



Miranda's Thick Regolith Indicates a Major Mantling Event from an Unknown Source

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Abstract

We investigated “muted” craters and scarps across Miranda’s cratered terrain. The morphologies of the muted craters are most consistent with modification by regolith deposition instead of erosion or viscous relaxation. We used three techniques to estimate regolith thickness. (1) Analysis of muted crater depth–Diameter (d - D) ratios near the South Polar Terrain Chasma indicates that regolith mantling their floors ranges from 0.3 to 1.2 km thick. Because older craters may have collected more regolith than younger craters, the true thickness may be similar to the highest estimate. (2) Analysis of crater size–frequency distributions across the cratered terrain indicates a thickness of 1.0 ± 0.2 km. (3) Analysis of a central mound within Alonso Crater indicates a thickness of $1.4^{+0.3}_{-0.4}$ km near Verona Rupes and may represent an upper limit. These results indicate that Miranda has one of the thickest regoliths in the solar system, which has important implications for Miranda’s interior thermal properties. Regolith appears to mantle some scarps within Arden but not Elsinore or Inverness, indicating that Arden may be the oldest corona, contrary to previous relative age estimates. In this scenario, the mantling event was ongoing during Arden’s formation but before Elsinore or Inverness formed. We propose three possible sources for Miranda’s thick regolith: (1) giant impact ejecta, (2) plume deposits, and (3) Uranian ring deposits. We favor the ring deposit hypothesis, which is consistent with the regolith’s large spatial extent, substantial thickness, and Miranda’s slightly spectrally blue color. Follow-up studies that rigorously investigate these scenarios are required.

Unified Astronomy Thesaurus concepts: [Uranian satellites \(1750\)](#); [Planetary science \(1255\)](#); [Regolith \(2294\)](#); [Craters \(2282\)](#); [Tectonics \(2175\)](#)

1. Background

1.1. Miranda’s Geology

The Uranian satellite Miranda has a complex surface, with multiple large regions of intense deformation called “coronae” that were likely formed by tectonism (e.g., Greenberg et al. 1991; Pappalardo et al. 1997) and might be at least partially cryovolcanic in origin as well (e.g., Schenk 1991; Beddingfield & Cartwright 2020). Miranda also displays a Global Rift System, composed of large sets of faults and fractures, that cut across its surface and bound the coronae (e.g., Greenberg et al. 1991). Miranda’s geologic setting and high heat flux estimates (e.g., Beddingfield et al. 2015a, 2022b) point to long-lasting geologic activity, likely driven by past orbital resonances shared between its neighboring moons Ariel and Umbriel (Tittlemore & Wisdom 1990; Ćuk et al. 2020). The coronae are fairly young—in particular, Inverness Corona near Miranda’s south pole ($0.1 \text{ Ga}^{+0.1}_{-0.4}$ Ga), which represents the youngest geologic terrain observed in the Uranian system (Kirchoff et al. 2022). In contrast to the youthfulness of Miranda’s coronae, its cratered terrains are more ancient (>1.2 Ga; Kirchoff et al. 2022) and exhibit a large number of “muted” impact craters and tectonic structures (Figure 1). The muted craters and scarps exhibit more subtle features (e.g., smooth crater rims and scarp slopes) that are more difficult to distinguish from the surrounding terrain than sharper “non-muted” craters and scarps.

The Voyager 2 flyby occurred during the southern summer, while Miranda’s northern hemisphere was in darkness, and it was therefore not imaged (Smith et al. 1986). A large part of Miranda’s imaged southern hemisphere is covered by “cratered terrain materials” (Figure 2) (Croft & Soderblom 1991). Age estimates vary for these regions, from 1.2 to 3.4 Ga for the lower crater density areas to $3.4 \text{ Ga}^{+0.9}_{-1.1}$ Ga for the areas of higher crater density (defined in Kirchoff et al. 2022). Based on the presence of muted impact craters and tectonic features, it was previously suggested that the cratered terrain was mantled by regolith (e.g., Croft & Soderblom 1991; Cartwright et al. 2020a), but the source of this material remains poorly understood. The cratered terrain is roughly split into two units called the heavily mantled (Mh) and lightly mantled (Ml) cratered terrain materials (Figure 2) (Croft & Soderblom 1991). The majority of Miranda’s cratered terrain is included in the heavily mantled unit, and a smaller region near Elsinore Corona on the anti-Uranus quadrant (between longitudes 150° and 250°) is included in the lightly mantled cratered terrain.

The Global Rift System has large faults that cut across the cratered terrain, including Verona Rupes, the largest known fault scarp in the solar system, which is ~ 10 km in elevation in some places. Several smaller systems of fractures are also present within the cratered terrain, some of which have been inferred via the presence of polygonal impact craters (PICs). Prior work demonstrated that the straight rim segments of PICs on Miranda might result from fractures that are either below the resolutions of available Voyager 2 Imaging Science System (ISS) images or are buried by regolith (Beddingfield & Cartwright 2020). The oldest corona is thought to be Elsinore ($1.2 \text{ Ga}^{+1.9}_{-0.8}$ Ga) while the youngest is likely Inverness ($0.1 \text{ Ga}^{+0.4}_{-0.1}$ Ga) (Zahnle et al. 2003; Kirchoff et al. 2022), and Arden Corona is estimated to have formed 0.1–1 Ga



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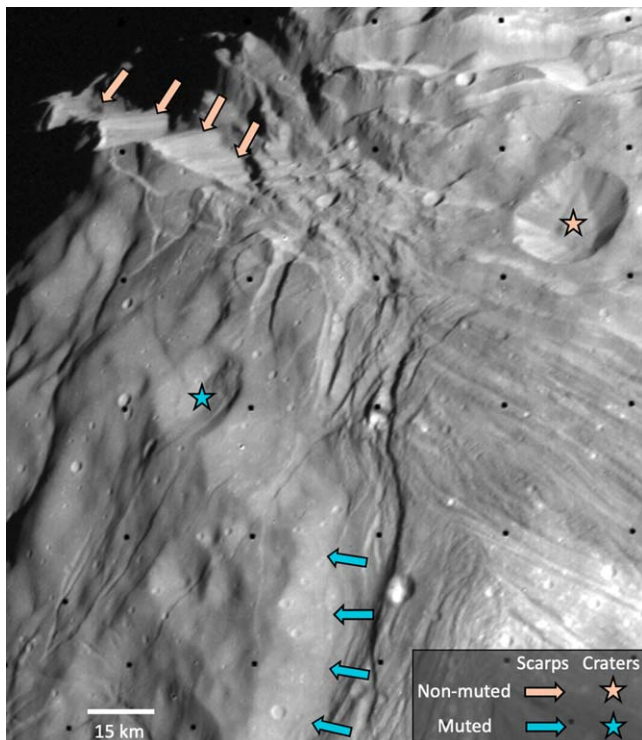


Figure 1. Examples of muted scarps and craters with more subtle features, and non-muted scarps and craters with more prominent features. Verona Rupes appears to be a non-muted scarp (orange arrows), with little evidence for mantling material on its surface. Individual corrugations are observable on the face of Verona Rupes, indicating the exposure of “bed-ice” material. In contrast, the Argier Rupes is muted (blue arrows) and has small impact craters, which may be secondary craters. Similarly, Alonso Crater (orange star) is non-muted, whereas Prospero Crater (blue star) is muted. This figure covers a portion of Voyager 2 ISS image c2684617 (284 m pixel⁻¹).

(Zahnle et al. 2003). However, the ranges in age estimates for each corona overlap, and therefore the order in which they formed is somewhat uncertain. The formation of Miranda’s coronae may have been driven by convective cells in its interior (e.g., Hammond & Barr 2014), and the high heat flux estimates for the regions bounding Arden Corona (Beddingfield et al. 2015a) and Inverness Corona (Beddingfield et al. 2022b) indicate endogenic geologic activity, supporting this scenario.

As summarized in Čuk et al. (2020), the most recent mean-motion resonance (MMR) in the Uranian system, the Ariel-Umbriel 5:3 MMR, likely resulted in significant tidal heating in Miranda’s interior that generated high heat fluxes. This MMR could have modified Miranda’s orbit and may explain its current 4.3° inclination. Additionally, other tidal resonances shared by Miranda with Umbriel and/or Ariel (Tittlemore & Wisdom 1990; Čuk et al. 2020) may have driven endogenic activity. However, Miranda likely does not have a subsurface liquid ocean today, unless it experienced tidal heating in the very recent past ($\lesssim$ a few tens of Ma; Castillo et al. 2022).

1.2. Impact Crater Modification across the Solar System

The morphologies of impact craters can be altered by one or more types of processes that are identifiable in remotely sensed imaging data sets. As shown in Figure 3(a), there are three types of end-member modification processes that may cause impact craters to appear muted: (1) regolith deposition forming mantled craters with subtle rims (Section 1.2.1), (2) erosion forming degraded craters with sharp rims (Section 1.2.2), and

(3) viscous relaxation that forms relaxed craters with sharp rims and uplifted crater floors (Section 1.2.3). Examples of these processes on rocky and icy bodies are shown in Figure 3(b), and other examples that include combinations of these processes are shown in Figure 4. Morphological characteristics expected for each end-member are summarized in Table 1.

1.2.1. Mantled Craters

On a planetary body, erosional and/or depositional processes can modify the morphologies of impact craters (e.g., French 1998; Osinski & Pierazzo 2013; Beddingfield et al. 2018), which is common for craters on Mars (Figure 4(a)). The presence of exposed bedrock or “bed-ice” in a crater rim indicates that the dominant modification processes are erosional, whereas depositional (i.e., mantling) processes dominate in regions where crater rims do not display exposed bedrock or bed-ice (e.g., Grant & Schultz 1993; Craddock et al. 1997). In locations where depositional processes dominate, the rate of regolith accumulation on crater rims is greater than the rate of removal. Once a sufficient amount of material has accumulated on crater rims, they appear muted in spacecraft images (see “Mantled” panels in Figures 3(a) and (b)). Impact craters mantled by dust and sand have been identified across Mars (Figure 4(a)); for example, in the Tisia Valles region, crater rims exhibit a wide degree of mantling morphologies (e.g., Beddingfield et al. 2018). On Vesta, many impact craters have been identified as mantled by regolith (Figure 4(c)) (e.g., Jaumann et al. 2012; Marchi et al. 2014; Denevi et al. 2016), possibly sourced from ejecta generated during large impact events including those that formed the Veneneia (400 km diameter) followed by the Rheasilvia (500 km diameter) impact basins, between 1 and 2 Ga (e.g., Schenk et al. 2012). In addition, some regions of Vesta with thick regolith may have accumulated more exogenic debris than other regions (Denevi et al. 2016). As summarized in Denevi et al. (2016), the region exhibiting a thick regolith may be a result of the presence of an ancient terrain that has accumulated a larger amount of exogenic material than other areas on Vesta.

On Enceladus, some impact craters near the south polar terrain (SPT) have been identified as “muted” (Figure 4(d)) (e.g., Kirchoff & Schenk 2009). As summarized in Kirchoff & Schenk (2009), these muted craters are likely mantled by plume deposits sourced from the nearby active Tiger Stripes. In addition to being mantled, these muted craters may have also experienced some degree of viscous relaxation, which is a dominant process for other craters on Enceladus, including those near the north pole (Figure 4(b)) (Bland et al. 2018). Regolith-mantled craters have also been identified elsewhere, including within the highlands on the Moon (e.g., Bart et al. 2011) and on Saturn’s small moons Helene, Telesto, and Calypso (e.g., Hirata et al. 2014).

1.2.2. Degraded Craters

Where erosional processes are dominant, mass-wasting and/or aeolian activity continuously expose the crater rims, revealing the bedrock/bed-ice (see “Degraded” panels in Figures 3(a) and (b)). During degradation, erosional processes are able to efficiently break apart high-standing crater rims, depositing much of this rim material downslope on the low-lying crater floor, thereby keeping the rim mostly exposed while burying the crater interior (e.g., Grant & Schultz 1993; Craddock et al. 1997; Barlow 2004; Robbins & Hynes 2012). These rims remain exposed because the

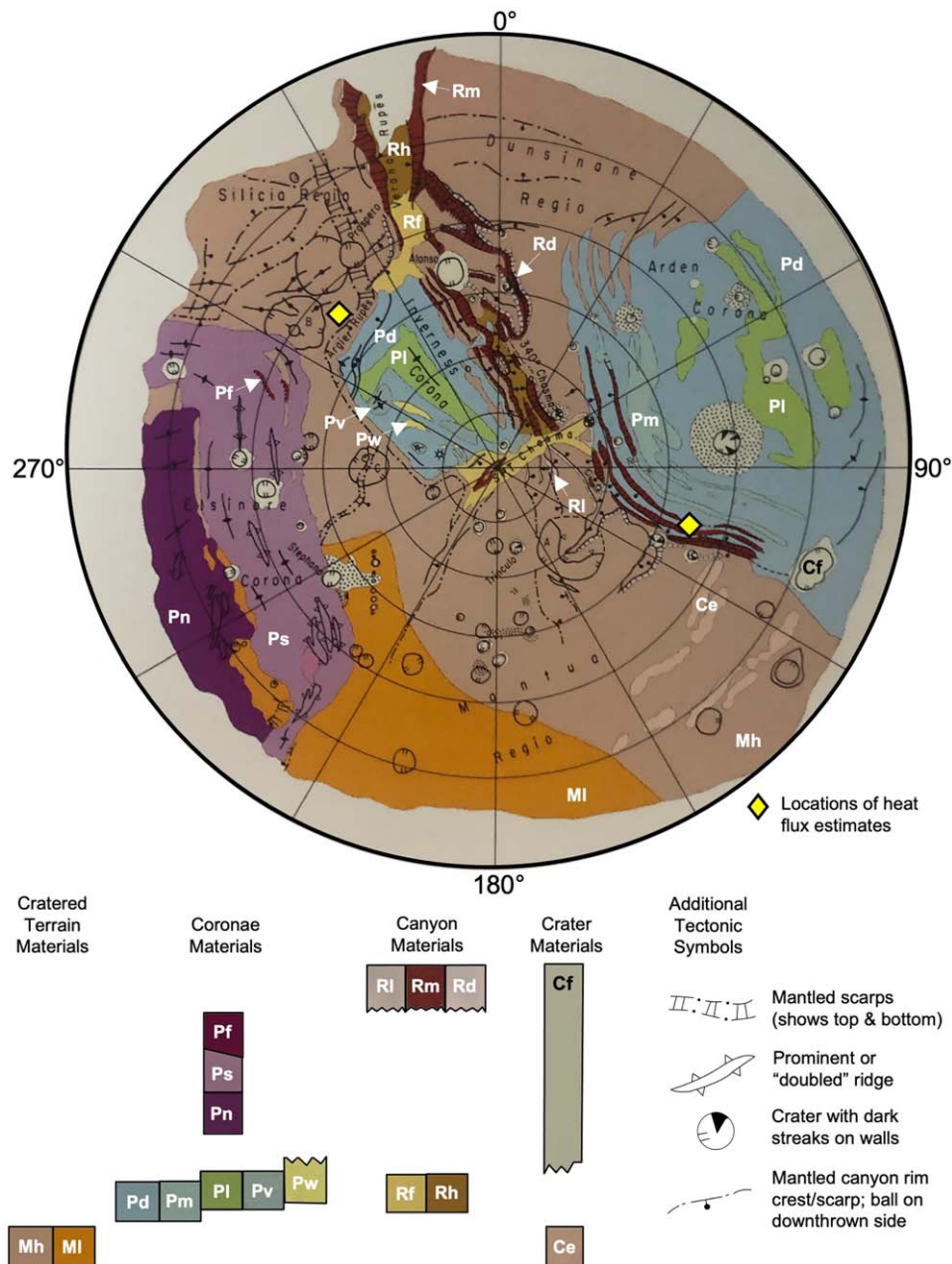


Figure 2. Annotated version of the geologic map of Miranda's imaged southern hemisphere (Croft & Soderblom 1991). The locations of heat flux estimates for Miranda are shown (Beddingfield et al. 2015a, 2022b). The Global Rift System is made up of the large fault scarps including Verona Rupes, Argier Rupes, the 340° Chasma, and the SPT Chasma (see the canyon materials geologic units). See Figure 6 for the highest-resolution Voyager 2 ISS images of Miranda and their locations. As summarized in Croft & Soderblom (1991), the geologic map units of Miranda are as follows. Mh: heavily mantled cratered terrain, MI: lightly mantled cratered terrain, Pf: flow material in Elsinore, Ps: southern ridged plains in Elsinore, Pn: northern ridged plains in Elsinore, Pd: dark material in Arden and Inverness, Pm: medium albedo material in Arden, Pl: light albedo material in Arden and Inverness, Pv: very dark albedo material in Inverness, Pw: wisp-like light material in Inverness, Rl: light albedo canyon slope material, Rm: medium albedo canyon slope material, Rd: dark albedo canyon slope material, Rf: fractured canyon floor material, Rh: hummocky canyon floor material, Cf: fresh impact craters, and Ce: elongate/chain craters.

rate of regolith removal is greater than the rate of local regolith production on the rims and/or deposition of material from a nonlocal source (e.g., Grant & Schultz 1993). Some regolith sourced from the rims may settle within the crater floor, or outside the crater along the base of the rims, as observed on Callisto (Figure 4(e)) (Chuang & Greeley 2000).

