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Short title: A climatic origin of crater-ice in Mars' North Polar Region

Abstract

This paper explores the origins and evolution of ice-rich interior mounds found within craters of the north polar region of Mars. We present a systematic study of impact craters above 65°N, and identify 18 craters that have interior mounds. At least eleven of these mounds are composed of water ice and geometric similarities suggest that dune-covered mounds may also have a water ice core. The mounds are found in the deeper craters in the north polar area and we suggest that these form a specific microclimate favorable for mound initiation and growth. It is likely that at least seven of the mounds have evolved as individual outliers, rather than conterminous with the main polar cap. Our observations suggest that the mounds are built up by atmospheric deposition, similar to that of the north polar layered deposits. Using a combination of remote sensing techniques enabling topographic, spectral, radar and image data analyses, we have documented the morphology, composition and stratigraphy of selected mounds. We advance and test four hypotheses for formation of these mounds: artesian outpouring from a deep aquifer, hydrothermal activation of ground ice, remnants of a more extensive polar cap, and atmospheric deposition on ice caps in meteorologically isolated locations. We propose that during periods when the perihelion was located in northern summer (most recently 10-25 ka before present) the microclimate in these craters retarded the sublimation of CO_2 and water ice in northern spring, thus creating a cold trap for volatiles released as the seasonal cap retreated. This created a thick enough deposit of water ice to withstand sublimation over the summer and initiate a positive feedback leading to mound-building. Mounds without complete dune-cover may be in dynamic equilibrium with the ambient climate and show evidence of both present-day and past periods of erosion and aggradation. We conclude that the water ice mounds formed in deep impact craters in Mars' north polar region may contain sensitive records of past polar climate that may enhance our understanding of the CO_2-H_2O system in the polar regions.

Keywords: Mars; ices; Mars, polar geology; Mars, climate; Cratering.

Highlights:

- 18 potentially ice-cored mounds were found in craters in Mars' north polar region.
- The stratigraphy of the mounds argues for deposition from the atmosphere.
- We argue many of them were deposited separately from the polar cap.
- The crater micro-environment is a potential explanation for mound initiation.
- These mounds are sensitive and important records of Amazonian climate on Mars.

1. Introduction

The northern lowlands of Mars have long been considered an important reservoir for water. It has been suggested that a northern ocean existed there in the Noachian and that water may now be present deep in the local subsurface (e.g., Clifford, 1993; Clifford and Parker, 2001; Perron et al., 2007). More pertinently, the northern polar cap, with an estimated volume of 1.3 x 10^6 km^3 (Selvans et al., 2010) consists of a mixture of water ice and dust, and a region of near-surface ground ice extends down to latitudes of 45° (Byrne et al. 2009). This was first shown by the hydrogen ion signature found by gamma ray spectrometry (Feldman et al., 2004; Jakosky et al., 2005) and later supported by spectral signatures from OMEGA (Bibring et al., 2005), the distribution of water surface frosts (Vincendon et al., 2010), the distribution of fluidized craters (Barlow and Perez, 2003) and the radar reflectivity properties of the surface (Mouginot et al., 2010), with ice found by the Phoenix lander (Mellon et al., 2008) providing a local ground truth. Seasonal cycles mobilize and redistribute these surface volatiles; principally those exposed on the polar cap. At present, the most noticeable seasonal change is the waxing and waning of the so called 'seasonal' polar caps, observable from Earth (e.g., Antoniadi, 1930). The northern hemisphere seasonal cap commonly extends down to 50°N and consists primarily of CO_2 ice. It has a thickness of ~ 0.5 m (Cull et al., 2010; Smith et al., 2001), and is sourced from the atmosphere by condensation of water followed by CO_2 (Ivanov and Muhleman, 2001). Deposition occurs during the polar night and so has not been directly observed, but is thought to involve atmospheric condensation possibly as snow from ubiquitous cloud-cover. The spring recession is better characterized and it comprises a gradual northward retreat first of the CO_2, then of the thinner water-ice annulus (Bibring et al., 2005; Wagstaff et al., 2008).

After recession of the seasonal volatile deposits there remain several substantial outliers composed of water ice, which are spatially separated from the northern polar cap (Langevin et al., 2005; Tanaka et al., 2008). Many of these ice bodies are found within craters, e.g., Korolev, Dokka and Louth. They are morphologically distinctive (Garvin et al., 2000), forming an interior mound, which is convex-up and dome-shaped, with low local slopes and a moat that separates it from the crater walls. These mounds are distinct from an impact central peak, because they have convex-up positive relief in the center, often with a large volume placed asymmetrically within the crater. These mounds are generally assumed to be remnants of a previously more extensive northern ice cap (Garvin et al., 2000; Tanaka et al., 2008). Other possible origins include: upwelling from an underground aquifer (proposed but not supported by Russell and Head, 2002), activation of near-surface ground ice by impact-induced hydrothermal systems (a possibility considered for paleolake formation by Osinski et al., 2005; Rathbun and Squyres, 2002) and atmospheric condensation as individual outliers of the main cap (Brown et al., 2008). Although the latter invokes atmospheric deposition, as for the polar cap, it differs significantly from the formation of polar cap remnants in that 1) a larger extent of the polar cap is not required and 2) more importance is given to microclimatic effects inside the craters. Each of these formative mechanisms has important, but different, implications for the dynamics of Mars' hydrosphere and climate. For example, if they are indeed remnants of a formerly more extensive polar cap, then these mounds are important records of both its extent and the conditions needed to preserve this ice. If sourced from near-surface, or deep ground-ice, then the mounds give information on the distribution of this ice and an indication of the volumes of water stored in the Martian crust. If the mounds are supplied from deep sources, then this could support the presence of a deep, global hydrosphere (Clifford, 1993). And if they are individual cap outliers, then the mounds could be sensitive to climate perturbations and hence, their spatial distribution, morphology and internal structure can help place constraints on recent climate.

The aim of this study is to explore the origins and evolution of interior mounds that are found within some craters in the North Polar region of Mars. To this end, we document and interpret the distribution, morphology and internal structure of these mounds, finding that the majority have likely formed by atmospheric deposition triggered by microclimatic effects inside the host craters, separate from the main polar cap.

2. Approach

2.1 Craters in the north polar basin

We have based our impact crater survey on information from the Mars Orbiter Laser Altimeter (MOLA) gridded data at 256 and 128 pixels per degree for the north polar region, recording the locations and morphometric properties of all craters that are hydrologically intact including craters with internal mounds. This survey is complete for craters with a diameter >5 km, located north of 65°N (Barnie, 2006).

Impact craters were identified and digitized from MOLA data using a watershed analysis technique aimed at delineating crater rims (*cf.* Stepinski et al., 2009). Each crater was given a unique identifying number (Barnie, 2006; Hovius et al. 2009). Based on visual inspection of the Thermal Emission Imaging System (THEMIS) and Mars Orbiter Camera Narrow-Angle (MOC-NA) image mosaic data we identified craters containing dune fields. Dune fields were identified by morphology where image resolution was sufficient and elsewhere by a characteristic combination of high IR emission values in THEMIS IR daytime data and low reflection of visible light in THEMIS VIS.

For all impact craters in our catalogue, we obtained elevation statistics from the MOLA digital elevation model (DEM) including the maximum, minimum and range. The latter was used as an estimate of the overall crater depth. The mean crater diameter was calculated by using $D = 2 \sqrt{(A/\pi)}$, where D is the diameter and A the area of the digitized rim polygon. This method assumes the rim traces a circle and is not biased by topographic irregularities or slight obliquity of the crater-form. The distance from a crater to the polar cap was measured along the shortest straight line from the crater centroid to the edge of the contiguous Polar Layered Deposits and Polar Ice Deposits as mapped by Tanaka et al. (2005).

From our catalogue, we have isolated all craters containing raised central topography for further analysis. The following additional criteria were applied:

1. To identify and exclude central peaks: North-south and east-west topographic profiles were taken across these craters. Craters with central peaks were recognized by the small extent (< 10% of the crater floor) and central position of the rise and the similarity of rise and crater wall albedo. These craters were not considered further.
2. For the remaining craters where the presence of dark dunes dominated the albedo of deposits within a crater, we applied a minimum relief of 150 m in our identification of mounds. This is the maximum relief of inner-crater dunefields measured outside of the survey area, and the cut-off eliminated only one mound from our survey (marked 332 in Fig. 2).

As the resulting mounds have geometries and surface attributes distinct from those of central peaks, they were deemed to be depositional in origin. We digitized the limit of the mounds using the topographic inflexion at the transition to concave-up crater topography as a guide. Mound volume was estimated from surface topography and an extrapolation of the crater floor below the mound (not including any estimate of a central peak). This involved rotating through 360° the power-law best fit to the median radial profile of the crater interior (excluding any areas containing the mound) and differencing this estimated surface with the MOLA topography to generate an isopach map for each mound. Using this map we also calculated the mound asymmetry by measuring the horizontal offset between the center of mass of the mound and the centroid of the crater rim.

2.2 Mound Composition

We have investigated the composition of surface materials on the crater mounds using data from Mars Express' Observatoire pour la Minéralogie, l'Eau, les Glaces et l'Activité (MEX OMEGA) imaging spectrometer and Mars Reconnaissance Orbiter's Compact

Reconnaissance Imaging Spectrometer for Mars (MRO CRISM) targeted observations to compute simple diagnostic ratios, including band depth and drop-off. OMEGA has a nearly complete coverage of the north polar region at a spatial resolution of ~3 – 7 km/pix and also covers selected areas at 0.5 - 1.5 km/pix. CRISM targeted observations have much lower global coverage, but a very high spatial resolution of 15.7 - 19.7 m/pix. Both datasets underwent standard atmospheric correction before spectral analysis (“volcano-scan” algorithm, cf. Morgan et al., 2009). Although aerosol effects, not accounted for in this correction, can be important in the north polar region (Vincendon et al. 2006; Vincendon 2008), the strong reflectivity of water and CO_2 ice overwhelms the atmospheric effects allowing reliable identification (Brown et al., 2010). Datasets from northern summer (Solar longitude, L_s 90 - 120°) were used to avoid seasonal CO_2 surface frost, which could mask any spectral signals from the material beneath (Bibring et al., 2005). The relative depth of the absorption band at 1.50 µm was used to estimate the water ice content. The depth of the absorption band at 2.35 µm was used to estimate the quantity of carbon dioxide ice (Brown et al., 2010).