Both erosional and depositional processes may simultaneously modify a crater (e.g., Grant & Schultz 1993; Craddock et al. 1997). Some craters on Mars (e.g., the Arabia Valles and Tisia Valles regions) have been mantled by material delivered by aeolian processes (e.g., Barlow 1995; Beddingfield et al. 2018, 2021b).

Some of these craters have also been eroded and exhibit segmented rims and other signs of degradation (examples shown in Figure 4(a)). In these cases, regolith may originate from nonlocal sources, and the amount of rim mantling is largely independent of the degradation state of the crater.

1.2.3. Relaxed Craters

Impact craters may also be modified by viscous relaxation processes (Figure 3). Viscous relaxation reduces stresses associated with topographic relief by reducing topography

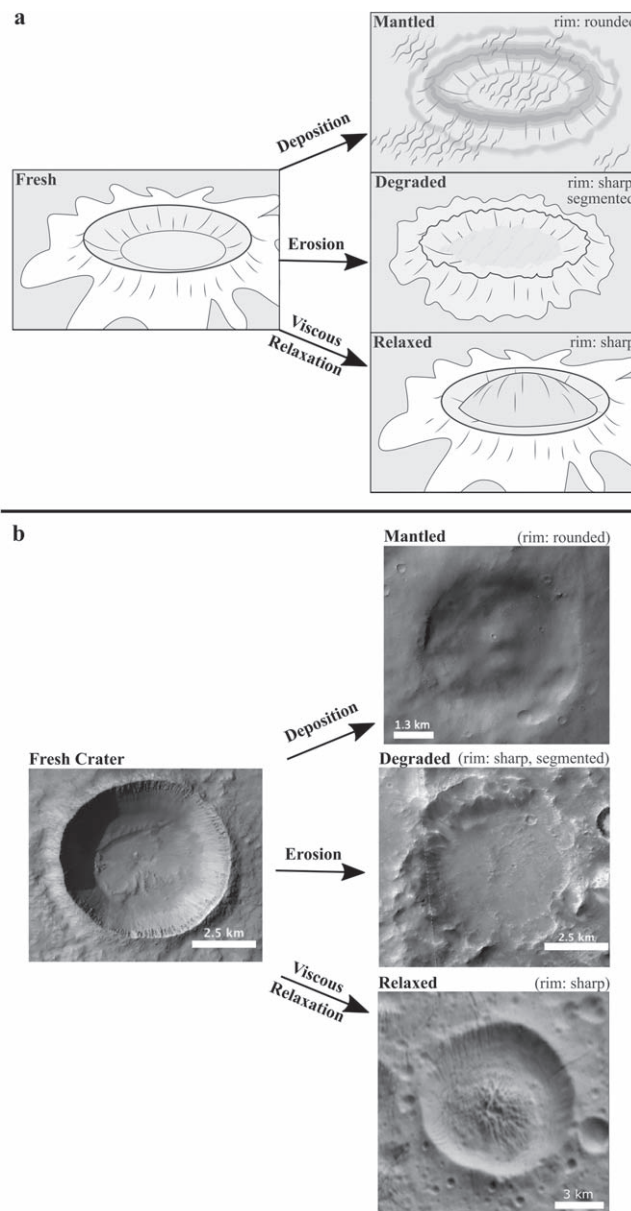


Figure 3. (a) Illustration showing that a fresh crater can become a mantled crater when located in a deposition-dominated environment, a degraded crater when in an erosion-dominated environment, or a relaxed crater if viscous relaxation has taken place. (b) Examples are shown of a fresh crater (Beddingfield et al. 2021b), a mantled crater (Beddingfield et al. 2018), and a degraded crater (Beddingfield et al. 2021b), all on Mars, as well as a relaxed crater on Enceladus (Bland et al. 2018).

over time (e.g., Scott 1967; Parmentier & Head 1981; Passey & Shoemaker 1982; Thomas & Schubert 1988), resulting in long-term ductile deformation of the material (e.g., Dombard & McKinnon 2006; Cathles 2015). Therefore, over geologic timescales, viscous relaxation acts to subdue topographic features (e.g., Melosh 1976; Grimm & Solomon 1988) and shallow slopes (e.g., Spencer 1984; Beddingfield et al. 2015b).

Fresh impact craters exhibit depth–Diameter (d - D) ratios that are largely consistent across the surface of a given planetary body (e.g., Robbins et al. 2018). However, floor uplifting caused by viscous relaxation can change d - D ratios for relaxed craters. Viscously relaxed craters have been identified on many icy bodies, including Enceladus, Tethys, Dione, Rhea, and Ariel (e.g., Schenk 1989; White et al. 2013, 2017; Bland et al. 2018, 2022) (Figure 4). Larger impact craters undergo viscous relaxation at a faster rate than relatively small craters, causing the floors of larger impact craters to uplift faster (e.g.,

Cathles 2015). As a result, viscous relaxation preferentially modifies the d - D ratios of large craters compared to smaller craters that formed at the same time.

2. Miranda’s Muted Craters and Scarps

2.1. Identification and Comparison of Muted and Non-muted Features

Miranda’s cratered terrains (labeled Mh and MI in Figure 2) show examples of both non-muted and muted craters and scarps (Figures 5 and 6). As summarized in Table 1, non-muted (fresher) craters have relatively sharp and prominent rims, whereas muted craters have rounded and more subtle rims (Figure 5) (Strom 1987; Croft & Soderblom 1991; Greenberg et al. 1991; Cartwright et al. 2020a). Non-muted fault scarps are composed of relatively sharp ridges and clear vertical corrugations, which reflect the displacement of the hanging

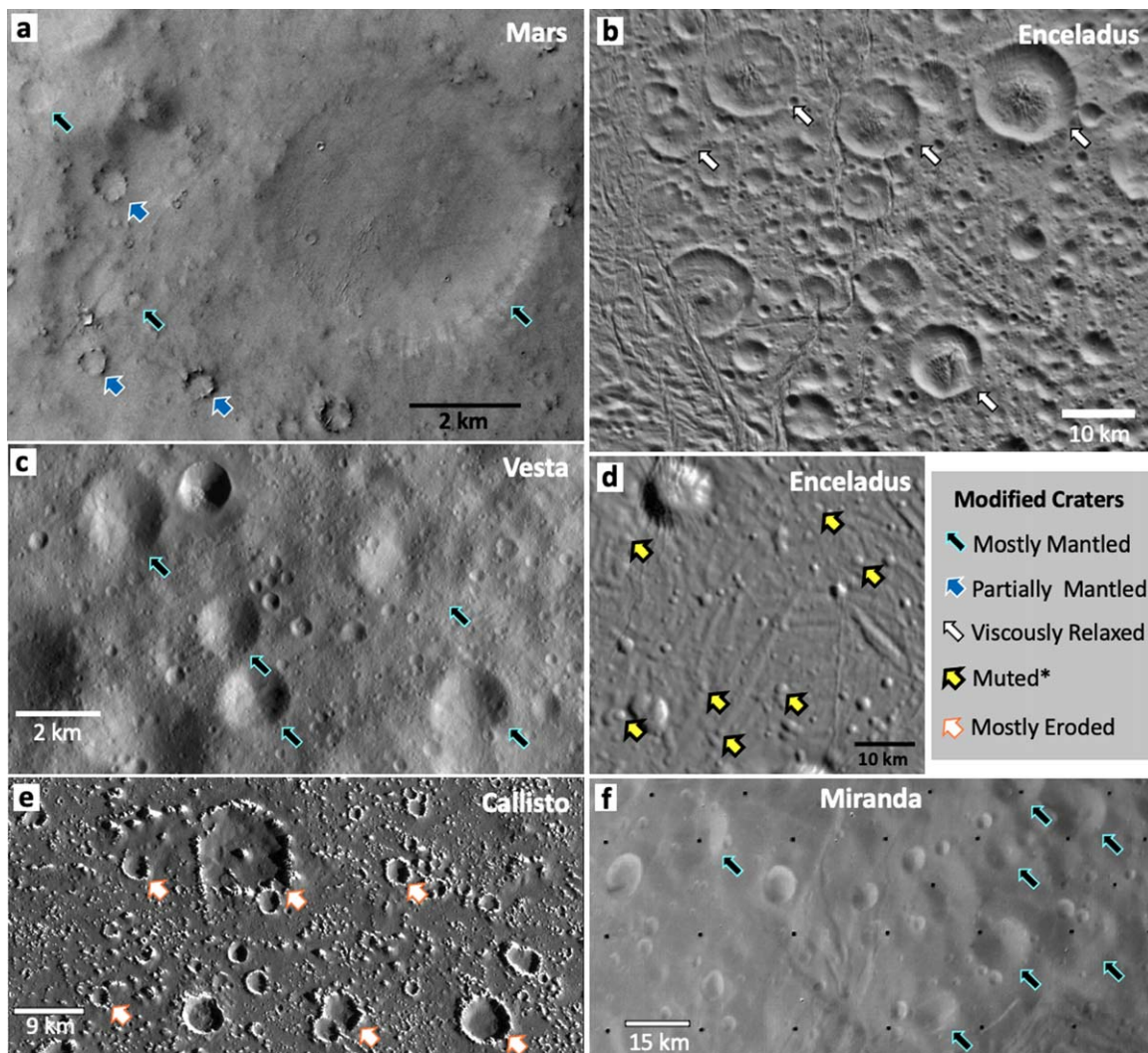


Figure 4. Examples of impact craters that have been modified. In some cases, combinations of processes affect an impact crater, so amalgamations of features detailed in Table 1 may be present. However, in this figure the craters are labeled by the most prominent modification process for simplicity. Panel (a) shows examples of “mostly mantled” and “partially mantled” craters in the Tisia Valles region of Mars, identified by Beddingfield et al. (2018). These craters are mantled by regolith deposited from a regional-scale deposition event and have also undergone some degree of degradation, forming irregular rims in locations where regolith cover is lower (Beddingfield et al. 2018). This figure shows a portion of Context Camera (CTX) image D20_035182_1681_XN_11S315W, and the small crater in the top right corner is located at $-10^{\circ}0$ latitude and $44^{\circ}3$ longitude. Panel (b) shows examples of impact craters that appear to have experienced viscous relaxation in the north polar region of Enceladus, north of Aladdin Crater and west of Samarkhand Sulci (shown in the global image mosaic of Enceladus (Bland et al. 2018)). The craters exhibit raised floors but have retained their sharp rims, showing how viscously relaxed impact craters are morphologically distinct from mantled impact craters. As shown in panel (c), many craters on Vesta are interpreted to be mantled by regolith (e.g., Marchi et al. 2014), and in some locations, the regolith is thought to be at least 1 km thick (Jaumann et al. 2012; Denevi et al. 2016). An example is shown in an image taken by the Dawn Framing Camera, NASA Photojournal image PIA15822 (NASA/JPL-Caltech/UCLA/MPS/DLR/IDA). Panel (d) displays examples of “muted” impact craters in the equatorial region of Enceladus that may be partly mantled, identified by Kirchoff & Schenk (2009). This figure shows the global mosaic of Enceladus at $8^{\circ}55$, 179° W. “The term “muted craters” was used by Kirchoff & Schenk (2009) to describe impact craters that may be partially mantled with plume fallout material, and that also may have experienced some viscous relaxation. Panel (e) presents examples of degraded craters on Callisto. This figure shows a portion of Galileo Solid-State Imaging (SSI) System image 2913r. The largest crater in this panel exhibits a lobate-shaped mass-wasting deposit on the crater floor. Similar mass-wasting deposits have been identified within many of Callisto’s craters (Chuang & Greeley 2000). Panel (f) presents examples of muted craters on Miranda, as shown in part of Voyager 2 ISS image c2684626 (Figure 6(d)).

wall during fault motion, and have albedos that may be different from the surrounding terrain. In contrast, muted scarps have rounded ridges at the top of their footwalls, with corrugations that are difficult to discern, and with scarp faces that are similar in tone and brightness to the surrounding terrain.

In the region between Inverness Corona and Miranda’s anti-Uranus hemisphere, south of Montua Regio, several examples of muted craters are proximal to non-muted craters of similar sizes and are observed at similar lighting geometries

(Figure 6(d)). The muted crater “Crater A” is present nearby the boundary of Arden Corona (Figures 6(d) and (e)) and appears to be relatively shallow. While Trinculo Crater is sharp, it exhibits a single streak of dark material along the interior of its rim in the region closest to Crater A and Arden Corona (Figures 6(d) and (f)). Similarly to Craters B and C, Crater A appears to be relatively shallow.

In addition to being present throughout Miranda’s cratered terrain, non-muted craters are present within Inverness (Figures 6(a) and (c)), Arden (Figure 6(d)), and Elsinore

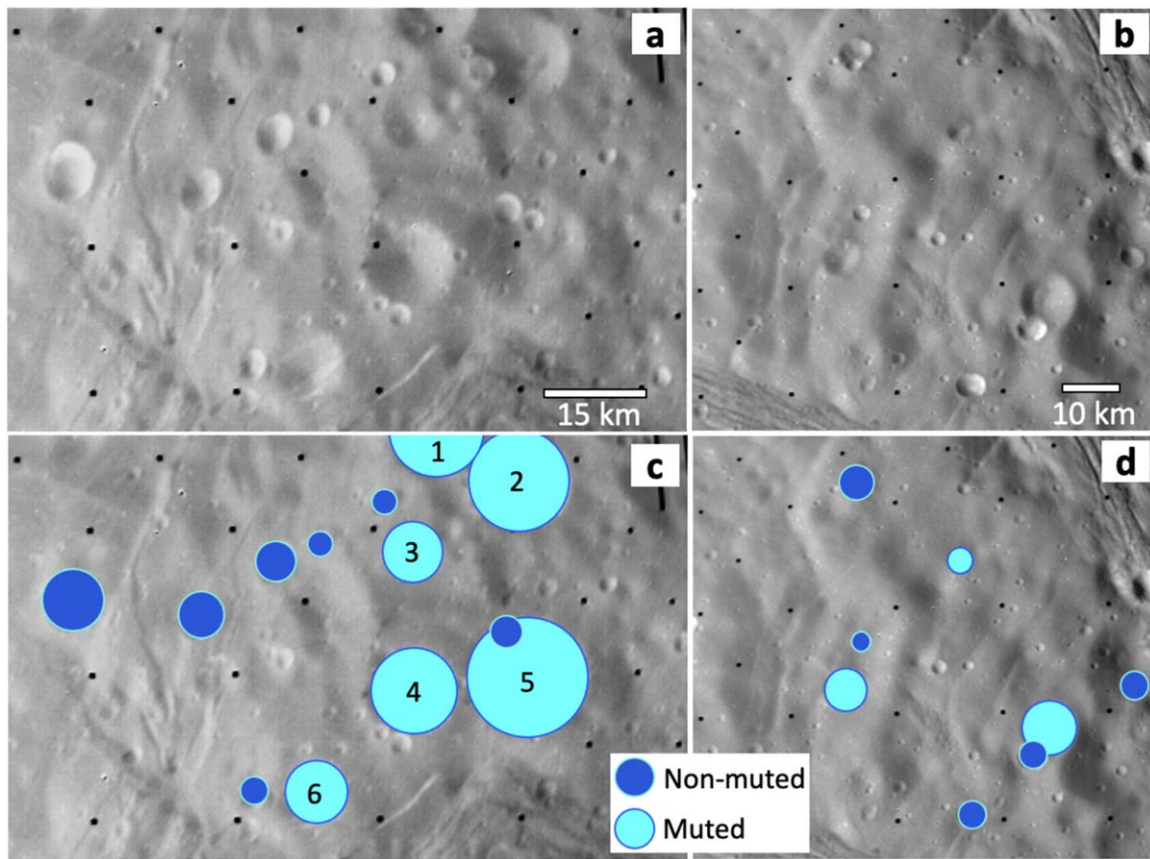


Figure 5. (a) A portion of Voyager 2 ISS image c2684626 (Figure 6(d)), in Miranda's cratered terrain. (b) A portion of Voyager 2 ISS image c2684629 (Figure 6(c)), in the cratered terrain between Inverness and Elsinore Coronae. (c) Annotated version of (a) showing the locations of muted and non-muted craters. (d) Annotated version of (b) showing the locations of muted and non-muted craters.