2.3 Internal Stratigraphy

The stratigraphy of mound materials was investigated with image data of surface outcrops and radar data capturing the interior stratigraphy. Surface stratigraphic information was obtained from images collected with THEMIS, High Resolution Stereo Camera (HRSC on Mars Express), MRO Context (CTX) and MRO High Resolution Imaging Science Experiment (HiRISE). All these images were either already geometrically corrected (HiRISE, HRSC), or corrected using routines in the software ISIS3. Alignment with the MOLA DEM was manually checked by verifying that local highs and local lows (e.g., crater rims, troughs) corresponded between the image and the MOLA data. For specific examples we measured the distance between the layers in plan-view. We assumed that the layers were approximately flat-lying and thus corrected the measured distances to account for exposure angle (which makes the plan-view width of layers wider than their true thickness if the slope is less than 45°). This was done by measuring the local slope, derived from MOLA topography, in the same orientation as the layer thickness measurements. As the length scale of the MOLA-derived slope is approximately ten times larger than that of the measured layers, the slope correction introduces a degree of error, however it affects all measurements and no better source of topography is available at present. For comparison purposes, we made equivalent measurements of layer thickness in three exposures of the Polar Layered Deposits (PLD) in HiRISE images PSP_010366_2590, TRA_000825_2665 and TRA_000863_2640 and one exposure of the Basal Unit (BU) in HiRISE image TRA_000863_2640.

At selected locations we measured the junction angles of angular unconformities and/or discontinuities exposed in outcrop, and ascertained younging directions from the layer geometry. Where possible, dip and strike measurements of large-scale layering were made by employing the geometric relationships between individual layers and 25 or 50 m contours derived from the MOLA 512 and 256 pixels per degree data (~115 and ~230 m/pix, respectively). Dips and strikes were measured over a length-scale of kilometers, so should be considered as an area average value. We estimate the error in the dip measurement to be 1-2° as detailed below.

Uncertainty in the dip measurement derives from: 1) errors in the layer digitization, 2) human errors in the distance measurements and 3) errors in the contours. We consider the layers to be digitized to within 2 pixels of their actual location on the images (50 cm for HiRISE and 12 m for CTX). We estimate the horizontal uncertainty of our distance measurements due to human error as being between 10 and 50 m, thus greater than the digitization uncertainty. Horizontal errors of this magnitude (over lengthscales of kilometers) account for at worst a 2° error in dip, but usually ±1°. Contour errors can originate from interpolated pixels in the MOLA DEM and misalignment of the DEM with the images. The interpolated pixels are a relatively small source of error, because we performed these measurements over smooth, continuous surfaces for which interpolation performs well, resulting in smooth contours with regular intervals. The potential horizontal misalignment between the images and DEM dominates the error and we estimate it is of the order of 1-2 MOLA pixels (460 m at worst, but more likely 100-200 m). However this does not necessarily change the horizontal distance between the contours (and thus the dip measurement) unless the slope changes significantly over distances of 460 m or less in the region of measurement. For the majority of measurements this was not the case. However, we took three examples where this was the case (one in Korolev and two in crater 663) and shifted the contours in relation to the images by ±460m in the direction of greatest change in slope. This simulated “worse-case” misalignment resulted in an error in the measured dip angle of ±1.5°.

We inspected all available MRO Shallow Radar (SHARAD) data intersecting craters with mounds for signs of internal structure. Where we found structure within mounds we performed clutter simulations using MOLA data to confirm that observed structures were not due to surface topographic effects (Russo et al., 2008). To assess layer spacing and thickness of mound deposits we converted the two-way-time into depth (cf. Plaut et al., 2009) using a permittivity of 3.15 as previously used by Putzig et al. (2009) for the nearly pure ice of the polar cap.

3. Results

3.1 Northern plains craters morphology and distribution

We found 397 craters with a diameter > 5 km north of 65°N. The depth-diameter relationship for these craters is shown in Fig. 1. The relationship is described by $d = 0.05D^{0.98}$, where d is crater depth (km) and D is crater diameter (km). For comparison, Garvin et al. (2000) found $d = 0.03D^{1.04}$ for 109 impact craters north of 57°N. These relationships differ significantly from the global depth-diameter relationship for pristine craters as determined by Garvin (2005). Compared to the polar depth-diameter relationships, the global depth-diameter relationship predicts deeper craters for a given diameter, apart from for the very largest diameter craters. The same trend has been noted by Boyce and Mouginis-Mark (2005) and Kreslavsky and Head (2006). This suggests that in the north polar area either, craters have been subject to significant infilling, or that the properties of local substrate differ substantially from the rest of Mars such that on formation impact craters are shallower compared to the global population. From morphological evidence and the existence of a small population of deep, young craters both Boyce and Mouginis-Mark (2005) and Kreslavsky and Head (2006) conclude that craters in the north polar area are infilled.

At least 38 craters with D > 5 km contain dune fields. Dune fields were found in 28% of craters > 400 m deep (Fig. 1), but not in shallower craters. Such dune fields are not restricted to the polar region; we have found them in 24% of craters between 50°N and 65°N. At these latitudes, the relief of dune fields above the crater floor usually does not exceed 150 m and the deposits usually cover relatively small areas within a crater. At higher northern latitudes some craters have extensive dune cover. This includes some craters with depositional mounds with relief in excess of150 m.

According to our criteria, eighteen craters were found to contain mounds (Fig. 2). They are located between 70°N and 82°N, and none are smaller than 9.5 km in diameter (Table 1). Seven are adjacent to the polar cap and the furthest is located > 600 km from the cap. No craters with mounds were found between 65 and 70°N. Many of the mounds consistently have high albedo in summer months (e.g., Armstrong et al., 2007; Garvin et al., 2000). However, many other craters in the north polar region that do not contain interior mounds also have this attribute (Calvin et al., 2009; Seelos et al., 2008; and informally by the MRO HiRISE team). Craters with mounds lie within the general depth-diameter distribution of craters in this region (Fig. 1), but for a given diameter they are relatively shallow compared to other craters of the same diameter. This can be accounted for only in part by the presence of a mound (predicted depths are given in Table 1 and shown on Figs. 1 and 3). Moreover, mounds are not found in all craters of any particular diameter and not all craters at a given latitude with a similar diameter contain a mound. In fact only 24% of all craters located above 70.1°N with diameters greater than 9.5 km have interior mounds.

Figure 3 shows two representative examples of impact craters with an interior mound (all the craters are shown in Fig. S1). Firstly, Dokka (388) with a clean ice mound, like Korolev (206) and six other, smaller craters in our survey and secondly crater 811 with a mound-shape similar to Dokka's, but entirely covered by dunes ("complete dune cover" on Fig. 2). Crater 814 shares this characteristic. Four other craters have complete dune coverage, but their mounds do not span the whole crater floor (480, 882, 904 and 934 – "dune covered mounds" on Fig. 2). There are four mounds with dune partial cover, including one in crater 503, Louth. In all cases dune patches coincide with the topographic high and only in the cases where there is complete dune cover do the dunes ingress into the moat. Despite the different levels of dune cover the mounds share the following morphological characteristics (Fig. S1): flat or gently sloping summit, asymmetrical placement and a partial or complete moat separating the mound from part or all of the crater wall.

At the scale of MOLA resolution (~230 m) the mounds are smooth (except where dunes are present, Fig. 3) and gently domed, with maximum local slopes of up to 34° and a half-range modal slope for all the mounds of 0.9° (mean of 3.6°). Mound relief ranges from 103 m to 1818 m above the current crater floor and mounds extend to within 13 – 1388 m below the crater rim, filling their host to 9 – 89 % of current crater depth from rim to floor (Table 1). Pole-centered, radial swath profiles including the craters containing mounds illustrate the local penetration depths of these craters into the crust and the relative heights of the mound summits (Fig. 4). The longitudinal extent of the swaths was chosen to minimize the longitudinal variation in elevation, so the penetration depth of the craters can be clearly seen. Figure 4 shows that the craters with mounds tend to be within the craters that penetrate to the deepest level for a given latitude and longitude region, but also that not all of the deeply penetrating craters have mounds. This figure also demonstrates that there is no characteristic height up to which the mounds extend vertically.

Estimates of maximum ice thickness range from 94 to 1890 m with a mean of 605 m, and mound volumes range from 0.8 to 3850 km^3, with a mean of 315 km^3 (Table 1 and Fig. 5). Our volume estimates are higher than those calculated by Garvin et al. (2000) for the largest craters. For example Garvin et al. (2000) listed volumes of 1356 and 463 km^3 for the infill of Korolev and Dokka, respectively and our corresponding estimates are 3850 and 1320 km^3 (Table 1). The discrepancy can be accounted for by the difference of methods used in this study and that of Garvin et al. (2000). Garvin et al. (2000) used single MOLA profiles to determine a polynomial fit to the cavity interior and constrained the predicted depths using empirical relationships between depth and diameter derived from 109 craters in the region. We used all available MOLA data to provide information on the cavity shape and did not constrain the crater depth.

Mounds are either completely or partially surrounded by a moat. Similar moats have been found around concentric crater fill (CFF) near the Martian dichotomy boundary (Levy et al., 2010). However, CFF is flat-topped, rough at MOLA resolution (~ 20-80 m of relief) and has not been found to extend above ~450 m below the crater rim. Westbrook (2009) described 75 craters in the south polar region, which also contain mounds. These mounds are similar to those described here in that they have a flat top, a distinct bounding slope (a moat), are often offset from the crater center, 35 of them have exposed ice and layers and 17 are topped by sand dunes. Notable differences include: the south polar mounds often have a bounding ridge encircling their flat top, some exhibit flow features, some have multiple superposed mounds, they are found down to lower latitudes (60.8°S) and their host craters are larger (diameters range from 18.6 to 114.1 km, mean 45.1 km). With the exception of the southern polar craters, mounds such as the ones described here have not been reported elsewhere on Mars.

3.2 Mound composition results

Spectral data acquired by OMEGA (Fig. 6) show that there are very high water-ice concentrations (up to ~0.7 band depth) at the surface of the mounds in Korolev (206), Dokka (388), Louth (503) and unnamed craters 544 and 579. The CO_2 ice signal of from the 2.35 μm band depth is less than 0.2, hence at the limits of the detection using this method. This indicates that the mounds are predominantly composed of water ice, similar to the inferred composition of the north polar cap (Bibring et al., 2005). The other mounds in our survey are too small to be resolved in low resolution OMEGA tracks and are not covered by high resolution tracks. Where possible we have used CRISM targeted images to investigate the composition of the mounds. For seven of the craters CRISM targeted images were not available in the summer season, or of insufficient quality (obscured by clouds or other atmospheric phenomena). Mounds that are completely covered in dunes had no discernible water ice signal during northern summer. Instead their spectra resembled those of background dust. For two mounds with partial dune cover (craters 663 and 769), a weak water ice signal was detected in high albedo patches, but this signal was not significantly stronger than that of the surrounding plains or crater interior. For crater 769 in particular the spectra across both available images (FRT0000b546 and FRT0000cf48) were dominated by dust, possibly due to the occurrence of small dust storms before these spectra were taken. These storms were observed in Mars Color Imager (MARCI) images (Malin, et al., 2008. MRO MARCI Weather Report for the week of 30 June 2008 – 6 July 2008. MSSS-40, http://www.msss.com/msss_images/2008/07/09/ and MRO MARCI Weather Report for the week of 13 October 2008 – 19 October 2008. MSSS-55, http://www.msss.com/msss_images/2008/10/22/). However, by inspecting the map of surface water ice abundance of Brown and Calvin (2010) derived from analysis of CRISM mapping data, we were able to confirm the presence of water ice in craters:332, 515, 577, 663, 697 and 795, whereas 436, 480 and 769 had no water ice signal.

Spectral instruments (such as OMEGA and CRISM) can only penetrate the optical surface and give information on the composition of the top few microns of the substrate in the case of rock or dust, and the top few centimeters in the case of ice (Brown et al., 2010). Hence the signals that we have observed could be from a thin surface veneer. However, other lines of evidence support the hypothesis that the mounds are composed of a substantial thickness of water ice, notably radar and stratigraphic arguments presented in Section 3.3 and also thermal arguments in Section 4.5.1. Having confirmed the presence of water ice in eleven of the crater mounds and having found striking morphometric similarities in all 18 mounds, including craters completely covered by dunes, we propose that all these mounds have a similar composition.

3.3 Mound structure and stratigraphy

Visible layering is common but not universal in mound deposits. It is picked out by reflectivity contrasts, probably due to differences in dust and/or frost concentration, or different resistance to erosion. Layers intersect the mound surface, forming contours on the gentle surface slopes, and outcrop geometries suggest that layering persists at depth within the mounds. We have mapped and measured spacing between layers on six crater mounds, three with partial dune cover and three without cover (Fig. 7). In addition, we have observed, (but did not measure) layers in mounds in craters 697, 795, 544 and 577 located in close proximity to the polar cap. Layers were not observed in craters 934, 904, 882 814, 811 and 480, where mounds are entirely covered by dunes. Crater 515 has no dunes, but no layers were observed, as in crater 436 which has partial dune cover. We have also found layering in a small mound in crater 332 (Fig. 2), which was excluded from our survey, because the mound does not have sufficient relief (> 150 m) to allow us to reliably distinguish it from a dunefield, a criteria set out in our mound identification in Section 2.1.

Interior layering of mounds is exposed in areas with erosion. These areas are systematically positioned on the south and east facing slopes of mounds (Fig. 7) and usually on the steeper than average slopes. These zones spatially overlap with significant proportions of water ice in Fig. 6, in particular zones with water ice band depth > 0.7 water ice in Korolev (206), Dokka (388) and crater 579. This strongly suggests water ice is the principal component in the subsurface. We have estimated the thickness of 1438 layers in the six crater mounds with extensive exposures. Where CTX images were used, HiRISE images within the same scene were used to confirm that layers thinner than those easily visible in the CTX images were not present. Layer thickness ranged from 0.007 to 12 m, with an average of 0.42 m and a standard deviation of 1.55 m (Fig. 8). The layering observed in HiRISE images of the mound in Louth crater (503) is much finer than in all other mounds (median 0.15 m). In all cases apart from the fine-scale layering in Louth crater (503) layers are laterally continuous on scales of 100-1000 m.

Dips and strikes of layers were measured at 35 locations within mounds with extensive exposures (Fig. 7). Measured dips range up to 20° with a mean of ~ 5°. This is consistent with the distribution of surface slopes at sites with layer exposures (Fig. 9), with a mean of 4.4° and a maximum of about 20°. The difference between layer dip angles and surface slope at any given location ranges from 0 to 9°, with a mean of 2° and little skewing. This implies that the dip of the layers could be explained by draping of a topography similar to that of the present day outcrop sites. Figure 10 shows a detailed example of the interaction of layers with topography, highlighting that the dip directions are robust, even if the measured angles are subject to uncertainty (see Section 2.3 for a detailed discussion). The patterns of layer orientation are similar in all the mounds: strike orientations are similar to those of the present surface with dip directions away from the top of the mound and in some cases sub-parallel to crater-rim interior slopes. In all mounds with extensive exposures, layering is disrupted by unconformities. As an example, Fig. 10 shows unconformities observed within the mound in crater 579. Younging of these deposits towards the top of the mound can be inferred from the truncation of layers that are located stratigraphically below truncating layers (e.g., layer packet A (below) is truncated by layer packet B (above) in Fig.

10). Unconformities were also observed in Louth crater (503) with the same apparent trend. Although the younging directions are similar for coarse and fine-scale layering in this crater, exposures of the two layer types are separate. Hence, we cannot tell if the fine layered deposits are located stratigraphically beneath (i.e. older), or above (i.e. younger) the coarse layered deposits. The fine-scale layers in Louth (503) pinch out laterally on the scale of meters and have frequent discontinuities, which we interpret as evidence of small-scale topographic drape. Recent wind erosion of these layer outcrops has caused the formation of sharp, irregular groves or sastrugi (previously noted by Brown et al., 2008), draping of which would give rise to the observed stratigraphy.

SHARAD radar data shows that the layered structure of mound deposits extends deeper into the subsurface. We have found internal structures in the mounds of Korolev (206) and Dokka (388) craters. Verified with clutter simulations, radargrams of the interiors of these craters (e.g., Fig. 11) indicate that the main stratigraphy follows the contours of the present-day mounds. Layer thinning is observed on the gently dipping south facing slope in Korolev (Fig. 11). Layers deeper in the mound interior have a lower dip for both craters, but are more undulating in the case of Korolev. The following vertical measurements are directly dependent on the assumed permittivity, as we have chosen pure water ice they are maximum estimates (lower water ice contents would give reduced lengths). The minimum spacing between reflectors in the radargram for Korolev is estimated to be 6-18 m with approximately ~ 20 distinct reflectors visible over a thickness of ~ 2.5 km. The reflectors picked up by radar instruments, such as SHARAD are generally acknowledged to delineate packets of layers (Phillips et al., 2008; Putzig et al., 2009), rather than individual beds. The last return corresponds to ~ 3.9 km below the rim for Korolev, which compares favorably to the estimated depth of ~ 2.9 km from the relationships of Garvin et al. (2000) and 3.1 km from our results, implying that the permittivity of the mound material does not differ much from that of water ice, as was assumed in the depth correction. Thus, the SHARAD data confirms that layering continues throughout the mound, and that the layered material is likely to consist mainly of water ice over the entire depth of fill.

Other, smaller crater mounds did not produce internal radar reflections, probably because they do not contain sufficiently continuous layering or sufficiently strong dielectric contrasts. In the along and across track directions reflectors must be at least 300-1000 m and 1500-8000 m long, respectively, to produce a strong radar return (Alberti et al., 2007). The mound in crater 579 is likely to be sufficiently large to allow radar detection of internal structures, but it had no SHARAD coverage, at the time of writing.

Given that layering appears to be pervasive within the mound deposits, we can surmise that mound patches without exposed layering in Fig. 7 represent areas with present-day deposition of water ice.

4. The origin and evolution of the crater-mounds

Eleven, possibly eighteen intermediate and large impact craters at high northern latitudes on Mars have interior mounds consisting mainly of water ice. These ice-filled craters are some of the deepest depressions in their surroundings. The mounds have a smooth, convex-up shape, often asymmetrically placed within their crater and inclined to the south, surrounded by a distinct moat. They have internal layering with unconformities indicating upward younging of deposits and an alternation of phases of deposition and erosion. Layers appear to drape over the extant topography of the mounds. This layering may be pervasive down to the base of the deposits where they rest on the crater floor. Any explanation of the origin and evolution of the crater mounds must address these key features. In this section we consider four distinct scenarios of mound formation after a brief exploration of controls on the present day geometry of the mounds.

4.1 Present-day form

The surface form and position of the mounds inside the craters may be determined by the mechanism that drives mound formation, or it could be due to subsequent processes changing the initial shape of the deposit. Only in the former case does the mound geometry contain information about the formative mechanism. Therefore we consider briefly whether post-depositional erosion alone can account for the mound shape. Using a radiation balance approach, and including effects of shadowing caused by the crater walls, atmospheric scattering, temperature-dependent re-radiation from the surface and conduction, Russell et al. (2004) have demonstrated that a north-south asymmetry and an exterior moat can be developed from an initially flat ice body inside a crater over limited time under current orbital configuration. They showed that any east-west asymmetry cannot be produced from radiation alone, but could be explained by aeolian processes, as winds coming off the cap are deflected longitudinally by planetary vorticity. Figure 12 shows the asymmetrical placement of the mounds within our studied craters. Mounds proximal to the cap have a westward displacement from the crater center and this matches the prevailing easterly wind direction as indicated by dune slip faces (Tanaka and Hayward, 2008). Mounds closer to the cap tend to be more asymmetric than those further away. These observations suggest that these mounds are primarily shaped by the wind, which is known to be stronger near the cap (e.g., Howard, 2000). Further away from the cap the mounds are placed obliquely or orthogonal to the prevailing wind, suggesting a stronger influence of radiation. The orientation of mound patches with exposures of internal layering in six of the craters (Fig. 7) matches this explanation. Dominance of south-facing layer exposures can be attributed to directionality of incident radiation, and east- or west-facing exposures can be related to regional winds. Together the asymmetry of the mounds and the orientation of layer exposures suggest that radiation-driven ablation and wind work together (with different relative magnitudes) to shape the existing ice deposits. This likely obviates the need to invoke a primary formation mechanism to explain the present-day large-scale morphology of crater mounds.

4.2 Remnant of a more extensive polar cap

As a first possible scenario of mound formation it has been proposed that the interior mound of Korolev and similar features elsewhere could be erosional remnants of a formerly more extensive polar cap (Fishbaugh and Head, 2000; Garvin et al., 2000; Tanaka et al., 2008), along with outliers such as Olympia Mensae. Specifically, Tanaka et al. (2008) have suggested that the Korolev deposit and those similar to it could be related to stratigraphic unit "ABb1", the Polar Layered Deposits (PLD). A previously larger extent of the cap (including the PLD and the Basal Unit beneath) is surmised based on the large erosional scarps at the current margins of the cap, however the precise spatial extent is not known. With reference to the position of outliers, Fishbaugh and Head (2000) and Johnson et al. (2000) have proposed that the cap extended symmetrically down to 80-75°N. Rodríguez et al. (2010) have argued that an extension of the PLD down to ~ 60°N and subsequent retreat could account for the distribution of Late Amazonian pedestal craters. If the polar cap was more extensive, then there is general agreement that it must have occurred under higher obliquity conditions than at present (which favors migration of ice to lower latitudes), but the exact timing is debated.

The PLD was laid down in the Amazonian as sub-horizontal layers from regular (possibly seasonal) cycles of atmospheric deposition of water-ice mixed with small and variable amounts of dust, producing regularly spaced layers with differing albedo of meter to tens of meters in thickness. Layers are extremely laterally continuous (e.g., Fishbaugh and Hvidberg, 2006) and angular unconformities occur mainly in association with the spiral troughs (Putzig et al., 2009; Smith and Holt, 2010; Tanaka et al., 2008). These layers are observed to generally dip at less than 1° from image (Milkovich et al., 2008) and SHARAD radar data. Even layers draping trough walls do not have dips in excess of 1° (e.g. Fig. 3e of

Smith and Holt, 2010). The maximum total deposit thickness is estimated to be 1800 m from MARSIS radar data (Selvans et al., 2010). These deposits are thought to provide a record of past climate, similar to the polar ice caps on Earth (e.g., Cutts 1973; Laskar et al., 2002; Milkovich and Head, 2005; Milkovich et al., 2008). Estimates of past accumulation rates are on the order of 0.5 mm/yr (e.g., Laskar et al., 2002), or lower (Perron and Huybers, 2009). The modern polar cap is undergoing both accumulation, as shown by the burial of craters in fine-grained ice (Banks et al., 2010) and the lack of dust accumulation on its surface, and erosion, as shown by the exposure of coarser grained ice with lower albedo at scarps (Langevin et al., 2005).