Table 1
Description of Expected End-member Morphologies for Different Types of Fresh and Muted Impact Craters

End-member Crater Type	Ejecta	Rim	Walls	Floor
fresh (no pre-existing regolith)	– possibly observable	– sharp – not segmented	– steep – possibly exposed layers	– flat or central peak (sharp)
fresh (pre-existing regolith)	– possibly observable	– sharp – not segmented	– shallow – possibly exposed layers	– flat or possibly central mound (rounded)
mantled	– not observable	– rounded – not segmented	– shallow – no exposed layers	– flat or possibly central mound (rounded)
degraded	– possibly observable	– sharp – segmented	– steep – possibly exposed layers	– flat or central peak (sharp)
relaxed	– possibly observable	– sharp – not segmented	– steep – possibly exposed layers	– upraised with radial fractures

Note. See Figure 3 for examples of mantled, degraded (eroded), and relaxed craters on planetary bodies. It is plausible that combinations of these processes affect impact craters, so amalgamations of the listed features may be typical in many cases.

Coronae (Figures 6(c), (e), and (f)). For example, the non-muted crater, Stephano, overprints both the boundary of Elsinore Corona and the lightly mantled cratered terrain (Figures 6(e) and (f)).

Verona Rupes is the largest observable non-muted scarp on Miranda, and the nearby Argier Rupes is a muted scarp (Figures 1 and 6(a), (b)). Nearby Argier Rupes, the muted crater called “Crater B” within Silicia Regio, is partially

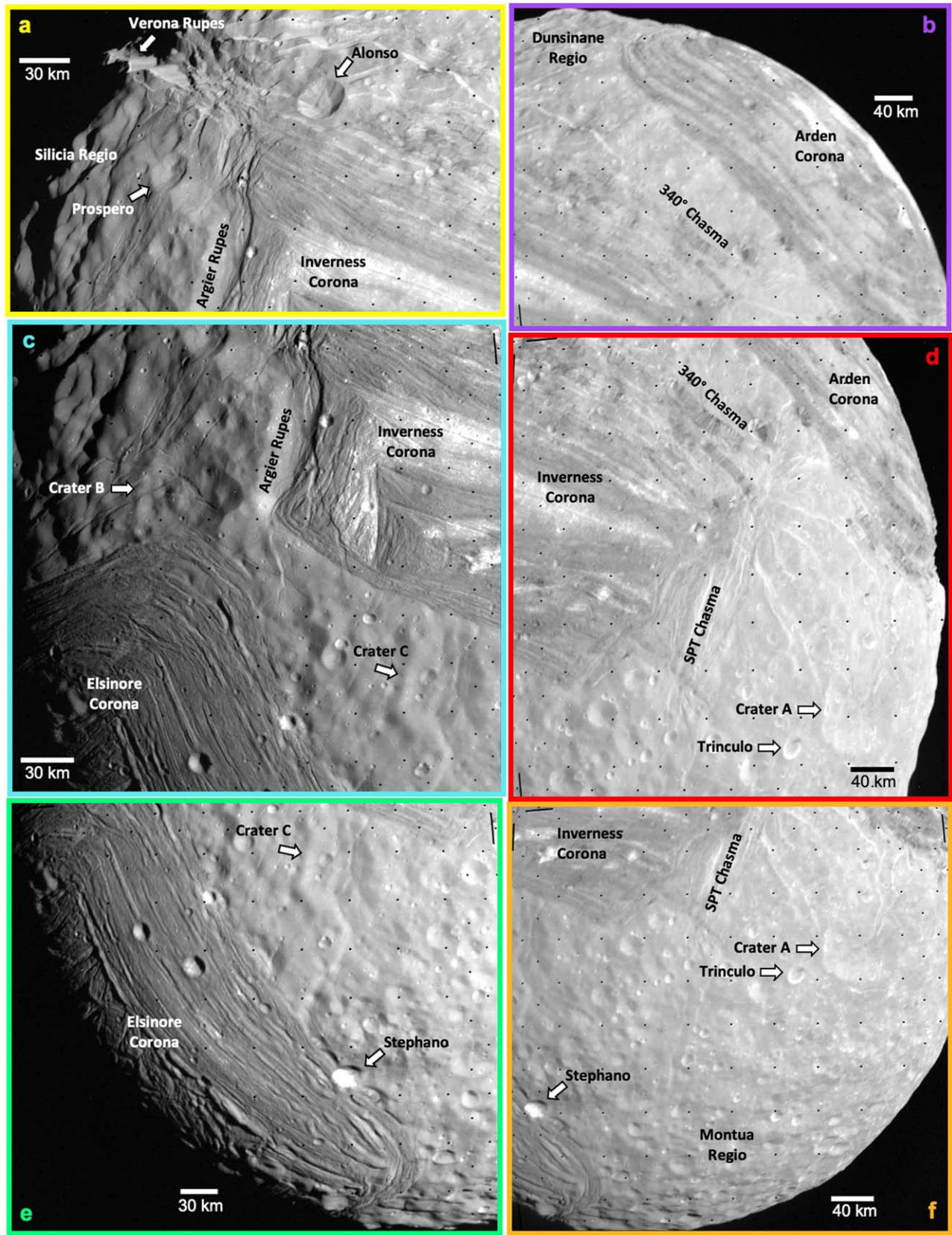


Figure 6. Voyager 2 ISS images of Miranda with annotated nomenclature. See Figure A1 for the unannotated version of this figure and Figure A2 for the specific regions covered by each image. (a) Image c2684617 (284 m pixel⁻¹), showing Inverness and Verona Rupes along Miranda's limb. (b) Image c2684611 (313 m pixel⁻¹), showing Arden and part of Inverness. (c) Image c2684629 (237 m pixel⁻¹), showing Inverness, Arden, and the cratered terrain. (d) Image c2684626 (247 m pixel⁻¹), showing Inverness, Arden, and the cratered terrain. (e) Image c2684620 (270 m pixel⁻¹), showing Elsinore and the cratered terrain. (f) Image c2684614 (298 m pixel⁻¹) showing the cratered terrain and small sections of all three coronae.

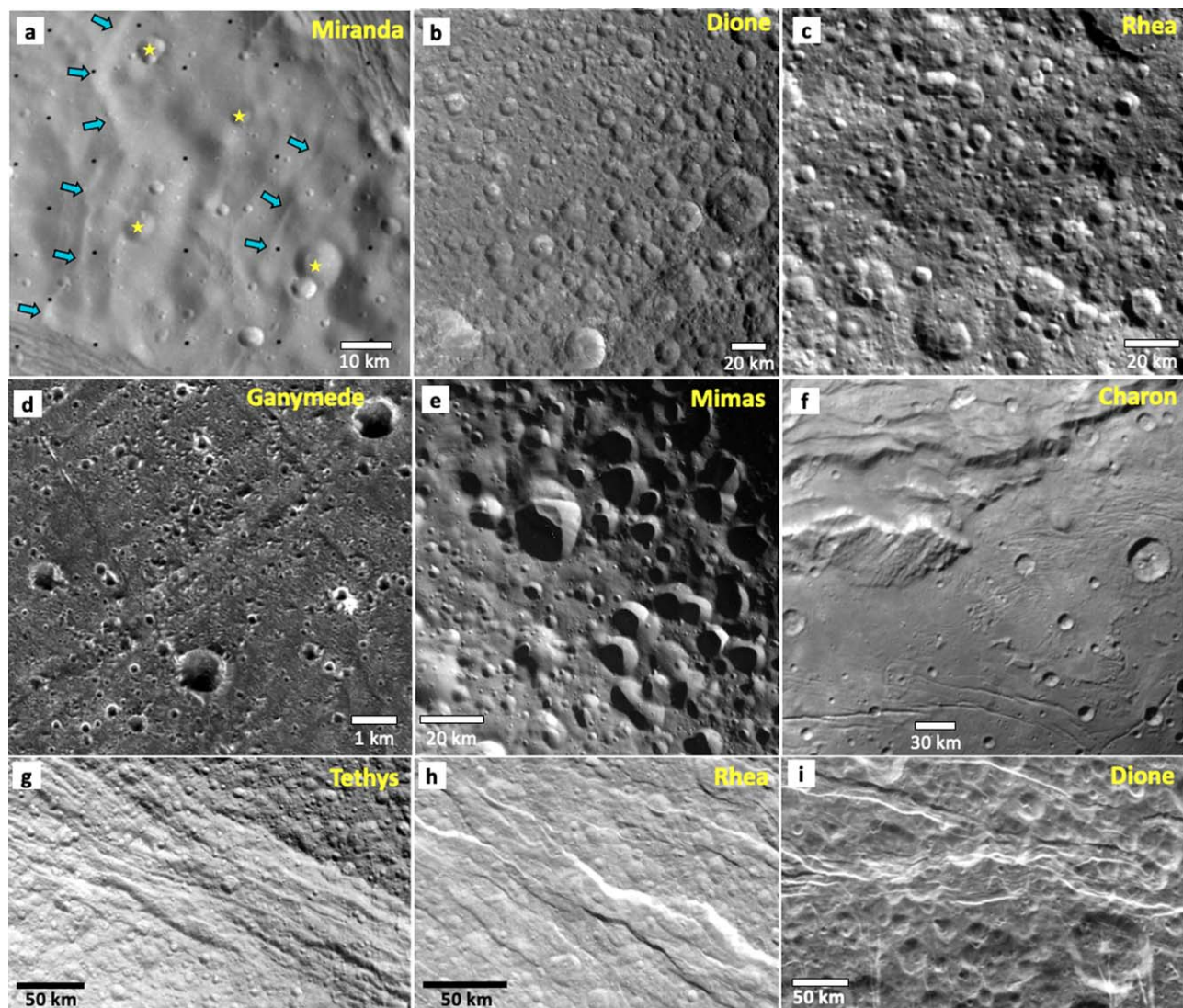


Figure 7. Comparison between muted impact craters and scarps on Miranda’s cratered terrain and cratered terrains on other icy bodies. Examples of craters on other planetary bodies that have been modified by mantling, degradation, and/or viscous relaxation are shown in Figure 4. (a) Muted impact craters (yellow stars) and “muted” scarps (blue arrows) on Miranda, shown in Voyager 2 ISS image c2684629 (Figure 6(c)). (b) Craters on Saturn’s moon Dione, shown in Cassini Imaging Science Subsystem (ISS) image N1714687700_1. (c) Craters on Saturn’s moon Rhea, shown in Cassini ISS image N1734911170_1. (d) Craters on Jupiter’s moon Ganymede, shown in Galileo Solid-State Imaging (SSI) image 3600r. (e) Craters on Saturn’s moon Mimas, shown in Cassini ISS image N1734911170_1. (f) Craters on Pluto’s moon Charon, shown in New Horizons LORRI image lor_0299175682. (g) A portion of Cassini ISS image N1716178094, showing normal faults within Ithaca Chasma on Tethys. (h) A portion of Cassini ISS image N1637520407, showing normal faults within Avaiki Chasmata on Rhea. (i) A portion of Cassini ISS image N1662199979, showing normal faults within the Wispy Terrain on Dione.

overprinted by Elsinore Corona, as well as two smaller muted craters near 27° longitude (Figures 6(c) and (e)). In these same images, the large muted crater termed “Crater C” is present between Elsinore and Inverness Corone and is overprinted by faults within the Argier Rupes system. Craters B and C appear relatively shallow, and Crater B has a central topographic high. In the region of Verona Rupes and Inverness Corona, near Miranda’s Uranus-facing hemisphere, a notable difference in morphology between non-muted and muted craters is apparent (Figure 6(a)). For example, within Dunsinane Regio, the non-muted crater Alonso, which overprints the 340° Chasma, and the muted crater Prospero, which overprints Argier Rupes within Silicia Regio, are similar in size (Figure 6(a)).

In some locations, non-muted tectonic features exhibit trends consistent with trends of muted tectonic features (Figure 2) (Croft & Soderblom 1991). For example, alternating sections of muted and non-muted scarps make up the 340° Chasma.

Relatively non-muted tectonic features also exhibit orientations concentric and radial to Arden Corona (Figures 2(c) and (e)).

2.2. Comparisons with Other Planetary Bodies

The presence of widespread muted craters across Miranda’s cratered terrain (Figure 7(a)) is unusual compared to other icy bodies. Typically, icy body surfaces exhibit craters with rims that are relatively sharp (Figures 7(b)–(f)), similarly to Miranda’s non-muted craters. For example the majority of impact craters on Saturn’s moon Dione do not appear to display a significant amount of mantling anywhere on the satellite (Figure 7(b)) (see Beddingfield et al. 2016 and Ferguson et al. 2022 for additional descriptions and images of craters across Dione). Similarly, impact craters on Rhea (Figure 7(c)) and Mimas (Figure 7(e)) exhibit non-muted craters with relatively sharp rims, even in their most heavily cratered and ancient

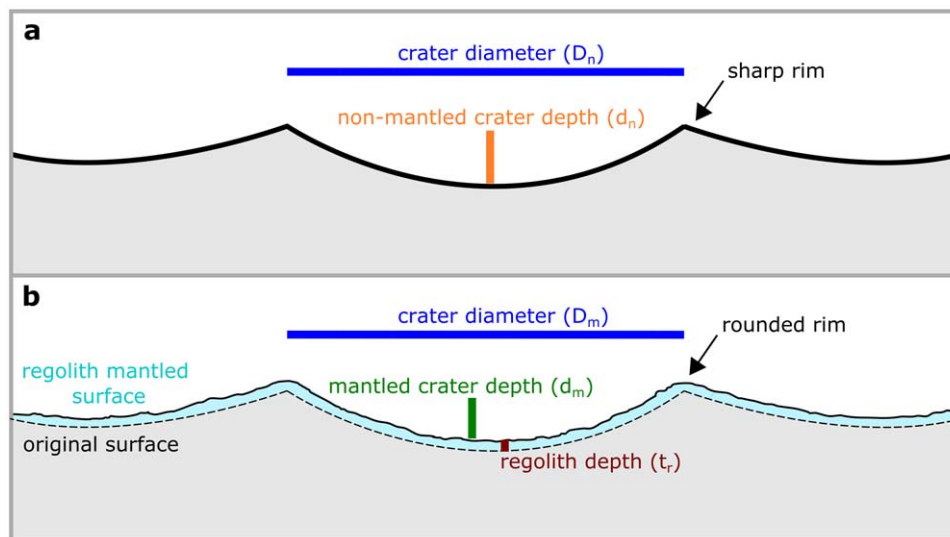


Figure 8. (a) Labeled diagram of a non-muted impact crater without the presence of regolith. (b) Diagram of a muted impact crater that is mantled by regolith.

terrains. Some large impact craters on Dione and Tethys exhibit evidence for viscous relaxation due to their low d - D ratios (White et al. 2017), as do some craters on Rhea (White et al. 2013). Jupiter’s moon Ganymede (Figure 7(d)) and Pluto’s moon Charon (Figure 7(f)) exhibit craters that also appear relatively fresh compared to Miranda’s muted craters. Additionally, some craters on Ganymede exhibit raised floors, indicating that viscous relaxation has taken place (Bland et al. 2017).

Like muted craters, the presence of widespread muted scarps, like those observed on Miranda, is uncommon on other icy bodies. Typically, scarps located in cratered terrains are relatively sharp (Figures 7(f)–(i)). For example, normal fault scarps on Charon exhibit sharp rims (Figure 7(f)), but some localized regions contain landslide deposits at the base of scarps in the Serenity Chasma region (Beddingfield et al. 2020). Similarly, normal faults that make up Ithaca Chasma on Saturn’s moon Tethys (Figure 7(g)), scarps within Avaiki Chasma on Rhea (Figure 7(h)), and scarps within Dione’s wispy terrain (Figure 7(i)) are all relatively fresh, with rims that appear sharp.

Miranda’s muted craters are morphologically most similar to mantled craters on Vesta (Figure 4(c)), and they are somewhat similar to the “mostly mantled” craters on Mars (Figure 4(a)) and the “muted craters” on Enceladus (Figure 4(d)), which may be mantled by plume deposits. Thus, after identifying the dominant morphologies for craters and scarps on Miranda and comparing them to craters and scarps on other planetary bodies, we find that the best explanation for Miranda’s muted features is mantling by regolith and not degradation or relaxation. Our findings are in agreement with the interpretation by Croft & Soderblom (1991) that Miranda’s cratered terrain has been mantled by regolith.

3. Estimating Miranda’s Regolith Thickness

3.1. Crater d - D Ratios

We investigated examples of non-muted and muted craters within Miranda’s cratered terrain region immediately north of Inverness Corona and the SPT Chasma. This area was imaged at relatively high resolution in comparison with other cratered

terrain regions on Miranda, allowing us to derive a more precise and higher-resolution digital elevation model (DEM) compared to other regions (see Appendix). We measured the depths (d) and diameters (D) of five non-muted and six muted craters in this region, calculating their d - D ratios (Figure 8). We used these measurements to estimate the regolith thickness in this region. These 11 craters are proximal to each other and are similar in diameter (Figure 5, also shown in Figures 6(f) and A3). Investigating craters that are proximal allowed us to mitigate confounding variables that can influence crater morphologies, including lithospheric thermal properties that may vary across Miranda’s surface, which in turn may cause different degrees of viscous relaxation. Additionally, comparing craters that are proximal allows for more consistency in lighting and spacecraft viewing geometries.

Based on craters analyzed by Schenk (1989), the average d - D ratio of impact craters on Miranda is approximately 0.132, which includes examples of both non-muted and muted craters. As shown in Figure 9 and Table 2, our results show that the d - D ratio specifically for non-muted craters in this particular region is $0.15^{+0.02}_{-0.03}$. In contrast, the muted crater d - D ratio in this region is $0.10^{+0.03}_{-0.02}$. We assumed that the impact craters underlying the regolith (Figure 8) had initial d - D ratios that were consistent with nearby non-muted craters before they were buried with regolith.

Because there is no evidence of bowed-up floors within these craters, it is unlikely that viscous relaxation notably altered their d - D ratios. While these craters have not been modified by viscous relaxation, other large craters elsewhere on Miranda may be relaxed (e.g., Crater A in Figure 6(c)). Additionally, the craters we analyzed are relatively small (<20 km in diameter), and viscous relaxation has been found to have a minimal effect on the d - D ratio of small craters relative to larger craters and impact basins (e.g., Solomon et al. 1982; Cathles 2015).

The depth of a non-muted impact crater (d_n) is equal to the depth of a muted crater (d_m) plus the thickness of the regolith (t_r) on its floor (Figure 8). The diameter of a non-muted crater (D_n) should not change after mantling has occurred (D_m). Therefore, $d_n/D_n = (d_m + t_r)/D_m$, and the thickness of the

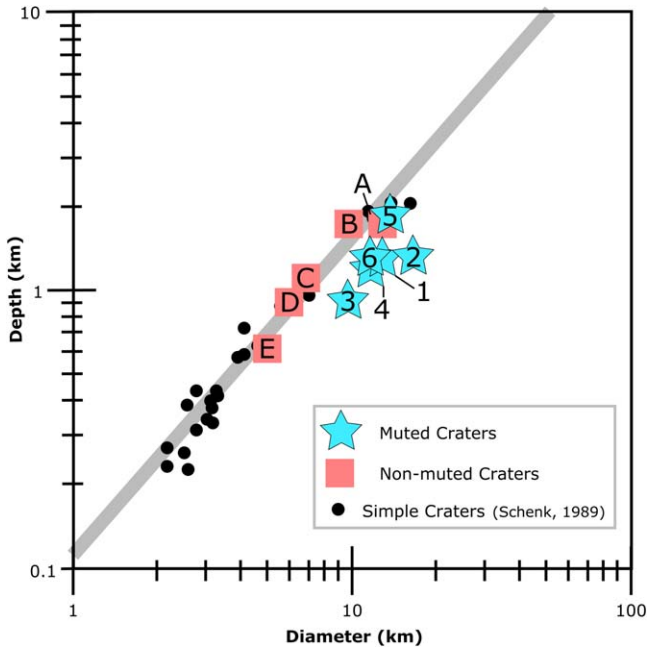


Figure 9. The d - D ratios of impact craters shown in Figure A3 for the six muted craters (Figure A5) and five non-muted craters (Figure A4 craters). The average d - D ratio estimate for impact craters in general on Miranda is 0.132 (Schenk 1989) (black dots). In comparison with this value, we estimate that the average d - D ratio is $0.15^{+0.02}_{-0.03}$ for non-muted craters and $0.10^{+0.03}_{-0.02}$ for muted craters in Miranda’s South Polar Terrain Chasma region (Table 2).

Table 2

Depth–Diameter (d - D) Ratios of the Impact Craters Shown in Figure A3 and Plotted in Figure 9

Crater ID	Crater Type	d (km)	D (km)	d - D Ratio	t_r (km)	α (°)
A	Non-muted	1.7	13	0.13	...	23
B	Non-muted	1.7	10	0.17	...	26
C	Non-muted	1.1	7	0.16	...	27
D	Non-muted	0.9	6	0.15	...	23
E	Non-muted	0.6	5	0.12	...	17
1	Muted	1.3	13	0.10	0.6 ± 0.3	...
2	Muted	1.3	17	0.08	1.2 ± 0.5	...
3	Muted	0.9	10	0.09	0.6 ± 0.3	...
4	Muted	1.2	12	0.10	0.5 ± 0.3	...
5	Muted	1.8	14	0.13	0.3 ± 0.3	...
6	Muted	1.3	12	0.11	0.4 ± 0.3	...

Note. Our estimated regolith thicknesses and the angle of repose (α) (see Section 3.3) are also provided. The average d - D ratio for muted craters is $0.10^{+0.03}_{-0.02}$, while the average ratio for non-muted craters is $0.15^{+0.02}_{-0.03}$ in the region north of Inverness and the SPT Chasma. The difference in the d - D ratio between these two groups is likely due to the presence of regolith within the muted craters.

regolith layer mantling the floor of a muted crater is given by

$$t_r = \frac{d_n}{D_n} * D_m - d_m, \quad (1)$$

where $\frac{d_n}{D_n}$ is the expected d - D ratio of non-muted impact craters, which we estimate to be $0.15^{+0.02}_{-0.03}$ in this region of Miranda.

Using our measurements of the depths and diameters of the five non-muted and six muted craters in Equation (1), we estimate that t_r within the muted craters ranges between 0.3 and

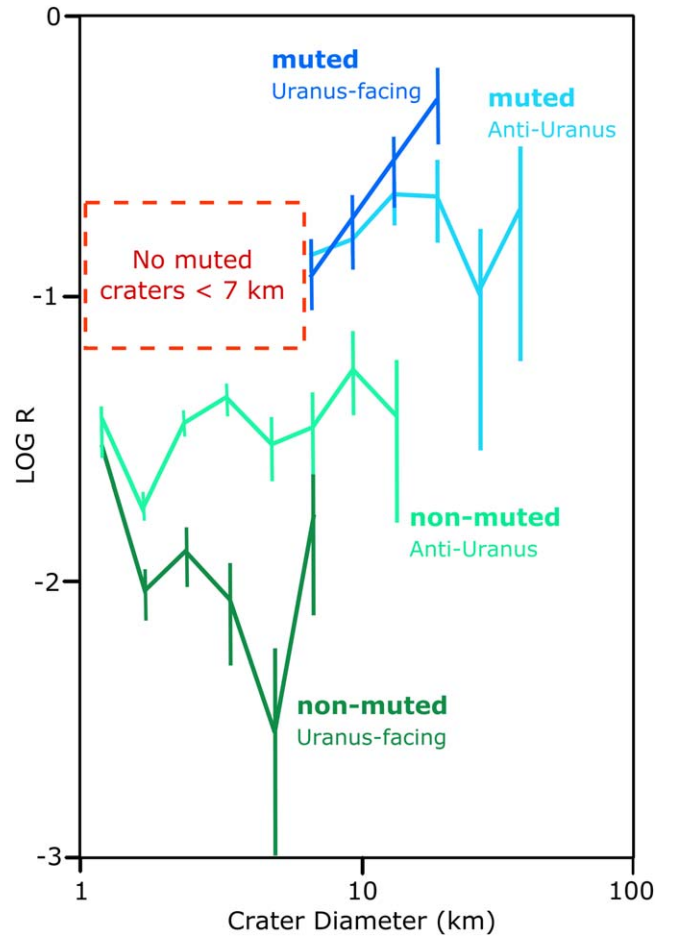


Figure 10. Annotated version of Figure 12 in Greenberg et al. (1991), which shows the relative size–frequency distribution (R-plot) of non-muted and muted impact craters within the cratered terrain on Miranda’s Uranus-facing (centered at $\sim 0^\circ$ longitude) and anti-Uranus (centered at $\sim 180^\circ$ longitude) hemispheres (see Figure 2 for locations of geologic features with respect to these hemispheres). Muted craters that are smaller than ~ 7 km are not visible in available images. Larger $\text{Log } R$ values correspond to higher spatial densities of craters. Because most impact crater populations throughout the solar system are approximated by a cumulative power law index of -3 , this value is chosen for normalization of the data, and deviations from this value appear as variations from a horizontal line. See Barlow (1988) for more details on interpreting crater R-plots. As described in Greenberg et al. (1991), the densities of muted craters on Miranda’s Uranus-facing and anti-Uranus sides are similar, but non-muted crater densities are higher on its Uranus-facing side. Greenberg et al. (1991) notes that there is a concentration of small fresh (and therefore non-mantled) craters in a local region offset from the sub-Uranus point, but the specific location was not provided.

1.2 km, with an average thickness of 0.6 km (Table 2). The range in t_r between the six mantled craters could result from differences in formation ages. In this scenario, older craters would have had more time to accumulate thicker layers of regolith. Alternatively, perhaps there were local-scale spatial variations in the rate of regolith deposition in this region of Miranda.

The upper end of the muted crater d - D ratio range overlaps with the lower end of the non-muted crater range (see muted Crater 5 and the non-muted Crater A). This similarity in d - D ratios is likely due to the presence of a thin mantling regolith within muted Crater 5 and mass-wasting deposits within non-muted Crater A. There is evidence for mass-wasting within this crater in the form of a dark wall streak, supporting this interpretation (see Figure A3, and see also Section 3.3 for

Table 3
Estimated Regolith Thickness Values (t_r) from Various Methods

Method	Geologic Feature	Section/Reference	t_r (km)
d - D Ratio	Non-muted and Muted Craters (e.g., Trinculo)	Section 3.1	0.3–1.2 km*
CSFD	All Known Muted Craters	Section 3.2	1.0 $\pm$ 0.2 km
Central Mound	Alonso (Non-muted Crater)	Section 3.3	1.4 ^{+0.3} _{-0.4} **
Bright Layer	Scarp and Crater Walls	Smith et al. (1986)	≥ 1 (in some areas)

Note. *The large variation in estimated thicknesses within individual muted craters may result from different crater formation ages. We would expect older muted craters to have thicker regoliths than younger muted craters. Therefore, the true thickness is likely closer to the higher end of the range estimated using this technique (see Section 3.1). **The thickness estimate from the central mound of Alonso may represent an upper limit and has important caveats (see Section 3.3).

additional information on these mass-wasting features within non-muted craters).

3.2. Crater Size–Frequency Distributions

Miranda’s muted and non-muted craters exhibit notably different crater size–frequency distributions (CSFDs) (Figure 10) (Greenberg et al. 1991). As discussed in Greenberg et al. (1991), Miranda’s muted craters are confined to the cratered terrain, whereas non-muted craters overprint both the cratered terrain and the coronae. The muted craters within Miranda’s heavily cratered terrain exhibit a CSFD with a differential -2 slope. However, the non-muted impact craters exhibit a size/frequency distribution with a differential -3 slope for crater diameters >1.5 km (Figure 10). Impact craters with diameters below this value show a CSFD similar to secondary impact craters on other bodies. Greenberg et al. (1991) suggest that these putative secondary impact craters may have resulted from ejecta that initially escaped Miranda but later reaccumulated onto its surface. This ejecta may have originated in a large basin-forming impact event that might have spurred the formation of Arden Corona (Croft & Soderblom 1991). However, Arden Corona does not display evidence for typical impact related structures (e.g., crater rims) that would support this scenario (McKinnon et al. 1991).

Based on our estimated d - D ratio for Miranda’s non-muted craters, the regolith depth required to completely fill the interior bowl of a crater, so that it is equivalent to the elevation of the surrounding terrain, is given by

$$t_r = (d - D)_n * D_{\min}, \quad (2)$$

where $(d - D)_n$ is the mean d - D ratio of non-muted craters ($0.15^{+0.02}_{-0.03}$) and $D_{\min}$ is the minimum muted crater diameter identified on Miranda (~ 7 km). Therefore, the resulting value of t_r indicates that a 1.0 ± 0.2 km thick layer of regolith is required to fill Miranda’s smallest muted craters, making them indistinguishable from the surrounding terrain (Table 3).

3.3. Crater Central Floor Mounds

Some of Miranda’s craters show evidence of debris slides along their interior rim walls (Figure 11(b)) and central floor mounds within their bowls (Figures 11(c), (d), and (e)) (Pappalardo et al. 1997). For example, a small crater on the cratered terrain, located between Inverness and Elsinore, shows a partially concentric slump within its crater bowl (Figure 11(c)). Additionally, Alonso Crater, overprinting the 340° Chasma near Verona Rupes, is a non-muted crater with a floor that appears to consist of mass-wasting deposits (Figure 1(d)). “Crater A,” as

defined in Croft & Soderblom (1991), exhibits a tongue of material that covers most of the floor and buries the side of its rim facing Arden. This crater sits within the heavily mantled cratered terrain (Figures 6(d), (f)). “Crater B” (45 km diameter), as defined in Croft & Soderblom (1991), in Silicia Regio, exhibits a central mound (Figures 11(e) and 6(c)). This central mound is partially overprinted by a corner of Elsinore Corona, and it could be a mass-wasting deposit.

As derived from physical laboratory experiments (Quaide & Oberbeck 1968), and utilized to estimate lunar regolith thicknesses (Bart et al. 2011; Bart 2014; Yue et al. 2019), the thickness of a pre-existing layer of regolith that experiences an impact event can be estimated using central mounds on their floors (Bart et al. 2011). In a non-muted crater with a central mound, t_r is estimated by

$$t_r = (k - D_{\text{mound}}/D) * D * \tan(\alpha)/2, \quad (3)$$

where k is an empirical constant (0.86) that relates the true diameter of a crater to the apparent diameter (Quaide & Oberbeck 1968), D_{mound} is the diameter of the inner central mound of an impact crater, D is the crater diameter, and α is the angle of repose of the regolith.