Underlying a large part of the PLD is the Basal Unit (BU) or Planum Boreum Cavi unit (ABbc) using the nomenclature of Tanaka et al. (2008). This unit is characterized by a lower albedo compared to the PLD and is inferred to be rich in sandy material containing a variable quantity of volatiles. Layers have variable regularity and cross-bedding and a thickness up to decameters (Herkenhoff et al., 2007). Layers can be laterally discontinuous particularly on the 100 m scale with frequent angular unconformities. Lighter-toned layers are more resistant and thought to be more ice-rich. The BU is hypothesized to be of Middle to Late Amazonian age, originating from aeolian deposition, sometimes in the form of ice-cemented dunes. The dip angle of these layers has not been measured, but cross-bedded layers are locally steeper than strata of the PLD. The maximum deposit thickness is estimated to be 1100 m from MARSIS data (Selvans et al., 2010). It is possible that the BU, like the PLD has been more extensive in the past.

If the mounds in our study are remnants of a larger, continuous polar deposit then they might be expected to have a similar stratigraphy. Measured layer thicknesses on the mounds are generally within the range reported for the PLD (Milkovich et al., 2008). According to our own assessments of ~ 350 layers over three outcrops (see Section 2.3 for details) the PLD have layer spacing of 0.5 to 40 m with a median of 4.7 m and BU between 0.05 and 7.0 m with a median of 1.4 m, presenting no statistically significant difference with layering in crater mound deposits (Fig. 8). The only exception is the fine-scale layering found in the central part of the mound in Louth crater (503), which is significantly and consistently finer than layers measured anywhere else. However, the frequency of discontinuities (e.g. Fig 10) in the mounds in general seems to be greater than within the PLD (where they are rare).This implies that the mounds have experienced more erosive episodes. In addition, mound layering is laterally less continuous, but this could be due to limited exposure compared to the polar cap. Layer dip angles in the mounds are up to 20°, which is much greater than the ~1° dips measured by Milkovich et al. (2008) for the PLD. Moreover, it would be expected in the case of PLD-like deposition that layers should drape the crater walls, dipping gently towards the center of the crater, with a flat lying central part as observed for craters emerging from the polar cap (see Fig. 4 of Rodríguez et al., 2010). This contrasts with the observed outward dip of mound layers that extends into the mound interior, indicating growth from a central core (Figs. 7, 11).

Preservation of the mounds in crater interiors during cap-retreat has been attributed to the shielding effect of the crater geometry (Garvin et al., 2000). For a given crater depth/diameter ratio, then, the ice preservation rate should depend to a degree on the radiation regime which changes with latitude. Confirmation of this link comes from the south polar region, where the volume of ice bodies inside craters has been observed to decrease away from the pole (Russell et al., 2003). However, there exists no such relationship either between the mound volume as a proportion of the total cavity volume (a least squares linear fit gives R^2 of 0.302 and p-value of 0.011), or the absolute mound volume and distance from the polar cap in the northern hemisphere (R^2 of 0.058 and p-value of 0.806). This could indicate a) a different origin for the mounds in the south compared to the north, or b) different environmental factors determining mound survival between the two hemispheres (e.g., geometric shielding is less important than, for example, wind patterns in the north).

On balance we feel that stratigraphic, structural and volumetric evidence are not compatible with the theory that crater mounds distal from the current polar cap formed as part of a larger contiguous PLD or BU deposit. In particular this applies to Louth (503), Dokka (388), Korolev (206), craters 579, 663 769 and 436, but we cannot rule out the role of a slightly more extensive polar cap in the formation of mounds located proximal to the present-day polar cap deposits.

4.3 Impact-driven water release

Our second scenario involves the mobilization of water from permafrost. It has been hypothesized that relatively long-lived hydrothermal systems could be driven by the heat released by a meteorite impact (Abramov and Kring, 2005; Osinski et al., 2005; Rathbun and Squyres, 2002), or that the seismic energy of an impact can mobilize water/permafrost (Harrison et al., 2010). This mechanism was suggested for the formation of a channeled scabland extending from the ejecta blanket of Lyot crater located at 55°N, 330°W and a band of paleo crater lakes (El Maarry, et al., 2010), both south of our study area. The northern plains region of Mars should be particularly susceptible to the mobilization of ground ice as it is abundant in the shallow sub-surface. The possibility of impact-induced hydrothermal systems in permafrost-rich ground on Mars was first suggested by Brakenridge et al. (1985). In this process the impact melt sheet of larger impact craters contains sufficient heat energy to melt the surrounding permafrost and water contained in topographic highs drains into the crater hollow. The resulting lake freezes at the surface, retarding evaporative loss. Additional water can be sourced from subsequent hydrothermal circulation, driven by temperature gradients within the surrounding regolith, which draws water into the crater interior. Once the system has cooled sufficiently water starts to drain downwards out of the crater lake and eventually a frost front may propagate upwards from the subsurface, leaving a solid body of ice in the crater interior.

Rathbun and Squyres (2002) have shown that impacts creating craters with a diameter less than 7 km have insufficient energy to sustain a crater lake. This cut off is relatively close to the observed lower diameter limit for craters with mounds of ~ 9.5 km. For larger craters, the presence or absence of a mound should then depend on the time since impact, and we would expect to find mounds preferentially in the youngest craters. At received cratering rates (e.g., Hartmann and Neukum, 2000), over our study area (excluding the area of the cap), an impact such as Dokka (51 km diameter) could be as old as several hundreds of millions of years and Korolev (80 km) over 1 Ga. It is unlikely that relatively small ice bodies would have survived for this length of time, especially given that the polar cap is believed to be no older than 4-5 Ma (e.g. Levrard et al., 2007). There are not sufficient craters in this size range without a mound to confirm, or refute the impact heat theory, although we note that Lyot crater, a large, 215 km diameter crater just south of our study area, has no evidence of water ponding in its interior either as a mound, or a desiccated lake (Harrison et al., 2010; Russell and Head, 2002).

In this scenario mound size should have no particular spatial trend, assuming that the same amount of sub-surface water is available throughout the polar region, but depend primarily on the energy of impact. As crater diameter is a proxy for impact energy, we might expect it to scale with the mound volume (an approximation of the amount of water released). Although the percent of the crater cavity in-filled by the mound does not correlate well with the diameter of the crater, the absolute volume of infill increases with the crater diameter, apparently as a power law with the form (0.04±0.09) $D^{2.8\pm0.2}$ with an R^2 of 0.983 and a p-value of 9.6×10^{-11} where the uncertainty attached to the estimation of mound volume has not been taken into account. However, the hydrothermal system is unlikely to have sustained a hydraulic head permitting significant flow above the plains surrounding impact craters. In fact, Rathbun and Squyres (2002) estimated a lake depth of 300 m for a 7 km diameter crater and of ~ 1 km for a 180 km diameter crater (which is almost twice the diameter of Korolev 206). Several craters in our survey have mounds that significantly exceed these values. Additional elevation of the mounds (of up to 0.5-0.8 km in some cases) would have to be accounted for, for example by invoking frost-heaving processes active in the waning period of hydrothermal activity, similar to pingo formation on Earth, as suggested by Sakimoto et al. (2005) and Bacastow and Sakimoto (2006).

It is perhaps most difficult to reconcile our observations of mound deposit stratigraphy with the requirements of the scenario of impact-driven water release. Creation of the mounds by this mechanism implies that the mounds have geologically the same age as the impact that created their host crater, and are formed in a single event. Moreover, the mechanism could give rise to complex, out-of-sequence layering as periods of hydrothermal injection into an ice covered lake should cause under-plating and thermal erosion, similar to that of aufeis (Gillespie et al., 2005). This could be followed by freezing of the water from both the top and the bottom, producing frost-heave features, folding and faulting when the hydrothermal system wanes. Water circulated through the subsurface at high temperatures may also contain impurities consisting of salt and sediment that could be frozen into the deposits. These expected characteristics contrast with our observation of regular, upward younging layers of relatively pure ice, without major faults, or folds. In Louth crater (503) small-scale layering has more folds, but is still relatively regular. Finally, the pervasive presence of large-scale discontinuities in the mound deposits suggests an evolution involving multiple events. This, and lack of correspondence with proposed crater lake dimensions are all problems for this mechanism.

4.4 Artesian flow

The third scenario of mound formation appeals to the presence of deep-seated faults below impact craters (e.g., Christeson et al., 2001). These faults could tap into a deep aquifer confined by the cryosphere that is hypothesized to underlie the north polar plains (as predicted by Clifford, 1993; Clifford and Parker, 2001; Clifford et al., 2010), providing a conduit to the surface for the confined water. Water would come to the surface at times when the cryosphere enlarges and impinges on the aquifer producing sufficient confining pressure. This mechanism has been proposed for the creation of crater lakes at lower latitudes by Newsom et al. (1996). Upwellings would not necessarily occur at the same time as the impact event and could occur many times, producing a layered stratigraphy as observed. Assuming that the top of the aquifer is located on a single level, then mounds should be found in large craters with large fault systems, but also smaller craters located on low-lying topography. Our results show this trend, where mounds are found in craters which tend to penetrate to the deepest local elevation (Fig. 4). If the aquifer is fully connected and the mounds formed at the same time then the mounds should extend up to a similar topographic elevation, equal to the hydraulic head. This is not the case. If flow rates are low, then it might be expected that smaller mounds should be found in craters with a base located at higher absolute elevation. We do observe a very weak negative relationship between the inferred elevation of the crater base and both relative (linear) and absolute mound volume (exponential), with R^2 values of 0.006 and 0.37 respectively and p-values of 0.31 and 0.003 respectively.

As in the hydrothermal hypothesis mound formation would involve processes akin to those occurring in pingos. This, together with wind and/or radiation ablation could give rise to doming of the mounds. The internal stratigraphy of such mounds should be complex. Water could pond on the surface and gradually freeze as a lake system, or could inject underneath pre-existing ice. Subsequent pulses would deform, fracture and fold, possibly under-plating the roof, or breaking through to the surface, filling topographic hollows. Potentially the age of the ice could increase, or decrease towards the center, depending on the dominant emplacement mechanism. Ice interior to the mounds should have many discontinuities, folds and faults. Such complexity is not characteristic of the mounds in our study, which display regular layering, large-scale discontinuities, minor folding and no faulting. The fine-scale layering in Louth crater (503) is the only example that shows sufficient complexity to fit with this model, but it is the exception, rather than the rule. Interestingly, the mound in that crater is placed very eccentrically, as would be expected from the peripheral position of impact fault systems. However, neither there nor elsewhere have we found spectroscopic evidence for the presence of salts in mound deposits, which is expected for upwelling water sourced at depth. In addition, there appears to be a discrepancy between the observed distribution of crater mounds and the geometry of the putative deep aquifer (Clifford, 1993). The aquifer is thought to be located deepest close to the polar cap, where surface temperatures are lowest, but there we have found mounds in several craters of limited diameter and depth. It is possible that the aquifer is located at shallower depths than proposed by Clifford (1993) very near to the pole, due to water being supplied from the base of the polar cap (e.g., Longhi, 2006). Away from the polar cap, the cryosphere is thought to thin southward, making mound or lake formation more likely in this scenario. The lack of evidence for hydrological activity in Lyot crater (Russell and Head, 2002; Harrison et al., 2010) is at odds with this expectation. Despite the mounds occurring in the deeper penetrating circumpolar craters, inconsistencies in stratigraphy, composition, position and distribution argue against artesian flow from a deep aquifer as a formative mechanism for these mounds.