The angle of repose, the maximum slope angle at which a non-cohesive granular material can be at rest (Lowe 1976), is likely independent of gravitational acceleration (Nakashima et al. 2011; Atwood-Stone & McEwen 2013; but see Kleinhans et al. 2011 for an alternative view). However, α values can vary significantly depending on properties of the granular material (i.e., dominant grain size range and shape of the regolith grains), ranging from $\sim 15^\circ$ to 45° (e.g., Carrigy 1970; Pohlman et al. 2006; Beakawi Al-Hashemi & Baghabra Al-Amoudi 2018). By measuring the interior slope angles of fresh craters on the Moon, prior work reported an estimated α value of $31^\circ \pm 2^\circ$ for the lunar surface (Quaide & Oberbeck 1968). Furthermore, craters that form on a planetary surface mantled by a pre-existing thick regolith layer have interior walls with slopes that are consistent with the α of the regolith (Quaide & Oberbeck 1968).

Because the non-muted craters on Miranda’s cratered terrain formed in regions that are mantled by a thick blanket of regolith, we measured the interior slope angles of the five non-muted craters on Miranda’s cratered terrain to estimate α (Table 2). These measurements yield an average α value of 23° , with a full range of 17° – 27° . We used this α range to estimate the thickness of the regolith layer associated with the non-muted crater Alonso. Alonso Crater exhibits a central mound (Figure 11(d)) and it was imaged at relatively high resolution, allowing for good estimates of central mound

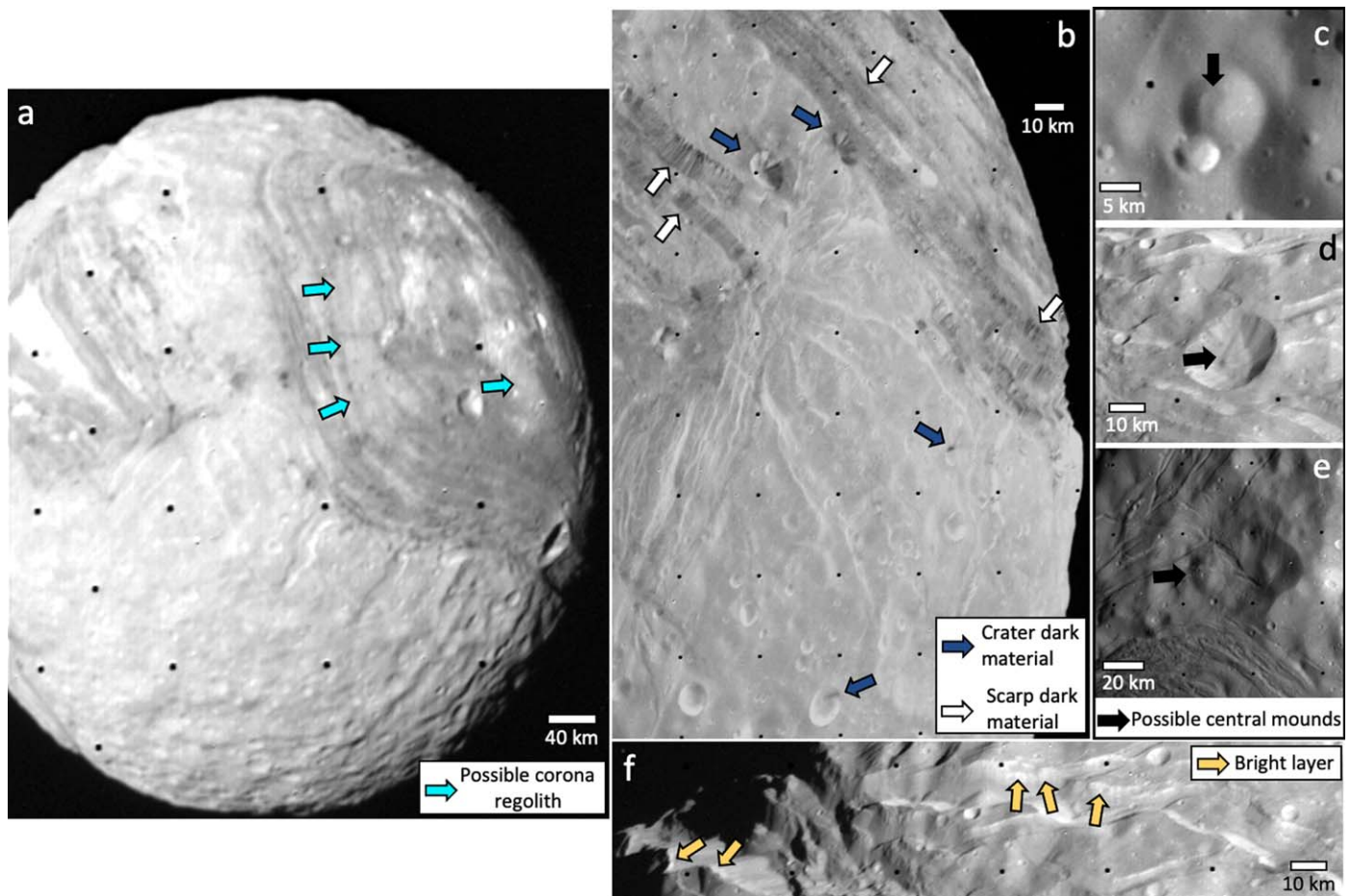


Figure 11. Annotated Voyager 2 ISS images showing locations of additional regolith and mass-wasting deposits. See Figure 2 for locations of geologic features. (a) Image c2684409 (1151 m pixel⁻¹) showing regolith that appears to mantle scarps within Arden Corona in some locations. (b) A portion of image c2684626 (247 m pixel⁻¹), showing the locations of dark material along crater and scarp walls. (c) A portion of image c2684629 (237 m pixel⁻¹), showing a crater with a central deposit. (d) A portion of image c2684617 (284 m pixel⁻¹), showing the non-muted crater Alonso, which has a central deposit. (e) A portion of image c2684629 (237 m pixel⁻¹) that shows “Crater B” as defined in Croft & Soderblom (1991). The positive topography feature within the crater may be a central mound deposit of regolith, an uplifted floor, or a central peak. (f) A portion of image c2684617 (284 m pixel⁻¹) showing locations of a bright layer, identified by Smith et al. (1986), along the top of scarp walls.

dimensions. The D of Alonso Crater is 24 km, and its D_{mound} averages 14 km.

We must make the assumption that additional mantling and mass wasting was minimal following the formation of Alonso Crater, and that all regolith contained within Alonso’s central mound is sourced from regolith that predates the impact event. Consequently, Miranda’s regolith thickness estimated from this method may represent an upper limit to the true thickness prior to the impact event that formed Alonso, because subsequent deposition of regolith may have occurred. Based on our results, we estimate that the regolith thickness in the area of Alonso Crater is $1.4^{+0.3}_{-0.4}$ km (Table 3). Another important caveat is that this method has only been utilized for craters within much thinner regolith. Ongoing laboratory studies are investigating the accuracy of this method for various experimental designs (e.g., Anderson et al. 2020, 2021; Cline & Cintala 2022).

3.4. Crater Streaks

Some non-muted impact craters exhibit evidence for low-albedo streaks along their walls (Figure 11(b)), which are darker than the ancient cratered terrain (Pappalardo et al. 1997). For example, two impact craters near the boundary of Arden (Figure 6(d)) exhibit multiple vertical bands of dark material on

the walls facing away from Arden (i.e., south-facing, toward Inverness and Elsinore). The walls facing toward Arden are difficult to discern, and it is uncertain whether dark streaks are also present on the Arden-facing (i.e., north-facing) crater walls as well. Other impact craters, further from the boundary of Arden, only exhibit dark streaks on the walls facing away from Arden, and craters even further from Arden show no discernible evidence for dark streaks.

In addition to crater walls, dark streaks are discernible on the sides of the large fault scarps that make up Arden Corona and the 340° Chasma, and they are also present along the low-lying bases of these large normal fault scarps (Figure 11(b)) (Pappalardo et al. 1997). The dark material observed along these features may have originated from Miranda’s subsurface (Smith et al. 1986; Pappalardo et al. 1997). Alternatively, perhaps this dark material is exogenic in origin, and was delivered via a large impact event. This type of event may be similar to the hypothesized delivery of dark material to Vesta’s surface via a giant impact event (Reddy et al. 2012). This material was either originally dark or it has been darkened over time following its exposure (e.g., due to charged particle weathering). As summarized in Pappalardo et al. (1997), the dark material may have originated near the top of the normal

fault scarps and progressed downslope due to mass-wasting processes, perhaps following along corrugations associated with movement of the hanging wall along the fault scarp. Based on the observation that dark streaks are not present on all crater walls nor all fault scarps, this material likely represents a discontinuous subsurface layer, which could be up to a few hundred meters thick, as estimated in Pappalardo et al. (1997).

In contrast to the dark-streaked scarps within and nearby Arden, other scarps like Verona Rupes and those comprising the SPT Chasma bounding Inverness exhibit streaks of material that are notably brighter than the surrounding cratered terrain (Figure 11(f)) (Smith et al. 1986). This bright layer is estimated to be ~ 1 km thick (Smith et al. 1986), and it may have originated in Miranda's interior. This bright layer may represent a regolith profile, and its thickness might be representative of the thickness of Miranda's regolith in that location (Table 3).

4. Discussion

4.1. Is Miranda's Regolith One of the Thickest in the Solar System?

The surface of Miranda exhibits a large number of impact craters and fault scarps within the cratered terrain that have been mantled by a thick layer of surficial regolith. We estimated the thickness of Miranda's regolith on the cratered terrain using three different techniques (Table 3): d - D ratios of muted craters (t_r : 0.3–1.2 km), size–frequency distributions of muted craters (t_r : 1.0 ± 0.2 km), and analysis of a central mound in Alonso Crater (t_r : $1.4^{+0.3}_{-0.4}$ km). These estimates are similar to a ~ 1 km thick bright layer observed within the walls of some scarps on Miranda, which could also represent a regolith profile (Smith et al. 1986). Therefore, our analyses indicate that Miranda's cratered terrain is mantled by a substantial layer of regolith, which has important implications for Miranda's internal processes and its ability to retain heat.

Thick regoliths serve as insulating layers, trapping heat in an icy body's interior and causing its ice shell to be warmer than its surface temperature (e.g., Passey & Shoemaker 1982; Passey 1983). Heat trapped in an icy body's interior could enhance endogenic activity, as examined for Enceladus by Martin et al. (2017) and Bland et al. (2012). Consequently, Miranda's thick regolith might have insulated its interior and promoted endogenic activity, such as the formation of one or more coronae and the expansion and reactivation of faults in the Global Rift System.

Thickness estimates of Miranda's regolith are most similar to Vesta, which displays regolith deposits of at least 1 km thick in some locations (Jaumann et al. 2012; Denevi et al. 2016). Mercury's regolith cover is estimated to be much thinner (20–90 m thick; Kreslavsky & Head 2015), as is the range of regolith thicknesses estimated for the lunar surface (4–15 m), with thinner deposits on the mare and thicker ones on the highlands (Quaide & Oberbeck 1968; McKay et al. 1991; Shkuratov & Bondarenko 2001; Bart et al. 2011). Both of Mars' moons, Deimos (10–100 m; Veverka & Thomas 1979) and Phobos (2–100 m; Basilevsky & Shingareva 2010; Ramsley & Head 2021), have notable regoliths that likely were formed by impact processes. Ceres is estimated to have a regolith between 150 and 300 m in depth (Otto et al. 2019).

On Enceladus, regolith near the South Polar Terrain is estimated to be 90–290 m thick, with this regolith likely

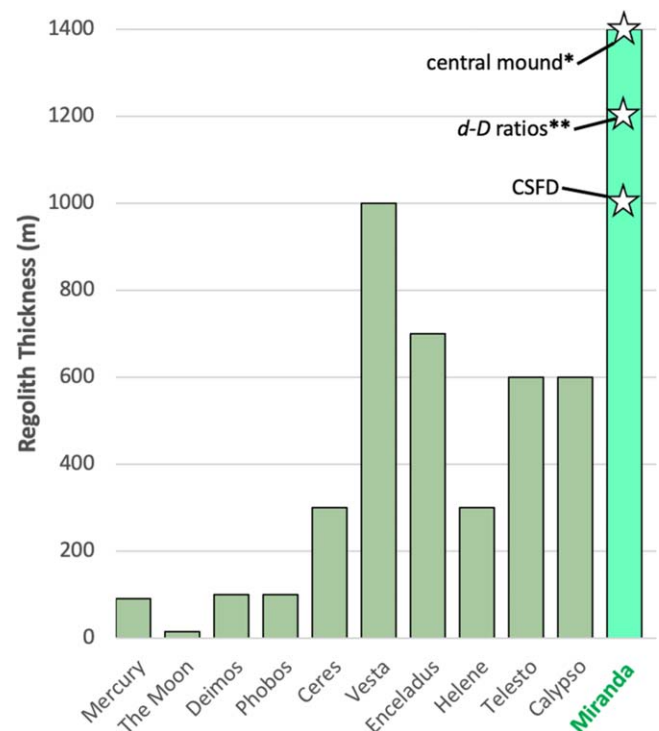


Figure 12. Comparison between the maximum regolith thickness estimates for Miranda and other planetary bodies where regolith layers have been observed and measured. These regolith thickness estimates are for surficial regolith and not megaregolith. Miranda's regolith thickness estimates derived from crater d - D ratios (Section 3.1), the crater size–frequency distribution (CSFD) of muted craters (Section 3.2), and the central mound within Alonso Crater (Section 3.3) are shown as white stars. *As described in Section 3.3, Miranda's regolith thickness estimated from the central mound technique may represent an upper limit and has important caveats. **As described in Section 3.1, the highest estimate for regolith thickness within the muted craters (1.2 km) may be most representative of the true thickness in this region.

originating from plume deposits and impact ejecta (Martin et al. 2017). Recent work indicates Enceladus' regolith is up to 700 m thick in some regions (Martin et al. 2022). Furthermore, the surfaces of Tethys' small Trojan moons Telesto and Calypso are blanketed by thick regoliths (~ 600 m thick), as is Dione's Trojan moon Helene (10–300 m thick) (Hirata et al. 2014). The regolith on Europa may be on the order of ~ 30 m thick, based on the thickness of a bright layer along a fault face that cuts a ridge within the ridged terrain (Sullivan et al. 1999). However, further measurements are needed in order to better constrain the range of regolith thicknesses across Europa. Thus, Miranda's cratered terrain is likely mantled by one of the thickest layers of regolith on any planetary body where regolith has been observed and quantified (Figure 12).

4.2. Implications for Miranda's Geologic History

Multiple locations within Arden Corona have normal faults that appear to have been mantled by materials with albedos similar to that of the regolith on the surrounding cratered terrain (Hillier et al. 1989; Buratti & Mosher 1991), in particular in the region located around -45° latitude, 60° longitude (Figure 11(a), mapped as part of the Pm unit in Figure 2). The presence of regolith on Arden indicates that mantling of Miranda's surface may have been ongoing following the formation of this corona. Alternatively, the material in the Pm

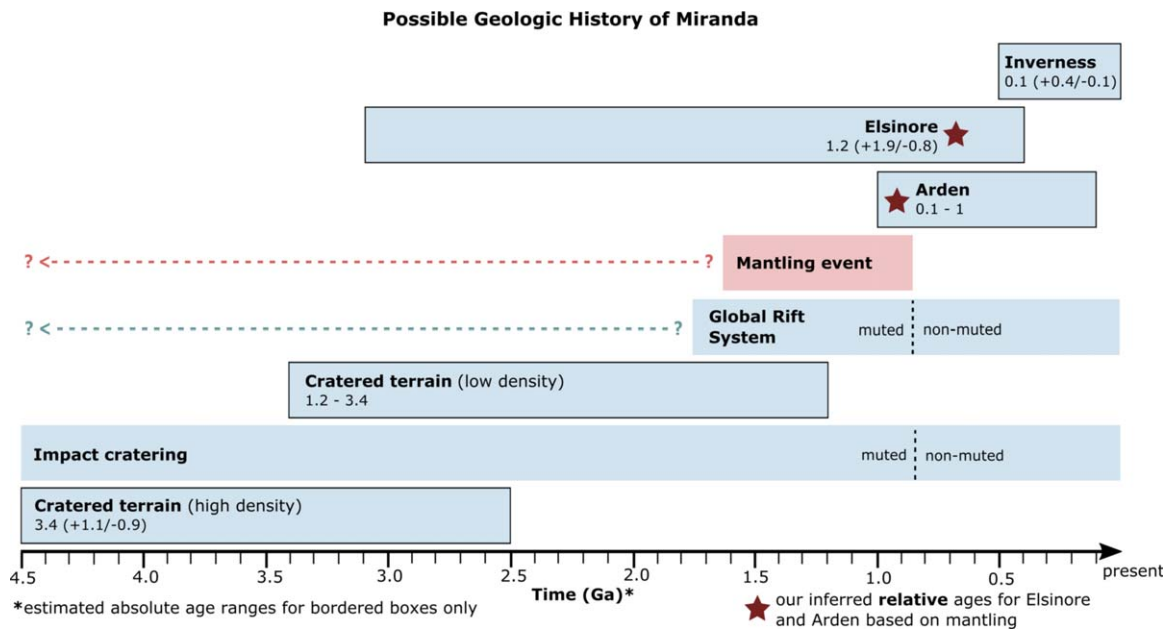


Figure 13. Timeline of Miranda's geologic history, updated to include the major mantling event (red). Ranges of absolute age estimates of terrains were originally reported in Kirchoff et al. (2022) and Zahnle et al. (2003). The boxes represent the range of possible formation periods for Miranda's different terrains and events. The order of events is based on both relative ages (non-bordered boxes) and absolute ages (bordered boxes; age estimates provided within the box are given in Ga). The bordered box boundaries represent the maximum and minimum ages estimates. Analysis of faults in the Global Rift System indicates that the mantling event occurred over a long period of time due to the presence of both muted and non-muted craters and scarps on the same terrains and proximal to one another. Regolith mantles some scarps within Arden but none in Elsinore, hinting that Arden Corona could be older than Elsinore Corona. This interpretation is contrary to previous relative age estimates of the coronae by Plescia (1988), Zahnle et al. (2003), and Beddingfield & Cartwright (2020), albeit their estimated absolute age ranges overlap.

unit might predate Arden and may have been incorporated into the corona, falling between exposed fault scarps, during formation. However, crater densities appear to be lower in the Pm unit than on the cratered terrain, and the mantling material appears to overprint the faces of individual scarps, suggesting mantling of Miranda's surface continued during Arden's formation.

Normal fault scarps within Arden Corona exhibit dip values that range from 27° to 43° (Beddingfield et al. 2015a). These dip values are within the expected range for the angle of repose ($\sim 15^\circ$ – 45° ; see Section 3.3). Therefore, it is unlikely that the topographic slopes associated with these faults have had a notable effect on the presence of regolith on these scarp faces.

Unlike Arden, neither Elsinore nor Inverness possess observable muted scarps or craters, suggesting that these two coronae formed after Arden. Therefore, we propose that Arden may be the oldest corona. In this scenario, the mantling event was ongoing during Arden's formation but ended before Elsinore or Inverness formed. Our interpretation contradicts the prediction that Elsinore is the oldest corona (Zahnle et al. 2003). As summarized in Zahnle et al. (2003), this interpretation was made because the majority of Elsinore's estimated age range is older than Arden's range. However, because these ranges of possible corona ages overlap (Figure 13), our interpretation that Arden is older does not contradict the younger end of Elsinore's range of estimated ages.

Here, we provide an updated possible geologic history of Miranda (Figure 13):

1. The regions of Miranda's surface that would become the high-density cratered terrain formed $3.4 (+1.1/-0.9)$ Ga (Kirchoff et al. 2022), and during this time, craters that would become muted began forming.

2. The parts of Miranda's surface that would become the low-density cratered terrain formed 1.2 to 3.4 Ga (Kirchoff et al. 2022).
3. The Global Rift System began forming, overprinting, and eventually being overprinted by craters that would become muted. These Global Rift System scarps would also become muted.
4. A major mantling event began blanketing Miranda's surface with a thick regolith. Craters and the existing Global Rift System faults became mantled with regolith, forming muted craters and scarps. The Global Rift System was still forming during this time, as evidenced by muted and non-muted fault scarps cutting muted craters (Figures 6(c) and (d)).
5. Arden Corona formed 0.1 to 1 Ga (Zahnle et al. 2003). Some scarps within Arden were mantled and became muted.
6. The mantling event ended. Faults within the Global Rift System continued to form, as evidenced by non-muted scarps.
7. Elsinore Corona formed $(1.2 \pm_{-0.8}^{+1.9})$ Ga (Kirchoff et al. 2022).
8. Inverness Corona formed $(0.1 \pm_{-0.1}^{+0.4})$ Ga; Kirchoff et al. 2022).
9. The Global Rift System remained active, as evidenced by large faults near Verona Rupes overprinting Inverness Corona (Figure 6(c)) and polygonal impact craters (PICs) overprinting Elsinore Corona (Beddingfield & Cartwright 2020).

Although this updated timeline provides insight into the timing of mantling processes, the source of Miranda's thick regolith remains unknown. We suggest three possible origin scenarios for Miranda's regolith: (1) burial by ejecta formed in

Table 4
Three End-member Hypotheses for the Source of Miranda's Regolith

Hypothesized Regolith Source	Supporting Evidence	Refuting Evidence
Giant Impact(s) (Section 4.3)	(1) Miranda's 4°3 orbital inclination (2) Localized concentration of possible secondary craters (3) Thick regolith—ejecta has been shown to create thick regolith (i.e., Vesta)	(1) No evidence for an impact basin, raised rim, concentric rings, or a central peak
Plumes (Section 4.4)	(1) Localized regolith overprinting Arden—if plumes are/were present and sourced from the coronae, they should be mantled by regolith (2) Miranda is spectrally blue—could result from small particles sourced from plumes	(1) Regolith not observed on Elsinore or Inverness—if plumes are/were present and sourced from the coronae, they should be mantled by regolith (2) Thick regolith—plume deposits have not been shown to create such a thick regolith (i.e., Enceladus)
Rings (Section 4.5)	(1) Regolith across entire imaged cratered terrain—distribution could be consistent with accumulation of ring deposits over time (2) Thick regolith—rings could provide a large source of material (3) Miranda is spectrally blue—could result from small ring particles, and consistent with Uranus' μ -ring	N/A

Note. All three processes, or combinations of two processes, may be responsible for generating Miranda's thick regolith. For simplicity, we only investigated end-member possibilities in this work.

a large impact event, (2) material sourced from Miranda's interior and deposited on its surface by plumes, and (3) accumulation of material sourced from the rings of Uranus (Table 4). These scenarios represent end-member processes, and it is plausible that one or more of these processes could have formed Miranda's regolith. We qualitatively consider these scenarios in the following subsections. More rigorous consideration and testing of these scenarios is beyond the scope of the paper and should be addressed by subsequent studies that build on the results and analyses reported here.

4.3. Regolith Formed by a Large Impact Event

A large impact event could have mantled Miranda with a layer of ejecta, forming the muted crater population (Table 4). By analogy, thick regolith layers have been identified on the surfaces of many asteroids (e.g., Helfenstein et al. 1994, 1996; Veverka et al. 2001; Saito et al. 2006; Sierks et al. 2011), including Vesta (Denevi et al. 2016). Regolith on Vesta is at least 1 km thick in some locations and has been largely attributed to ejecta blankets associated with impact basins (Jaumann et al. 2012; Denevi et al. 2016). Previous work has suggested that a large impact event on Miranda may have created the basin in which Arden Corona formed (Croft 1987; Croft & Soderblom 1991; Greenberg et al. 1991). Furthermore, a giant impact could have altered Miranda's orbit, and perhaps this could explain its 4°3 inclination. As summarized in Greenberg et al. (1991), small craters on Miranda with diameters <1.5 km show CSFDs similar to secondary craters (Section 3.2), which might have resulted from ejecta that initially escaped Miranda's gravity, formed a co-orbital debris disk, and subsequently fell back onto its surface.

There are no apparent impact structures (e.g., rims, concentric rings, or a central peak) surrounding Arden or elsewhere that suggest Miranda has experienced a giant impact event in the past. However, an unobserved impact basin might be present on Miranda's unimaged northern hemisphere.

Although a giant impact could explain Miranda's somewhat inclined orbit, other mechanisms, such as mean motion resonances (MMRs) and/or secular resonances shared by the Uranian satellites could also explain the intricacies of Miranda's orbit (Tittlemore & Wisdom 1990; Ćuk et al. 2020). By analogy to muted craters on Vesta, which are proximal to large impact basins, we would expect Miranda's muted craters to be concentrated near the source of impact ejecta (i.e., Arden), but muted craters are located across the southern hemisphere. Thus, we do not think the observational evidence best supports a giant impact scenario, but we cannot rule out the possible presence of large impact structures on Miranda's unimaged northern hemisphere. Future statistical modeling work is needed to determine the probability of a large impact event occurring on Miranda. In addition, simulations of giant impact events are needed in order to determine whether the amount of ejecta generated could explain Miranda's thick regolith.

4.4. Plume Deposits

Perhaps Miranda's regolith is sourced from cryovolcanic plumes (Table 4). The process(es) that formed Miranda's coronae may include cryovolcanism (e.g., Schenk 1991; Beddingfield & Cartwright 2020). In this scenario, perhaps the coronae exhibited plume activity as they developed, burying Miranda's pre-existing cratered terrain in material that fell back to its surface, thereby forming muted craters and scarps. The presence of regolith overprinting localized regions in Arden Corona could be consistent with a plume origin. A possibly analogous plume-mantling process could be responsible for muted craters on Enceladus (Figure 4(d)) (e.g., Kirchoff & Schenk 2009). Enceladus' muted craters are clustered near the active plumes emanating from the Tiger Stripes in the South Polar Terrain.

By analogy to Enceladus, if Miranda's muted craters were buried by plume material, then we would expect them to be

spatially correlated with one or more of the coronae, and the degree of crater and scarp mantling should decrease with distance from the coronae. This prediction is not consistent with the spatial distribution of Miranda's muted craters and scarps. Furthermore, Enceladus' muted craters exhibit a significantly thinner regolith than Miranda's muted craters (Figure 12). If plume deposits buried Miranda's muted craters, then this activity persisted for a much longer period time and/or it was much more vigorous than the current activity seen on Enceladus.

The surface of Miranda is spectrally blueish (i.e., slight decrease in reflectance with increasing wavelength, $\sim 0.4\text{--}1\ \mu\text{m}$, Karkoschka 2001), which might be expected for a surface mantled by small particles sourced from plumes. However, Miranda's blueish spectral slope can be explained by other processes, such as accumulation of fine-grain ring particles. Therefore, we find that the evidence does not best support a plume origin for Miranda's thick regolith cover.

4.5. Ring Material Accumulation

A possible explanation for the origin of Miranda's thick regolith is burial by material that originated in the rings of Uranus (Table 4). Uranus' outermost μ -ring is dusty, diffuse, and possibly composed of small H_2O ice grains (de Pater et al. 2006; Showalter & Lissauer 2006). The μ -ring spans an observed orbital radius of $\sim 86,000\text{--}103,000\ \text{km}$, encompassing the ring moon Mab (semimajor axis $\sim 97,700\ \text{km}$), which could sustain the μ -ring via ejection of dust particles from micrometeorite collisions with its surface (e.g., de Pater et al. 2006; Showalter & Lissauer 2006; Sfair & Winter 2012). The μ -ring has a prominent blue spectral color, attributed to the presence of small H_2O ice particles (submicron- to micron-sized diameters; de Pater et al. 2006; Showalter & Lissauer 2006). Material in the μ -ring could spiral inward and mantle the ring moon Puck (semimajor axis $\sim 86,000\ \text{km}$), thereby explaining the significantly bluer spectral slope of Puck's leading hemisphere compared to its trailing hemisphere (French et al. 2017).

Along with inward migration of μ -ring particles, perhaps some of this material spiraled outward on expanding orbits, due to electrically charged interactions with plasma in Uranus' magnetosphere and solar ultraviolet (UV) photons (Grundy et al. 2006), and eventually crossed into Miranda's orbital zone and collided with it (Cartwright et al. 2018, 2020c). The accumulation of small H_2O ice grains on Miranda is consistent with its H_2O ice-rich surface (e.g., DeColibus et al. 2022) and could explain its slight blue color (Karkoschka 2001). Although the accumulation of material from the μ -ring might explain some of Miranda's spectral properties, it is unlikely that the present-day μ -ring, sustained by tiny Mab (6–12 km diameter), could be the source of Miranda's substantial regolith (Figure 12).

An alternative possibility is that Miranda, along with Uranus' rings and ring moons, formed out of a massive Roche-interior ring of Uranus (Hesselbrock & Minton 2019). Uranus' high obliquity could have been caused by a giant impact with an Earth-sized object that tilted Uranus on its side and resulted in catastrophic disruption of its primordial satellite system (e.g., Morbidelli et al. 2012; Hesselbrock & Minton 2019; Rufu & Canup 2022; Salmon & Canup 2022). Subsequently, an outer debris disk composed of the remnants of the primordial satellites would have formed around Uranus' former

equatorial plane and then migrated to its new equatorial plane. Ariel, Umbriel, Titania, and Oberon would have formed out of the material in this outer debris disk. Miranda and the ring moons, however, would have formed out of a massive inner debris disk, representing the remnants of the Earth-sized object that collided with Uranus. Forming in a different disk of material could explain why Miranda is less dense than the other classical moons, as well as why its surface composition is notably different from them (e.g., Grundy et al. 2006; Gourgeot et al. 2014; Cartwright et al. 2018, 2020c; DeColibus et al. 2022).

In this scenario, Miranda could have formed and then migrated outward through the remaining disk of material. Forming in and subsequently migrating through such a debris disk could have resulted in the accumulation of a large amount of material, mantling it with a layer of thick regolith and forming muted craters and scarps. Once Miranda had finished sweeping up debris in its orbit, muted crater formation would have ended and non-muted craters would begin to form. Additionally, subsequent pulses of material might have been generated when Uranus' densely packed ring moons collided, forming new clouds of debris that might have expanded outward and mantled Miranda. These later pulses of material could explain why mantling appears to have continued over long stretches of Miranda's geologic history, possibly overlapping the formation of Arden Corona.

Although this scenario is intriguing, there are few constraints on Miranda's formation from an inner debris disk, how long it would take for Miranda to clear out its orbit after forming, and how much material might reach Miranda from subsequent ring moon collisions interior to its orbit. Nevertheless, we favor ring material accumulation as a possible explanation because this scenario may explain the wide spatial extent and substantial thickness of Miranda's regolith, and it can also help explain why Miranda's spectral properties are different from the other classical moons. Future modeling work that investigates how much ring material could have been deposited on Miranda is needed in order to determine whether this process could be responsible for Miranda's thick regolith.

4.6. Miranda Science Priorities for a Uranus Orbiter

A flagship-class mission to the Uranian system is the highest recommended priority for outer solar system exploration made by the 2023–2032 Planetary Science and Astrobiology Decadal Survey (National Academies of Sciences, Engineering, and Medicine 2022). Optimal launch opportunities for a flagship would be in 2031 and 2032, so that a Jupiter gravity assist could be utilized to shorten cruise time. Additionally, the launch time would allow a spacecraft to arrive when lighting conditions are favorable for analyzing the northern hemispheres of Miranda and the other Uranian moons, which were pointed away from the Sun and in shadow at the time of the Voyager 2 flyby in 1986.