4.5 Crater microclimate

Our fourth hypothesis calls on microclimate phenomena associated with crater topography for growth of ice mounds from atmospheric deposition. We start our exploration of this mechanism with some observations of temperature and albedo in and around craters in the north polar region.

4.5.1 Craters as cold traps

The temperature and albedo evolution of ice mounds in Korolev and Dokka through the Martian year have been reported by Armstrong et al. (2005) and Kuti (2009). These mounds are consistently ~ 10 K cooler than ice-free terrain at the same latitude for most of the year, with an accentuated daytime difference of ~20-40 K in summer months (Fig. 13). On these crater mounds, surface temperatures remains close to the water frost point throughout the year (particularly at night). In other craters without ice (an example is given in Fig. 13), temperatures inside and outside the craters are very similar for most of the year. These observations suggest that the presence of ice modulates local near surface temperatures in crater interiors. A minimum thickness of ice of several meters at the surface may be required to produce this effect (Armstrong et al., 2005). So how did this initial ice layer come about in the craters with mounds, and how and when does/did this volatile deposition occur?

The following thermal, spectral and image data reveal that crater cavities may enhance the deposition and/or preservation of volatiles at the present day. Viking Infrared Thermal Mapper (Kieffer 1977) and Thermal Emission Spectrometer (TES; Titus et al. 2001) observations have shown temperatures below the CO_2 sublimation point inside craters during winter. These cold spots are unlikely to represent real surface temperatures, but could be caused by cloud formation and thus possibly increased deposition of CO_2 ice (Ivanov and Muhleman, 1999). In agreement with these observations, we also observe that apparent crater floor temperatures with and without ice mounds are up to 10°K below those outside the crater between L_s = 240-320° (Fig. 13).

We have examined published OMEGA data (Appéré et al., 2011) and HiRISE and CTX images for presence of surface ice at latitudes greater than 60°N around the L_s = 60° crocus date, when CO_2 frost is receding. The OMEGA data show that crater floors remain covered in water ice for approximately 5° of L_s after the main seasonal cap has retreated to higher latitudes (Figures 5-8 Appéré et al., 2011). These deposits then degrade to an annulus around the crater rim, before disappearing. HiRISE and CTX images show that crater floors rarely have any remnant high albedo ice deposits after the crocus date. They do show, however, that high albedo patches remain after the crocus date on north-facing crater slopes (Fig. 14B) and sometimes as E-W ‘plumes’ outside the crater (Fig. 14A). Many of these patches disappear in the summer, but some remain as shown in CRISM analyses performed by Seelos et al. (2008). These authors have hypothesized that the ‘plumes’ are a result of wind-driven, orographic deposition of water ice. Hajigholi et al. (2010) monitored seven craters at > 55°N and also noted crater interiors were favored sites for ice deposition. These observations show that crater floors are not presently sequestering H_2O or CO_2 deposited in the seasonal polar cap under present day conditions.

If atmospheric deposition is the driving process in which layers are constructed, then to start mound formation, ice build-up in winter months on the crater floor must outstrip sublimation in spring and summer. The presence of high-albedo ice deposits throughout the year may then locally suppress near surface temperatures and form a long-lived cold trap for H_2O. It is possible that the ice layer needed to trigger further build-up is provided by a hydrothermally induced crater lake, or the remnant of a more extensive cap deposit. Below, we examine a scenario that does not involve an external start, and is instead entirely climate-driven.

4.5.2 Reduced sublimation of water ice in craters

We have demonstrated that at least seasonally all craters trap water ice. However, this does not explain why only some craters in the near polar area host mounds and why others do not. We suggest that this is because of differing preservation potentials between the craters, which are determined by differing rates of sublimation. Factors, such as shadowing within the crater, differences in surface thermophysical properties, atmospheric temperature, atmospheric pressure and the wind regime, play a role in defining the sublimation rate.

The higher the surface temperature the more rapid the sublimation, so cooler temperatures favor the preservation of ice mounds. The surface temperature is controlled by insolation (and thus shading) and surface themophysical properties. If we consider shadowing alone, water ice should be preserved on all north-facing crater walls rather than on crater floors and mounds should therefore have an off-center nucleus, positioned on the north-facing slope. In the craters where we have studied the stratigraphy the nucleus of the mound is located on the crater floor and not all craters have mounds, hence this factor alone does not explain our observations. We cannot determine the surface material albedo and thermal inertia of the craters prior to mound deposition; hence we cannot definitively say what role these thermophysical properties played.

The atmospheric pressure at the base of a crater is greater than on the surrounding plains (due to altitude) and this can act to reduce sublimation by free convection (e.g., Ingersoll, 1970). As the mounds are often found in craters that penetrate deeply into the crust (Fig. 4), this may be one factor that could help to explain the mound distribution. However, other factors must be playing a role, because not all deeply penetrating craters have mounds. Wind speed and its saturation level also play a key role in sublimation, with strong, dry winds enhancing forced sublimation (e.g., Ingersoll, 1970). Given this, it might be expected that craters with mounds should be located away from the strong summer katabatic winds, but we observe no systematic placement of craters with mounds with respect to these winds (see maps in Massé et al. 2012, Ewing et al., 2010 and Howard, 2000). At a smaller scale craters generate their own wind systems. Fenton and Hayward (2010) noted a peculiar “bullseye” dunefield morphology in craters close to the southern polar cap, and attributed it to inward flowing local katabatic winds. We see the same morphology inside many of the craters with mounds and inside other craters in the area, indicating the strong influence of local winds. The topography of each crater would have different effects on these wind patterns, but with the data available we find no systematic pattern to explain why some craters have mounds and others do not. To precisely determine the role of local-scale winds in triggering mound formation would require a full meso-scale climate model (e.g. Spiga, et al. 2011), which is outside the scope of this paper.

4.5.3 Initiation and mound building

We envisage the following sequence of events (Fig. 15) for the initiation and evolution of mound building:

1) In winter condensation of water ice is followed by the condensation of CO_2 intermixed with a smaller fraction of water ice. The micoenvironment inside the crater interior causes additional condensation of volatiles over the winter period, resulting in a thicker layer of both CO_2 and H_2O in the crater floor than on surrounding plains.

2) The CO_2 layer is first to sublimate in spring. This process is driven by direct insolation, but also draws heat from the surrounding atmosphere, and this maintains a low temperature in the crater interiors (~150 K). Simultaneously, water ice is sublimating at lower latitudes and this water vapor can condense back on to the sublimating CO_2 deposits at higher latitudes (e.g. Appéré et al. 2011). The thicker layer of CO_2 deposited on the crater floor, and lower local temperatures cause the crater floor to defrost later than the surrounding landscape. This means that the process of re-condensation of water vapor can continue for longer and in particular while H_2O is sublimating from the landscape immediately surrounding the crater.

3) Once the surrounding landscape has completely defrosted and the CO_2 sublimation in the crater has finished, there remains a layer of water ice condensed over winter and additional, re-condensed water ice layer.

4) The sublimation of this remaining water ice is retarded by the microclimate of the crater interior (likely pressure and wind effects). Due to the high albedo of the deposit both free and forced sublimation will also be slowed due to the lower atmospheric temperature above the deposit (further enhanced if the deposit is thick). A thick layer (with high thermal inertia) might permit re-condensation during the night if any sublimation has occurred.

5) Once a “thick” layer has been built (~1-4 m required according to Armstrong et al., 2005) this creates a positive feedback, whereby the high albedo and high thermal inertia form a cold trap which produces net accumulation over the course of a year (Fig. 13). Then, water vapor delivered to the crater at almost any time of year could freeze onto the existing deposit, and progressive build-up of an ice mound would ensue.

Atmospheric deposition would give rise to upward younging stratigraphy, with regular layering, reflecting cyclical deposition on seasonal, but more likely much longer timescales.

Changes in the balance of deposition and ablation could cause alternation of aggradation and decay of the deposits, recorded in (angular) unconformities within the mound stratigraphy. During an initial build up phase, deposits are expected to line the crater interior, but solar irradiation and directional ablation by winds would eventually cause the formation of a moat and give the deposit a characteristic mound shape. Further volatile deposition would drape the mound topography, dipping outward from an accumulation center, and following any erosional detail in the pre-existing ice surface. All of these expectations are matched by our stratigraphic observations.

4.5.4 Timing

At present, craters are not initiation points for mound building, nor is mound building very rapid. Assuming that current maximum deposition rates are similar to those on the north polar cap, 3-4 mm/yr (Banks et al., 2010), a mound such as the one in Korolev would take ~ 500 ka to form when deposition is uninterrupted. Orbital forcing of climate is important on this time scale. Using a global climate model and different orbital parameters Levrard et al. (2007) showed that obliquity (axial tilt, ~124 ka cycle) is the primary orbital control on ice accumulation in the polar region. Low obliquity favors polar cap accumulation and obliquities greater than 30° can destabilize ice in the whole polar region (which last occurred at ~0.4 Ma). Another important feature is the precession cycle (~51 ka), a combined effect of the eccentricity and longitude of the perihelion. The orbit of Mars is elliptical, so that northern hemisphere spring is the longest season, followed by northern summer, winter and the shortest season is autumn, the present-day location of the perihelion. Thus, northern winters are comparatively short and warm, and summers are long and cool. If the perihelion is located in northern summer, then winters become longer cooler and the summer becomes short and relatively hot. It is the length of the seasons, rather than their relative temperature, that governs ice accumulation at obliquities less than 25° (Levrard et al., 2004; Levrard et al., 2007). As the obliquity has been relatively stable over the last 500 ka, precession cycles dominate. We suggest that mound triggering and building may be favored when the perihelion is located in summer or autumn, allowing a longer period of winter deposition and a long spring period when volatiles are delivered from the receding seasonal cap and the main polar deposits. This situation last occurred between 10-25 ka before present (Montmessin et al., 2007). These cycles could account for the large angular unconformities that we observe. Further modeling is required to accurately ascertain orbital parameters that are favorable to mound building.

If the mounds are indeed formed by atmospheric deposition, then their overall form and internal stratigraphy are controlled by climatic conditions. If their timescale of adjustment is relatively quick compared to the timescale of climate change then they should provide a stratigraphic record of recent climate changes, driven by changes in Mars' orbital configuration.

4.5.5 Other ice outliers

We surveyed the outlying high albedo deposits that are not located within craters to ascertain if there was a common formation mechanism with the mounds inside craters. The majority of these deposits are mostly thin (they show underlying impact craters of ~ 500 m diameter with an expected rim height of ~ 25 m) and only around Olympia Mensae (between ~95-135°E and ~73-80°N, Fig. 2, marked "OM") do they reach thicknesses of > 100 m. The presence of relatively thin outliers located between Dokka, Korolev and the Olympia Undae dunefield (between ~150-240°E and ~75-80°N latitude) has been previously attributed to the presence of polar deposits underlying this area (Fishbaugh and Head, 2000). Tanaka et al. (2008) classified these high albedo deposits as 'ABb$_4$' or Planum Boreum 4 unit, which is the unit that caps the present polar cap, and suggested that these deposits are recent (last 21.5 ka) and relatively thin. Recent radar data has confirmed that the BU extends beneath the Olympia Undae dunefield (Selvans et al., 2010), making it quite plausible that thin deposits extend further north. These high albedo deposits occur where dunes are not present, so it could be that they are accumulations resulting from the exposure and resulting cold-trap effect of BU deposits.

The presence of thicker deposits in the Olympia Mensae area is more difficult to explain. They occur as discrete patches with steep boundary scarps with up to ~20° slopes and thickness of up to 350 m, but more generally around 100 m. They are located latitudinally between and to the east of Louth (503) and crater 579. This area is of relatively high elevation compared to the same latitude elsewhere around the pole. Three of the outcrops are elongate in the E-W direction and have layer exposures in all orientations, consistent with wind erosion rather than radiation as the primary shaping mechanism. The Olympia Mensae deposits have visually very similar stratigraphy to the crater mounds in terms of layer spacing and presence of low-angle unconformities, but only have very low dips (<1°) and no draping layers (Fig. 16). They have almost 100 % exposure of the layers, which implies they are currently undergoing retreat. This is supported by Brown et al. (2011), who suggest that sublimation in the Olympae Mensae region might be the main source of the water which masks the CO_2 seasonal cap during its spring regression. This suggests these outliers unlike the crater mounds have not undergone significant recent deposition and therefore their evolution is somewhat different to that of the ice mounds inside craters. The questions of how their deposition was triggered and their previous extent are left open as targets for future investigation.

5. Conclusions

We have identified and studied 18 inner-crater mounds in Mars' north polar region that have an origin different from the central peaks normally found in complex craters. They are restricted to craters with a diameter >9.5 km located at latitudes > 70°N. Two of these crater mounds consist largely of water ice according to radar data and nine others have substantial water ice at the surface as observed in spectral data. We infer by similarity of form and location that the other mounds are also composed of water ice. Their large-scale morphology, flat topped with a circumferential moat, can be explained by ablation of volatiles by solar radiation and wind, leaving no indication of the formation mechanism.

The formation of these mounds by either impact-induced hydrothermal circulation or artesian upwelling is inconsistent with the generally unfolded and unfaulted, regular layering with upward younging stratigraphy exposed on mound surfaces. Moreover, ice layers drape the existing topography on at least six of the mounds, implying growth from a central core. This is inconsistent with these mounds being erosional remnants of a previously more extensive polar cap. It is plausible that this form of growth is shared by eight other mounds that are not directly adjacent to the present polar cap. For the four mounds abutting the present polar cap we cannot rule out an origin by cap retreat.

Ice accumulations can be maintained or grow by cold trapping when they have attained a minimum thickness required to form a negative thermal anomaly that persists throughout the year. This minimum condition can be achieved in one of three ways: a frozen paleo-lake from an impact hydrothermal system, a remnant of a more extensive polar cap, or atmospheric deposition driven by microclimate processes inside the crater. We have no observational evidence to support any of these initiation mechanisms, but we have explored in more detail the possibility that mound formation is entirely due to the microclimatic characteristics of deep impact craters at high northern latitudes.

Mounds are found in deeply penetrating craters where atmospheric pressures are high compared to the areas surrounding the crater. This suppresses ice sublimation, but not all deeply penetrating craters have mounds, so other climatic factors, such as wind regime, must be playing a role. We suggest that microclimatic effects act to suppress the sublimation of seasonal ice in springtime, both increasing its lifetime and creating a cold-trap onto which volatiles released from the surrounding landscape during warm seasons can condense. It is likely that during periods when Mars' orbital perihelion was located in northern summer, longer winters have allowed enhanced deposition of volatiles, while the potential for sublimation was limited by the brevity of summers. Under these conditions, which last occurred 10-25 ka ago, enough ice may have built up in deep crater interiors to start a positive feedback permitting mound building due to the formation of long lived cold trap above bodies with high albedo and thermal inertia. The main supply of water vapor to these cold traps may be the seasonal cap with secondary input from katabatic winds flowing from the polar cap during spring and early summer.

Where mound surfaces are not clad by dunes, the fact that layers are seen to be exposed at the surface indicates recent or ongoing erosion in these areas. However, elsewhere the lack of visible layering can be attributed to recent deposition of ice draping the mound topography. Hence, the mounds are currently undergoing change, likely driven by changes in the local climate. Changes in climate may shift and resize the depositional and erosion areas, explaining the complex pattern of discontinuities visible in exposed mound stratigraphy. If this interpretation is correct, then the crater mounds may be sensitive recorders of climate change in the north polar area, located in closest proximity to the main source of volatiles on present day Mars. Further investigation of their stratigraphy and dynamics may yield new insights into the past and present H_2O and CO_2 cycles of the planet.

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Figure Captions

Figure 1: Depth-diameter plot of all the craters north of 65°N, with ice-filled craters and dune-filled craters highlighted. Lines extending above the mound-points are the predicted depths of the craters with the mounds removed. Depth-diameter relationship derived for polar craters by Garvin et al. (2000) and the general relationship for Martian craters (Garvin, 2005) are marked. An anomalous pixel in the MOLA data has resulted in a ~1.5 km crater with >1 km depth, we believe this pixel to be an error in the data; hence this point should be ignored.

Figure 2: Map of the north polar region of Mars made from MOLA topography with overlain hillshade. It includes the locations of the mounds studied and locations of the swaths used to produce Fig. 4. Outlined in grey are all the craters included in the survey. The “+” indicates the position of crater 332, which is mentioned in the text, but does not have a significant mound. “OM” indicates the position of Fig. 17, with the arrow showing the viewing direction. “Complete dune cover” means that the mound spans the whole crater floor and is covered in dunes and “dune-covered mounds” are those which are covered in dunes, but only occupy part of the crater floor (the rest of the crater floor is dune-free.)

Figure 3: Left, Crater 388 Dokka. Perspective view (with 3x vertical exaggeration) of a subset of HRSC nadir (RGB-color online) image H1177_0000 draped over MOLA topography with profiles A-A’ and B-B’ marked and shown beneath. Right: Crater 811. Perspective view (with 3x vertical exaggeration) of a subset of HRSC nadir (RGB-color online) image H3711_0000 draped over MOLA topography with profiles C-C’ and D-D’ marked. The 20 m scale vertical roughness on profiles C and D is a result of the overlying dunes. Arrow points to North in both figures. Profiles A, B, C and D were generated from

MOLA topography – note the exaggeration of the vertical scale. Credit for HRSC images ESA/DLR/FU Berlin (G. Neukum).

Figure 4: Histogram plots of the elevation variation with latitude within the swaths outlined in Fig. 2. Black (blue online) indicates craters with ice-mounds and the height to which those mounds reach, labels correspond to those in Fig. 2 and Table 1. Note that the vertical scale is exaggerated.

Figure 5: Bar chart of calculated mound volume for each crater in descending order. Note the logarithmic scale on the y-axis. The mean value (314.7 km^3) is indicated by the upper dashed line and the median value (17.01 km^3) by the lower dotted line.

Figure 6: Right optical image with scale and north arrow, and left optical image overlain by proportion of water ice detected (band depth at 1.5 μm).

(A) Korolev crater (206), optical images from MRO CTX P20_008831_2529_XN_72N195W (L_s 85.33°), P21_009042_2528_XI_72N197W (L_s 92.58°), P21_009332_2529_XN_72N194W (L_s 102.66°), P22_009477_2530_XN_73N193W (L_s 107.77°), P22_009754_2529_XN_72N195W (L_s 117.69°), and OMEGA track 0987_2 (L_s 105.9°).

(B) Dokka crater (388), optical images MRO CTX B01_010108_2573_XN_77N144W (L_s 130.74°) and B02_010385_2572_XN_77N146W (L_s 141.32°), OMEGA track 1221_1 (L_s 136.4°).

(C) Crater 544, optical images THEMIS V13697003 (L_s 144.508) and V12474005 (L_s 97.106°), OMEGA track 1257_0 (L_s 141.3°).

(D) Crater 579, optical image MRO CTX P02_001819_2570_XN_77N270W (L_s 151.16°), OMEGA track 1017_1 (L_s 109.5°).

(E) Louth crater (503), optical image HRSC h1343_0001 (L_s 153.5°), OMEGA track 1017_1 (L_s 109.5°). Credit for CTX images NASA/JPL/MSSS, for HiRISE images NASA/JPL/University of Arizona, for THEMIS images NASA/JPL/ASU, for HRSC images ESA/DLR/FU Berlin (G. Neukum), for OMEGA data ESA/OMEGA.

Figure 7: Maps of layer exposures and measured layer dips in Korolev (206), Dokka (338), crater 579, Louth (503), crater 769 and crater 663. Shaded areas represent areas where layers were observed to be cropping out, dotted lines are selected layers to illustrate the nature of the layering and the dip and strike labels are in units of degrees. Dotted box in panel 579 indicates the approximate location of Fig. 10. Images all MRO CTX: **206 Korolev**: P20_008831_2529_XN_72N195W, P21_009042_2528_XI_72N197W, P21_009332_2529_XN_72N194W, P22_009477_2530_XN_73N193W, P22_009754_2529_XN_72N195W, **388 Dokka**: B01_010108_2573_XN_77N144W and B01_010108_2573_XN_77N144W, **579**: P02_001819_2570_XN_77N270W, **503 Louth**: P01_001370_2503_XI_70N257W, **769**: B02_010407_2587_XN_78N028W, **663**: P20_008795_2592_XN_79N299W. Image credit NASA/JPL/MSSS.

Figure 8: Boxplots of the estimated layer spacing in mounds and polar deposits. The vertical bars are the median values, the boxes delimit the interquartile range, the whiskers the range and the dots are the outliers (data points outside 1.5 interquartile ranges from the box). In grey are measurements made outside craters. Shorthand labels are as follows: **206 Korolev** – layers measured in Korolev crater in CTX images P20_008831_2529_XN_72N195W and P21_009042_2528_XI_72N197W**, 388 Dokka –** layers measured in Dokka crater in CTX images B01_010108_2573_XN_77N144W and B02_010385_2572_XN_77N146W**, 503 Louth HR** - layers proximal to dunes in Louth crater measured in HiRISE PSP_001700_2505, **503 Louth CTX –** layers measured in CTX image P01_001370_2503_XI_70N257W at the edge of the mound in Louth, **579**: layers identified in

crater 579 in HiRISE image PSP_008926_2575, **663** - layers measured in crater 663 corrected for layer dip as well as exposure slope, in HiRISE image ESP_016087_2595, **769** – layers measured in crater 769 in HiRISE image PSP_008416_2585, **PLD1** – layers measured in the PLD in HiRISE image PSP_010366_2590, PLD2 - layers measured in the polar layer deposits in HiRISE image TRA_000825_2665 and **PLD3** - layers measured in the polar layer deposits in HiRISE image TRA_000863_2640, upper quartile is at 15.0 m and the upper whisker at 29.3 m, with one outlier at 39.8 m. **BU** – layers measured in the polar basal deposits in HiRISE image TRA_000863_2640.