An orbiter that performs multiple close flybys of Miranda could provide higher-resolution images that would allow for a more complete investigation of its surface, including previously unimaged regions in the northern hemisphere (Beddingfield et al. 2021a; Cartwright et al. 2021; Leonard et al. 2021; Cohen et al. 2022). New close-up images would provide key data for investigating the origin and properties of Miranda's thick regolith. Furthermore, higher-resolution images may show evidence for structures such as pit chains, similar to those

identified on Enceladus in Cassini ISS images with resolutions around 75 m pixel^{-1} (Martin et al. 2017). Identification of these features on Miranda would allow for further constraints on the thickness of its regolith and whether it is spatially variable (i.e., thicker near one or more of the coronae).

Even higher-resolution images would be useful to investigate finer morphological features that may provide further insight into regolith physical properties. For example, on Vesta, Dawn Framing Camera images with resolutions ranging from 15 to 20 m pixel^{-1} were used to assess blocks within crater ejecta, spur-and-gully type features in crater walls, and small impact craters (e.g., Denevi et al. 2016). Assessment of impact crater densities for muted and non-muted craters, as well as of overprinting and cross-cutting relationships between terrains and tectonic structures, would provide better constraints on the timing of Miranda's major mantling event and the absolute ages of coronae.

An orbiter with a visible/near-infrared mapping spectrometer would allow us to determine whether NH_3 -bearing species are spatially associated with Miranda's coronae, providing additional insight into the ages of its different terrains and the composition of its surface, lithosphere, and deeper interior (Cartwright et al. 2020b, 2021). A magnetometer on a Uranus orbiter could search for induced magnetic fields in Miranda and its neighboring satellites, which could emanate from briny liquid H_2O oceans (Cochrane et al. 2021; Weiss et al. 2021). Additionally, higher-resolution images of Ariel, Umbriel, Titania, and Oberon would provide opportunities for investigating the thickness of possible regolith on their cratered terrains. Thus, a Uranus orbiter, making multiple close flybys of the classical moons, would dramatically improve our understanding of their surfaces, interiors, and orbital histories, and could help us determine whether they are or were ocean worlds (Cartwright et al. 2021).

5. Conclusions

We analyzed muted and non-muted craters and scarps within Miranda's cratered terrain. Based on their morphologies, we found that Miranda's muted craters are most similar to craters that have been mantled by regolith on other planetary bodies, as opposed to craters that have been eroded or experienced viscous relaxation. We used three techniques to estimate regolith thickness: (1) Analysis of muted crater depth–Diameter (d - D) ratios near the South Polar Terrain Chasma indicates that the regolith mantling their floors ranges from 0.3–1.2 km thick. This range may result from differences in crater ages, and the true thickness may be similar to the highest estimate (1.2 km). (2) Analysis of crater size–frequency distributions across the cratered terrain indicates a thickness of $1.0 \pm 0.2 \text{ km}$. (3) Analysis of a central mound within Alonso Crater indicates a thickness of $1.4^{+0.3}_{-0.4} \text{ km}$ near Verona Rupes and may represent an upper limit.

Our estimates indicate that Miranda may have one of the thickest regoliths in the solar system, which has important implications for Miranda's interior thermal properties. Patches of regolith appear to mantle some scarps within Arden but not Elsinore or Inverness, indicating that Arden may be the oldest corona, contrary to previous relative age estimates. In this scenario, the mantling event was ongoing during Arden's formation but before Elsinore or Inverness formed.

We considered three different scenarios to explain the origin of Miranda's regolith, including deposition of ejecta from a

giant impact event, deposition of plume material associated with the formation of Miranda's coronae, and deposition of material from the rings of Uranus. We favor the ring hypothesis, which is consistent with the regolith's large spatial extent and substantial thickness, as well as Miranda's slightly spectrally blue color. Although we present a key first step for understanding Miranda's regolith, future projects are needed in order to more rigorously test these origin scenarios. Furthermore, higher-resolution images collected by a Uranus orbiter would dramatically improve our ability to constrain the properties of Miranda's regolith, image its unseen northern hemisphere, and investigate the source of the major mantling event.

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Appendix

We used various Voyager 2 ISS images to investigate Miranda. The unannotated version of Figure 6 is shown in Figure A1. The locations of images shown in Figure 6 are shown in Figure A2. To investigate the topography of Miranda's impact craters, we generated a digital elevation model (DEM) using Voyager 2 ISS images that cover this feature and the surrounding terrain (see Figure A2). The DEM was generated using the Ames Stereo Pipeline (ASP) software (Broxton & Edwards 2008; Moratto et al. 2010). We used a stereographic projection for the DEM, and the Voyager 2 ISS images were first pre-processed using the Integrated Software for Imagers and Spectrometers (ISIS; Anderson et al. 2004). The steps we followed to pre-process the images for input into ASP are described in Beddingfield et al. (2015a) and Beddingfield et al. (2015b). We applied photoclinometry (PC) techniques described in Alexandrov & Beyer (2018), which were utilized and discussed in previous work that generated DEMs for Miranda and Ariel using Voyager 2 ISS images (Beddingfield et al. 2022a, 2022b). We provide the topographic profiles for each crater analyzed within the cratered terrain near Inverness (Figure A3) for non-muted craters (Figure A4) and muted craters (Figure A5).

To assess the accuracy of DEMs, we used shadow measurement techniques, as discussed in Dameron & Burr (2018), to independently estimate the heights of ridges, where shadows were present and measurable in the Voyager 2 ISS images. The shadow measurements were conducted in order to calculate the topographic height, h , of a ridge. This value is given by $h = \frac{s}{\tan(\Phi)}$, where s is the observed shadow length and Φ is the solar incidence angle. Based on slight differences in ridge heights estimated using these two different techniques, we estimate that the topographic uncertainties associated with the DEM is approximately 20 m per 1 km in elevation. The DEM horizontal resolution is equivalent to the resolution of the input Voyager ISS image with the highest resolution.

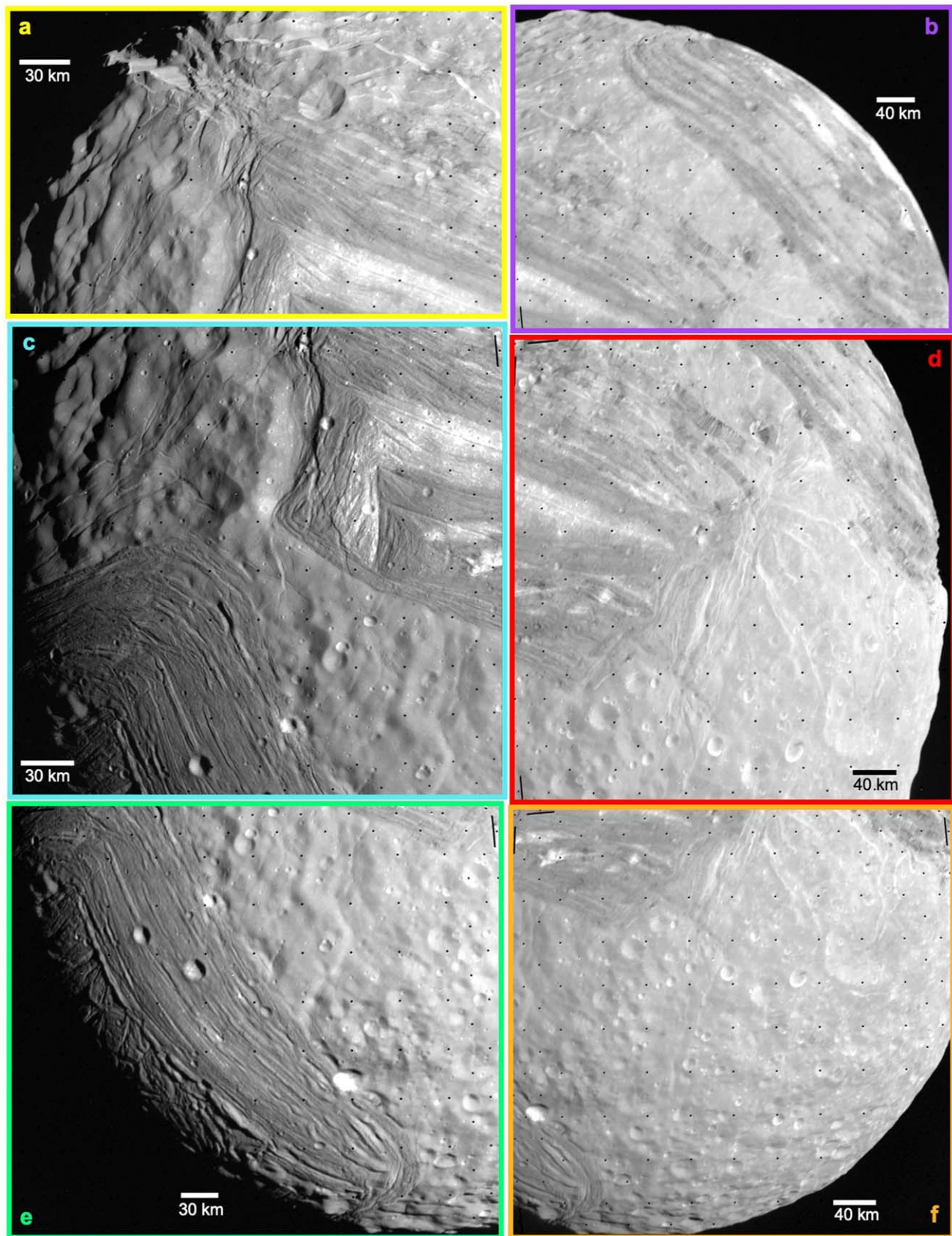


Figure A1. Unannotated version of Figure 6. See Figure A2 for the specific regions covered by each image.

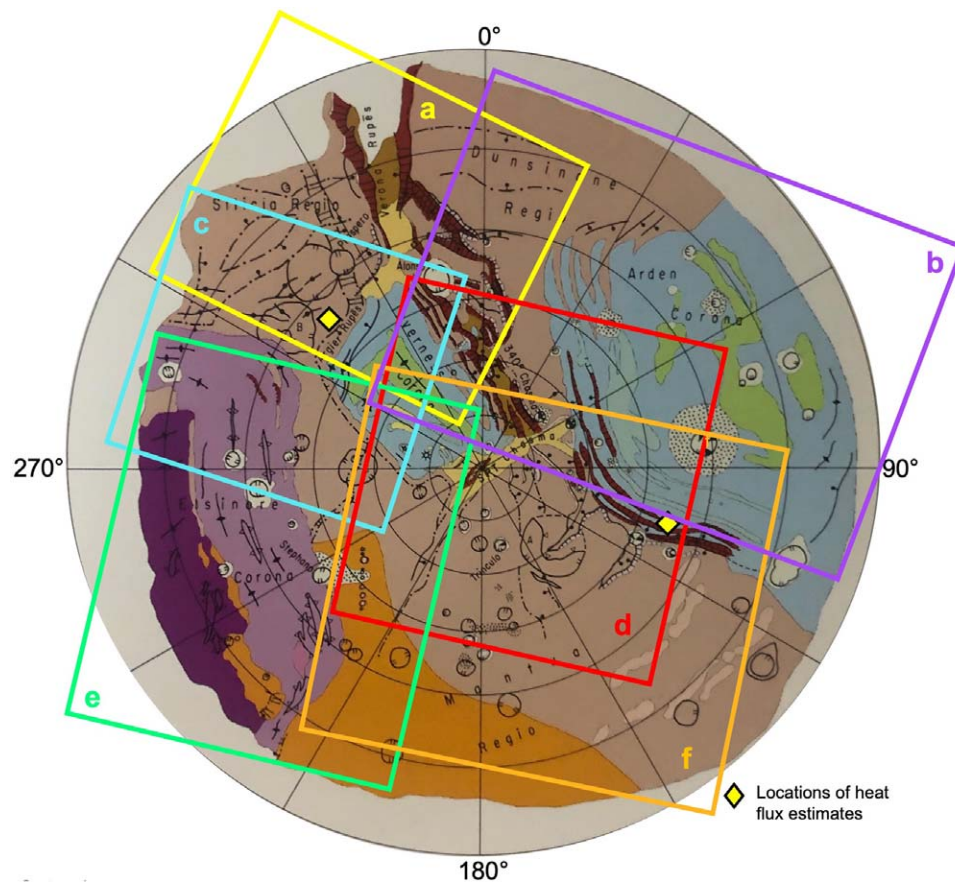


Figure A2. Annotated version of the map shown in Figure 2, presenting the locations of all images shown in Figure 6. See Figure 2 for legend.

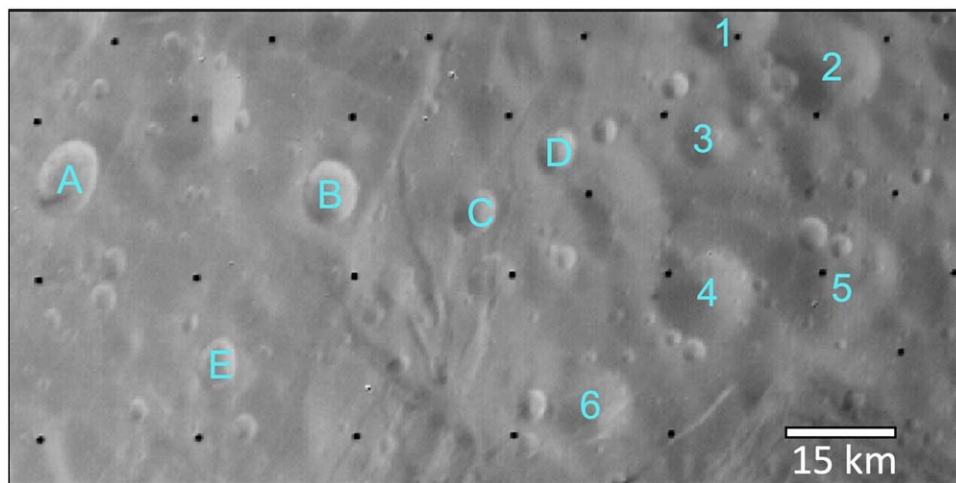


Figure A3. Locations of impact craters analyzed (shown in a portion of image c2684626, Figure 6(c)) in a digital elevation model (DEM) created for this work. Topographic profiles for less muted craters (A–E, Figure A4) and more muted craters (1–6, Figure A5) were generated.

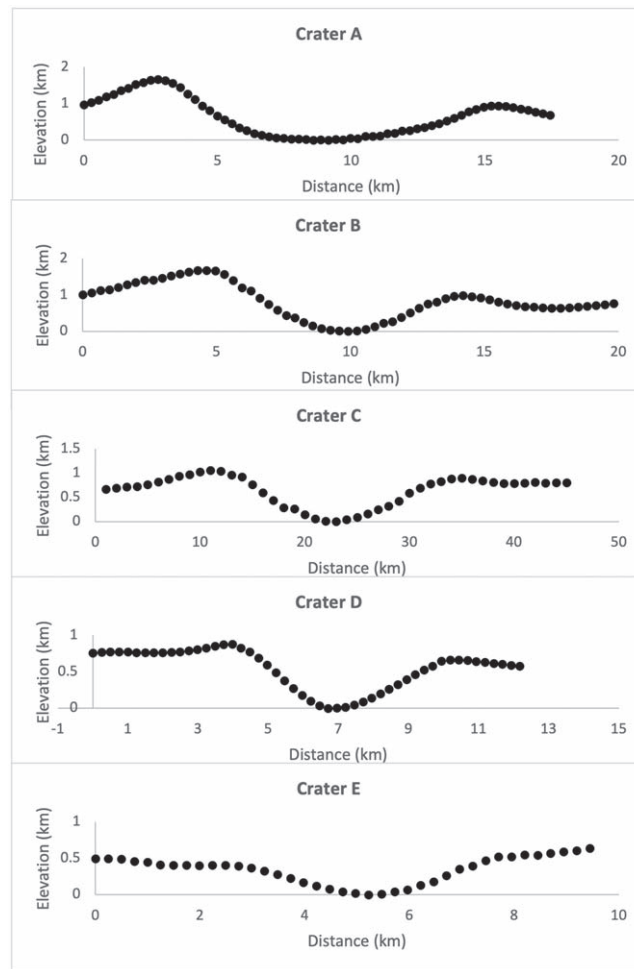


Figure A4. Topographic profiles for less muted craters A–E shown in Figure A3.

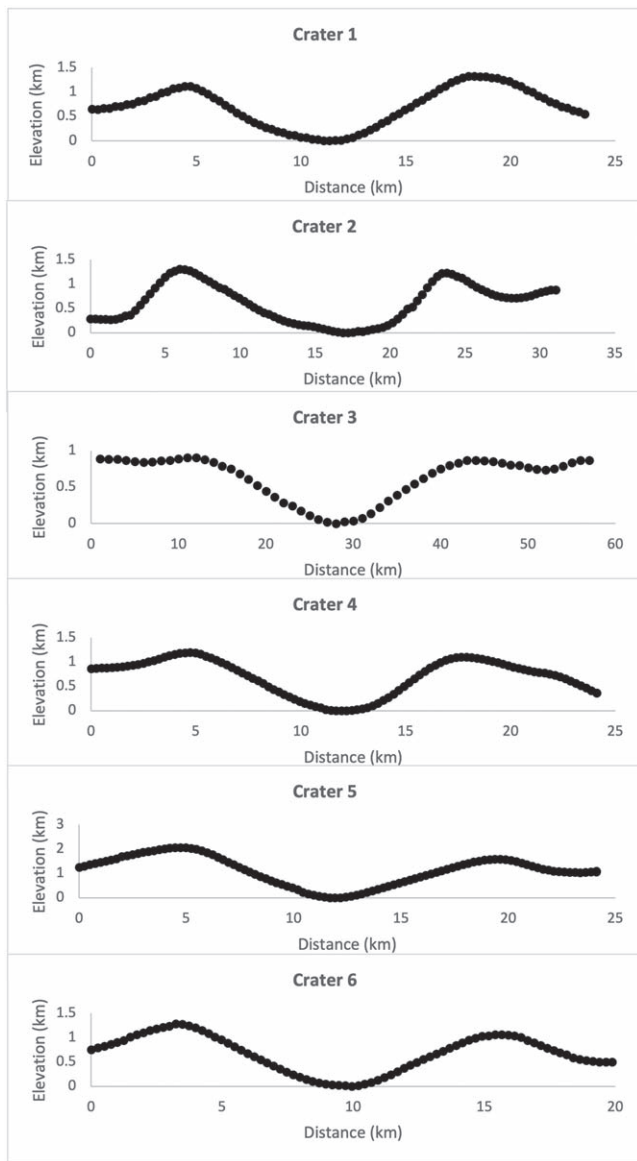


Figure A5. Topographic profiles for more muted craters 1–6 shown in Figure A3.

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References

- Alexandrov, O., & Beyer, R. A. 2018, *E&SS*, **5**, 652
- Anderson, J. L. B., Cintala, M. J., Cline, C. J., II, et al. 2021, *LPSC*, **52**, 1023
- Anderson, J. L. B., Dechant, L. E., Cline, C. J., II, et al. 2020, *LPSC*, **51**, 2791
- Anderson, J., Sides, S., Soltesz, D., Sucharski, T., & Becker, K. 2004, *LPSC*, **35**, 2039
- Atwood-Stone, C., & McEwen, A. S. 2013, *GeoRL*, **40**, 2929
- Barlow, N. G. 1988, *Icar*, **75**, 285
- Barlow, N. G. 1995, *JGR*, **100**, 23307
- Barlow, N. G. 2004, *GeoRL*, **31**, L05703
- Bart, G. D. 2014, *Icar*, **235**, 130
- Bart, G. D., Nickerson, R. D., Lawder, M. T., & Melosh, H. J. 2011, *Icar*, **215**, 485
- Basilevsky, A. T., & Shingareva, T. V. 2010, *SoSyR*, **44**, 38
- Beakawi Al-Hashemi, H. M., & Baghabra Al-Amoudi, O. S. 2018, *Powder Technology*, **330**, 397
- Beddingfield, C., Burr, D., & Emery, J. 2015a, *Icar*, **247**, 35
- Beddingfield, C., Li, C., Atreya, S., et al. 2021a, *BAAAS*, **53**, 121
- Beddingfield, C. B., Beyer, R. A., Singer, K. N., et al. 2020, *Icar*, **335**, 113383
- Beddingfield, C. B., Burr, D. M., & Dunne, W. M. 2015b, *JGRE*, **120**, 2053
- Beddingfield, C. B., Burr, D. M., & Tran, L. T. 2016, *Icar*, **274**, 163
- Beddingfield, C. B., & Cartwright, R. J. 2020, *Icar*, **343**, 113687
- Beddingfield, C. B., Cartwright, R. J., Leonard, E. J., Nordheim, T. N., & Scipioni, F. 2022a, *PSJ*, **3**, 106
- Beddingfield, C. B., Leonard, E., Cartwright, R. J., Elder, C., & Nordheim, T. A. 2022b, *PSJ*, **3**, 174
- Beddingfield, C. B., Moersch, J. E., & McSween, H. Y. 2018, *Icar*, **314**, 345
- Beddingfield, C. B., Moersch, J. E., & McSween, H. Y., Jr. 2021b, *Icar*, **370**, 114678
- Bland, M., Nordheim, T., Patthoff, D., & Vance, S. 2022, *LPICo*, **2678**, 1140
- Bland, M. T., Becker, T. L., Edmundson, K. L., et al. 2018, *E&SS*, **5**, 604
- Bland, M. T., Singer, K. N., McKinnon, W. B., & Schenk, P. M. 2012, *GeoRL*, **39**, L17204
- Bland, M. T., Singer, K. N., McKinnon, W. B., & Schenk, P. M. 2017, *Icar*, **296**, 275
- Broxton, M., & Edwards, L. 2008, *LPSC*, **39**, 2419
- Buratti, B. J., & Mosher, J. A. 1991, *Icar*, **90**, 1
- Carrigy, M. A. 1970, *Sedim*, **14**, 147
- Cartwright, R., Beddingfield, C., Showalter, M., Cruikshank, D., Nordheim, T., et al. 2020a, *LPSC*, **51**, 1699
- Cartwright, R. J., Beddingfield, C., Nordheim, T., et al. 2020b, *ApJL*, **898**, L22
- Cartwright, R. J., Beddingfield, C. B., Nordheim, T. A., et al. 2021, *PSJ*, **2**, 120
- Cartwright, R. J., Emery, J. P., Grundy, W. M., et al. 2020c, *Icar*, **338**, 113513
- Cartwright, R. J., Emery, J. P., Pinilla-Alonso, N., et al. 2018, *Icar*, **314**, 210
- Castillo, J. C., Weiss, B. P., Beddingfield, C. B., et al. 2022, *JGRE*, submitted
- Cathles, L. M. 2015, *Viscosity of the Earth's Mantle*, Princeton Legacy Library, Vol. 1362 (Princeton, NJ: Princeton Univ. Press) doi:10.1515/9781400867684
- Chuang, F. C., & Greeley, R. 2000, *JGR*, **105**, 20227
- Cline II, C. J., & Cintala, M. J. 2022, *M&PS*, **57**, 1441
- Cochrane, C., Vance, S., Nordheim, T., et al. 2021, *JGRE*, **126**, e06956
- Cohen, I. J., Beddingfield, C., Chancia, R., et al. 2022, *PSJ*, **3**, 58
- Craddock, R. A., Maxwell, T. A., & Howard, A. D. 1997, *JGR*, **102**, 13321
- Croft, S. 1987, *LPSC*, **18**, 207
- Croft, S., & Soderblom, L. 1991, *Uranus* (Tucson, AZ: Univ. Arizona Press), 561
- Čuk, M., El Moutamid, M., & Tiscareno, M. S. 2020, *PSJ*, **1**, 22
- Dameron, A. C., & Burr, D. M. 2018, *Icar*, **305**, 225
- de Pater, I., Hammel, H. B., Gibbard, S. G., & Showalter, M. R. 2006, *Sci*, **312**, 92
- DeColibus, D. R., Chanover, N. J., & Cartwright, R. J. 2022, *PSJ*, **3**, 119
- Denevi, B. W., Beck, A. W., Coman, E. I., et al. 2016, *M&PS*, **51**, 2366
- Dombard, A. J., & McKinnon, W. B. 2006, *JGRE*, **111**, E01001
- Ferguson, S. N., Rhoden, A. R., & Kirchoff, M. R. 2022, *JGRE*, **127**, e07204
- French, B. M. 1998, *Traces of Catastrophe: A Handbook of Shock-Metamorphic Effects in Terrestrial Meteorite Impact Structures*, LPI Contribution No. 954 (Houston, TX: Lunar and Planetary Inst.)
- French, R. S., Showalter, M. R., de Pater, I., & Lissauer, J. J. 2017, *AAS/DPS Meeting*, **49**, 214.19
- Gourgeot, F., Dumas, C., Merlin, F., Vernazza, P., & Alvarez-Candal, A. 2014, *A&A*, **562**, A46
- Grant, J. A., & Schultz, P. H. 1993, *JGR*, **98**, 11025
- Greenberg, R. J., Croft, S. K., & Janes, D. M. 1991, *Uranus* (Tucson, AZ: Univ. Arizona Press), 693
- Grimm, R. E., & Solomon, S. C. 1988, *JGR*, **93**, 11911
- Grundy, W., Young, L., Spencer, J., et al. 2006, *Icar*, **184**, 543
- Hammond, N. P., & Barr, A. C. 2014, *Geo*, **42**, 931
- Helfenstein, P., Veverka, J., Thomas, P. C., et al. 1994, *Icar*, **107**, 37
- Helfenstein, P., Veverka, J., Thomas, P. C., et al. 1996, *Icar*, **120**, 48
- Hesselbrock, A. J., & Minton, D. A. 2019, *AJ*, **157**, 30
- Hillier, J., Helfenstein, P., & Veverka, J. 1989, *Icar*, **82**, 314
- Hirata, N., Miyamoto, H., & Showman, A. P. 2014, *GeoRL*, **41**, 4135
- Jaumann, R., Williams, D. A., Buczkowski, D. L., et al. 2012, *Sci*, **336**, 687
- Karkoschka, E. 2001, *Icar*, **151**, 51
- Katsnelson, A. 2021, *Natur*, **598**, 224
- Kirchoff, M. R., Dones, L., Singer, K., & Schenk, P. 2022, *PSJ*, **3**, 42
- Kirchoff, M. R., & Schenk, P. 2009, *Icar*, **202**, 656
- Kleinhans, M., Markies, H., De Vet, S., In't Veld, A., & Postema, F. 2011, *JGRE*, **116**, E11004
- Kreslavsky, M., & Head, J. 2015, *LPSC*, **46**, 1246
- Leonard, E. J., Elder, C., Nordheim, T. A., et al. 2021, *PSJ*, **2**, 174
- Lowe, D. R. 1976, *JSedR*, **46**, 188
- Marchi, S., Bottke, W. F., O'Brien, D. P., et al. 2014, *P&SS*, **103**, 96

- Martin, E. S., Kattenhorn, S. A., Collins, G. C., et al. 2017, *Icar*, **294**, 209
- McKay, D. S., Heiken, G., Basu, A., et al. 1991, Lunar Sourcebook, A User's Guide to the Moon (New York: Cambridge Univ. Press), 285
- Martin, E., Whitten, J., Kattenhorn, S. A., et al. 2022, *Icar*, submitted
- McKinnon, W. B., Chapman, C. R., & Housen, K. R. 1991, Uranus (Tucson, AZ: Univ. Arizona Press), 629
- Melosh, H. J. 1976, *LPSC*, **7**, 2967
- Moratto, Z. M., Broxton, M. J., Beyer, R. A., Lundy, M., & Husmann, K. 2010, *LPSC*, **41**, 2364
- Morbidelli, A., Tsiganis, K., Batygin, K., Crida, A., & Gomes, R. 2012, *Icar*, **219**, 737
- Nakashima, H., Shioji, Y., Kobayashi, T., et al. 2011, *Journal of Terramechanics*, **48**, 17
- National Academies of Sciences, Engineering, and Medicine 2022, Origins, Worlds, and Life: A Decadal Strategy for Planetary Science and Astrobiology 2023–2032 (Washington, DC: The National Academies Press) doi:10.17226/26522
- Osinski, G. R., & Pierazzo, E. 2013, Impact Cratering: Processes and Products (Hoboken, NJ: Wiley) doi:10.1002/9781118447307
- Otto, K. A., Marchi, S., Trowbridge, A., Melosh, H. J., & Sizemore, H. G. 2019, *JGRE*, **124**, 1188
- Pappalardo, R. T., Reynolds, S. J., & Greeley, R. 1997, *JGR*, **102**, 13369
- Parmentier, E. M., & Head, J. W. 1981, *Icar*, **47**, 100
- Passey, Q., & Shoemaker, E. 1982, Satellites of Jupiter (Tucson, AZ: Univ. Arizona Press), 379
- Passey, Q. R. 1983, *Icar*, **53**, 105
- Plescia, J. B. 1988, *Icar*, **73**, 442
- Pohlman, N. A., Severson, B. L., Ottino, J. M., & Lueptow, R. M. 2006, *PhRvE*, **73**, 031304
- Quaide, W. L., & Oberbeck, V. R. 1968, *JGR*, **73**, 5247
- Ramsley, K. R., & Head, J. W. 2021, *SSRv*, **217**, 86
- Reddy, V., Le Corre, L., O'Brien, D. P., et al. 2012, *Icar*, **221**, 544
- Robbins, S. J., & Hynek, B. M. 2012, *JGRE*, **117**, E05004
- Robbins, S. J., Watters, W. A., Chappelow, J. E., et al. 2018, *M&PS*, **53**, 583
- Rufu, R., & Canup, R. M. 2022, *ApJ*, **928**, 123
- Saito, J., Miyamoto, H., Nakamura, R., et al. 2006, *Sci*, **312**, 1341
- Salmon, J., & Canup, R. M. 2022, *ApJ*, **924**, 6
- Schenk, P., O'Brien, D. P., Marchi, S., et al. 2012, *Sci*, **336**, 694
- Schenk, P. M. 1989, *JGR*, **94**, 3813
- Schenk, P. M. 1991, *JGRB*, **96**, 1887
- Scott, R. F. 1967, *Icar*, **7**, 139
- Sfair, R., & Winter, S. G. 2012, *A&A*, **543**, A17
- Shkuratov, Y. G., & Bondarenko, N. V. 2001, *Icar*, **149**, 329
- Showalter, M. R., & Lissauer, J. J. 2006, *Sci*, **311**, 973
- Sierks, H., Lamy, P., Barbieri, C., et al. 2011, *Sci*, **334**, 487
- Smith, B. A., Soderblom, L. A., Beebe, R., et al. 1986, *Sci*, **233**, 43
- Solomon, S. C., Comer, R. P., & Head, J. W. 1982, *JGRB*, **87**, 3975
- Spencer, J. E. 1984, *Geo*, **12**, 95
- Strom, R. G. 1987, *Icar*, **70**, 517
- Sullivan, R., Moore, J., & Pappalardo, R. 1999, *LPSC*, **30**, 1747
- Thomas, P. J., & Schubert, G. 1988, *JGR*, **93**, 13755
- Tittemore, W. C., & Wisdom, J. 1990, *Icar*, **85**, 394
- Veverka, J., & Thomas, P. 1979, Asteroids (Tucson, AZ: Univ. Arizona Press), 628
- Veverka, J., Thomas, P. C., Robinson, M., et al. 2001, *Sci*, **292**, 484
- Weiss, B. P., Biersteker, J. B., Colicci, V., et al. 2021, *GeoRL*, **48**, e94758
- White, O. L., Schenk, P. M., Bellagamba, A. W., et al. 2017, *Icar*, **288**, 37
- White, O. L., Schenk, P. M., & Dombard, A. J. 2013, *Icar*, **223**, 699
- Yue, Z., Di, K., Liu, Z., et al. 2019, *Icar*, **329**, 46
- Zahnle, K., Schenk, P., Levison, H., & Dones, L. 2003, *Icar*, **163**, 263