Figure 9: Histogram of mound slopes on layer exposure sites (as shown in Fig.7) as derived from 128 pixel/degree MOLA data. The mean of the distribution is 4.4° with a standard deviation of 3.6° and skew of 1.29.

Figure 10: Detailed sketch (left) cross section (top right) and detail-inset (bottom-right) of the layer stratigraphy in crater 579. On the sketch dotted lines are selected layers to illustrate the nature of the layering, bold lines are the locations of unconformities and dip and strike symbol labels are in units of degrees. The line labeled I-II indicates the position of the cross section adjacent to the image. Colored lines are contours in 25 m intervals derived from gridded MOLA 128 pixel/degree data (see online for color version). The crater wall is located to the right in both the sketch and the cross-section. Overlain on the sketch is the CTX image P02_001819_2570_XN_77N270W, credit NASA/JPL/MSSS. Labels A, B and C are packets of layers separated by unconformities. On the cross-section dotted lines show the estimated position of the unconformities and the relative vertical positions of the three layer packets A, B and C. From layer cross-cutting relationships we know that A is older than B, which is older than C; hence the packets were laid down in an upwards younging sequence. The location of the inset is indicated by the grey box. The image in the inset is HiRISE image PSP_008926_2575; credit NASA/JPL/University of Arizona.

Figure 11: Depth-corrected SHARAD radargram r_0554201_001 and clutter simulation for Korolev (206) crater. South is to the right.

Figure 12: Map of the mound center-of-mass offsets with respect to the crater center (normalized by crater diameter) with the mapped location of north pole dunefields and wind vectors from dune slipface measurements taken from Tanaka and Hayward (2008). The vector linking the center of mass of the ice mound to the centroid of the crater rim polygon we use as a representation of the asymmetry of the mounds. Dune density is given as the relative density parameter calculated by Tanaka and Hayward (2008) as the ratio of mean dune length vs. dune crest separation.

Figure 13: The variation of bolometric brightness temperature with season (L_s), inside and on the ice-free plains outside two example craters as derived from TES data downloaded using the TES Data Tool – http://tes.asu.edu/data_tool/ for Mars years 24 to 28. The dataset contains both data collected at 2pm and 2am local solar time; hence there are two distinct temperatures between 90-180° L_s when the sun is above the horizon at 2pm. Dokka Crater (388), a 51 km diameter crater which contains a dome of water ice, which suppresses the temperature almost all year round relative to that outside the crater. Crater 1065, at 70°N, 352°E diameter ~ 40 km, which has lower temperatures than the exterior only between L_s 250° and L_s 320°. The grey band indicates the water frost point temperatures corresponding to an atmospheric water content of between 10 and 60 prµm (Pankine et al., 2009).

Figure 14: Location of temporary ice deposits which remain after the seasonal cap has retreated, as indicated by arrows. The background is a MOLA hillshade image. A: Deposits around crater 561 at 267°E, 70°N form a plume extending to the east, matching dominant wind directions in this area (Fig. 12). CTX images P17_007799_2513_XN_71N093W and

P18_007944_2506_XN_70N092W. B: Deposits around crater 263 at 146°E, 72°N are located mainly on north-facing slopes, CTX image B01_009847_2522_XN_72N214W. Image credit MSSS/NASA/JPL.

Figure 15: Schematic diagram of the initiation and growth of ice domes, with the ices throughout as labeled in the first panel. North is to the right in all diagrams. The thickness of the frost and ice deposits are greatly exaggerated for illustration purposes. 1. In winter the microclimate in the crater provokes additional deposition of CO_2 and water ice inside the crater. 2. During the springtime sublimation, the CO_2 layer inside the crater remains (because a thicker layer takes longer to sublime), while the crater exterior is denuded of CO_2 and the H_2O starts to sublime. The H_2O released from the plains around the crater can be re-deposited in the crater interior, because of its lower temperature (as the CO_2 remains).3. Once the CO_2 has completely sublimed, relatively thick deposits of H_2O remain in the crater (H_2O from winter plus the H_2O re-condensed onto the CO_2), which act as a cold trap for H_2O subliming outside the crater. In addition the higher atmospheric pressure inside the crater reduces the rate of sublimation of this water ice. 4. The additional thickness and the reduced sublimation enable the survival of at least several meters thick deposit of water ice through the summer, which then builds up each year. A perennial deposit creates a positive feedback whereby higher thermal inertia and higher albedo than the surrounding landscape promotes continued deposition of water ice.

Figure 16: Perspective view of polar outliers near Olympia Mensae at 96.5°E, 74°N. Vertical exaggeration is five times. Outcrop in foreground is ~350 m tall and 10 km wide at the base. CTX images P21_009295_2550_XN_75N263W and P20_009018_2543_XI_74N262W used for topographic drape. Background hillshade and topography from the MOLA gridded 256 tile. CTX image credit MSSS/NASA/JPL.

Supp. Mat. Figure Captions

Figure S1: MOLA hillshade image and cross sections for all 18 craters with mounds. The black line on the image corresponds to the black north-south cross section on the plot, where “N” on the image indicates the northern end. The red line indicates the position of red east-west cross section on the plot (always oriented E-W apart from for crater 503, Louth). Data from the 256 pix/deg MOLA gridded data were used to produce the hillshade and cross sections for craters 206, 503, 579, 811, 814, 882, 904 and 934; for the others the data are derived from the 512 pix/deg gridded data.

Table 1: Summary of the characteristics of the mounds and their host craters.[a]

Crater ID	Depth (m)	Diameter (km)	Longitude (°E)	Latitude (°N)	Distance to cap (km)	SwathID	Predicted depth (m)	Max. mound thickness (m)	Mound volume (km^3)	Garvin¤ ID	Garvin¤ mound volume (km^3)	Garvin¤ predicted initial depth (m)	Presence of Dunes	Mound planimetric area		Mound relief		Max. slope (°)	Mean slope (°)	Mound summit to rim distance (m)	
														absolute (km^2)	as % of crater area	absolute (m)	as % of depth			absolute	as % of depth
206 Korolev	3102	82.83	164.49	72.75	624	5	3133	1845	3848.5	E	1356	2860	none	3134	56	1818	59	33.5	3.5	1284	41
332[#]	1364	21.35	195.59	76.97	330	5	n/c	n/c	n/c	G	2.5	660	partial	n/c	n/c	117*	9*	n/c	n/c	1039*	76*
388 Dokka	2508	51.26	214.29	77.17	301	6	3053	1890	1318.9	D	463	1540	none	1115	53	1443	58	28.7	4.0	1065	42
436	1248	19.66	190.08	81.59	60	5	2069	1640	95.9	A	59	640	partial	129	43	909	73	23.1	6.2	339	27
480	936	16.13	240.14	78.11	119	6	930	383	4.4				100%	30	14	407	44	22.7	5.7	529	56
503 Louth	1788	36.15	103.24	70.17	568	4	1766	267	13.3				partial	136	12	400	22	14.5	3.2	1388	78
515	250	11.21	117.25	81.36	0	4	305	139	2.4				none	37	37	118	47	3.8	1.1	133	53
544	773	20.96	255.07	81.27	0	6	788	378	21.0	F	35	610	none	107	42	686	89	10.4	3.7	87	11
577	248	11.73	88.93	81.16	0	4	284	133	2.4				none	32	52	157	63	4.5	1.4	72	31
579	1173	30.53	89.13	77.11	130	4	1517	932	201.6	B	248	1170	none	485	65	736	63	19.4	2.5	437	37
663	1451	24.60	60.92	79.13	0	3	1836	1017	59.1	H	3.4	600	partial	126	28	822	57	23.7	4.3	629	43
697	151	9.52	47.62	79.76	0	3	161	94	1.0				none	21	46	103	68	3.8	1.6	13	11
769	1058	20.19	331.72	78.59	8	1	1103	512	21.1	C	34	680	partial	96	29	685	65	15.1	4.9	374	35
795	371	12.55	347.00	78.61	0	1	371	178	3.1				none	41	32	274	74	6.8	2.5	98	26
811	571	18.82	309.02	71.20	452	2	773	565	40.4				100%	138	49	380	67	17.2	3.3	191	33
814	694	16.95	319.18	74.25	257	2	666	408	20.7				100%	97	41	457	66	19.8	4.1	237	34
882	572	16.32	340.23	75.05	190	1	543	150	1.8				100%	29	13	192	33	9.2	2.5	381	67
904	993	22.59	346.90	74.63	186	1	942	121	0.8				100%	84	20	178	18	20.1	2.9	488	73
934	666	15.86	348.27	73.54	249	1	653	240	7.2				100%	28	14	368	55	9.8	2.2	626	63

Min.	151	9.52		70.17	0	161	94	0.8		21	12	103	9	3.8	1.1		13	11
Mean	1048	24.33		76.97	183	1161	605	314.7		326	36	563	54	15.9	3.3		495	44
Median	936	19.24		77.17	130	859	381	17.0		96	39	403	59	16.2	3.2		381	41
Max.	3102	82.83		81.59	624	3133	1890	3848.5		3134	65	1818	89	33.5	6.2		1388	78

[a] Abbreviations and symbols used: Max. = maximum, Min. = minimum, Garvin¤ refers to results in Garvin et al. (2000), # refers to the crater that has layered deposits, but no significant mound, n/c = not calculated, and * indicates that the relief was estimated from a MOLA cross profile, rather than taking the difference between the max. and min. MOLA elevations found within the mound polygon.

Highlights

Highlights:

- 18 potentially ice-cored mounds were found in craters in Mars' north polar region.
- The stratigraphy of the mounds argues for deposition from the atmosphere.
- We argue many of them were deposited separately from the polar cap.
- The crater micro-environment is a potential explanation for mound initiation.
- These mounds are sensitive and important records of Amazonian climate on Mars.

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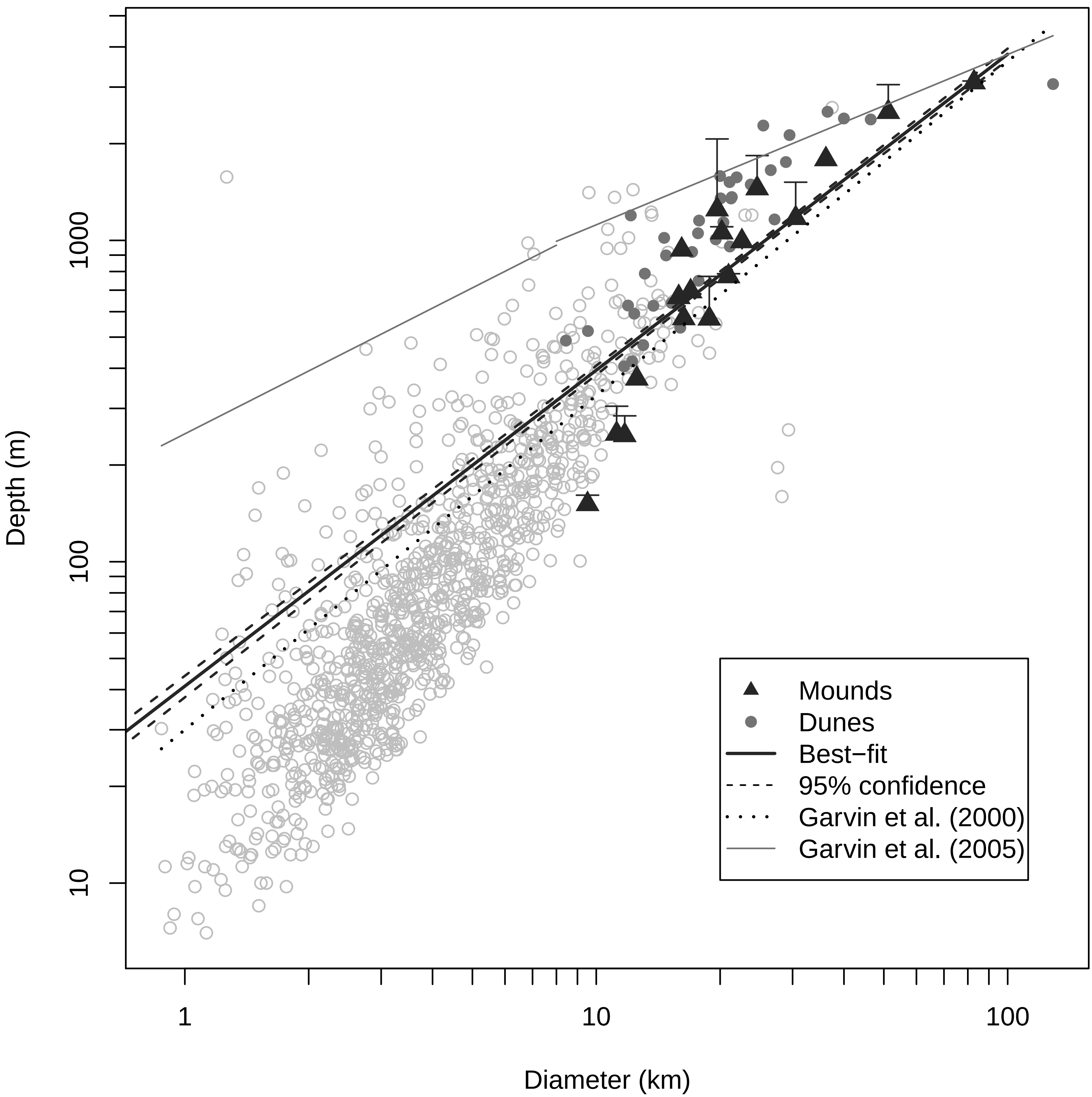
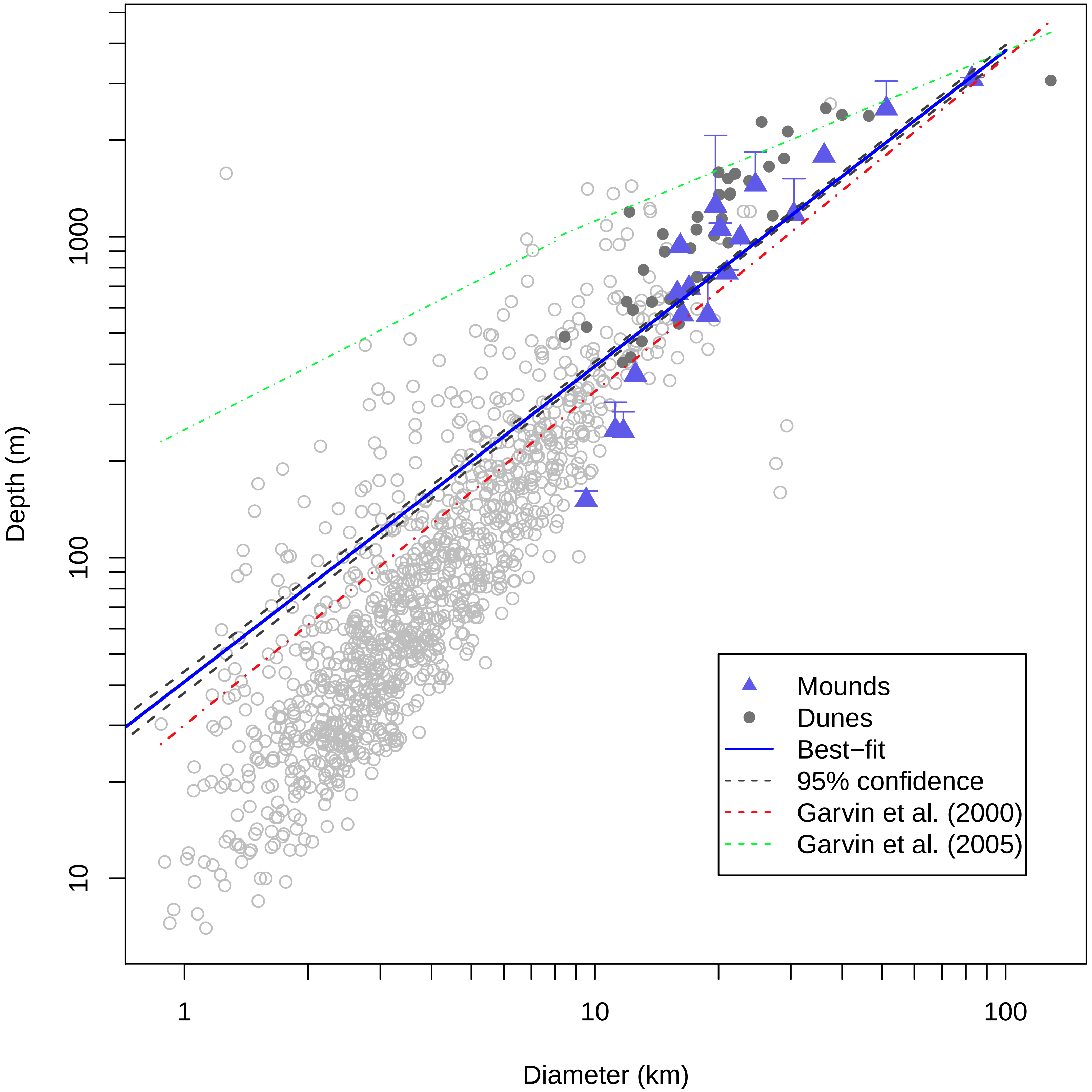


Figure1



Mounds
Dunes
Best−fit
95% confidence
Garvin et al. (2000)
Garvin et al. (2005)
Depth (m)
Diameter (km)
10
100
1000
1
10
100

Figure 2 (black and white)
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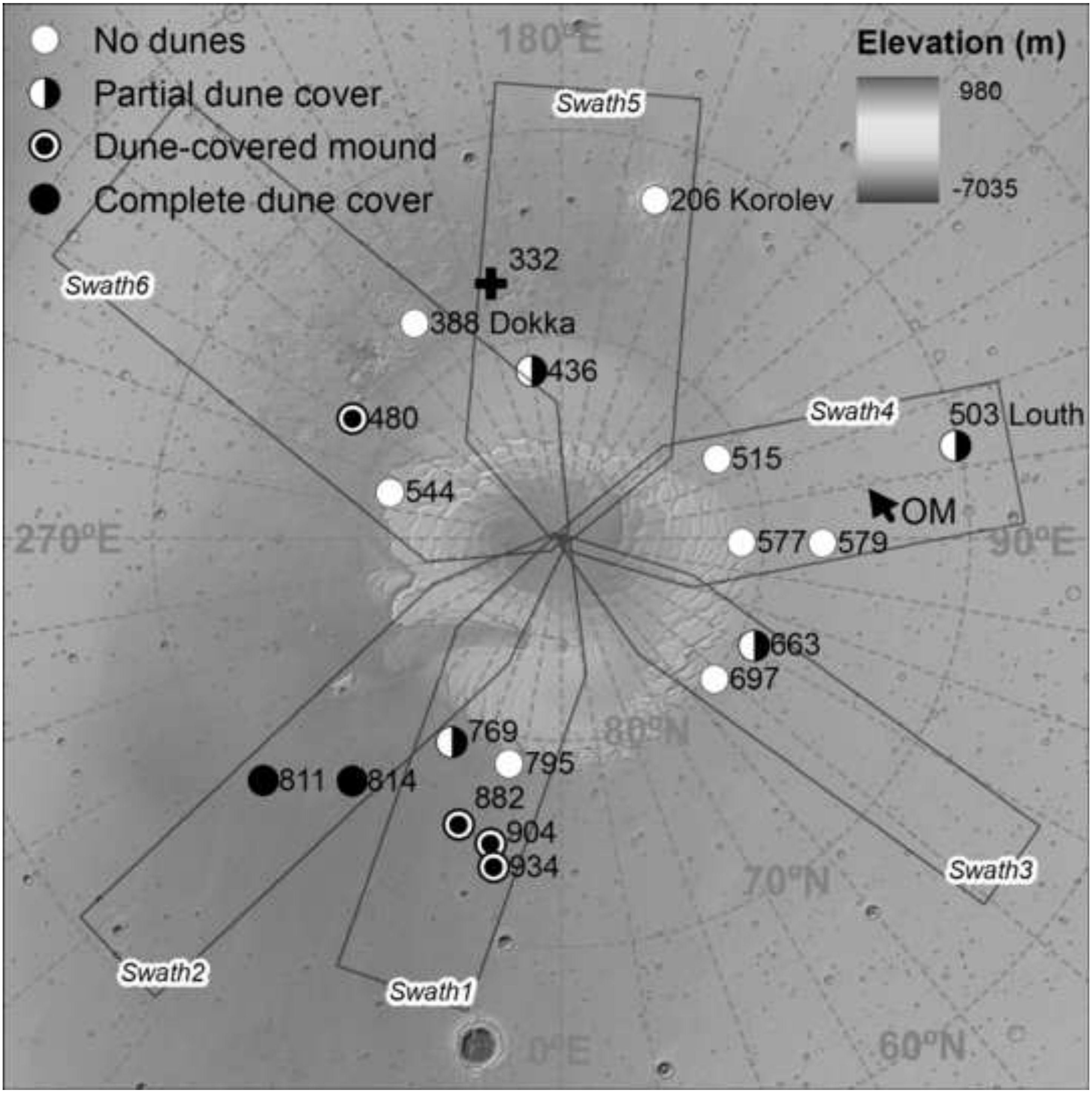

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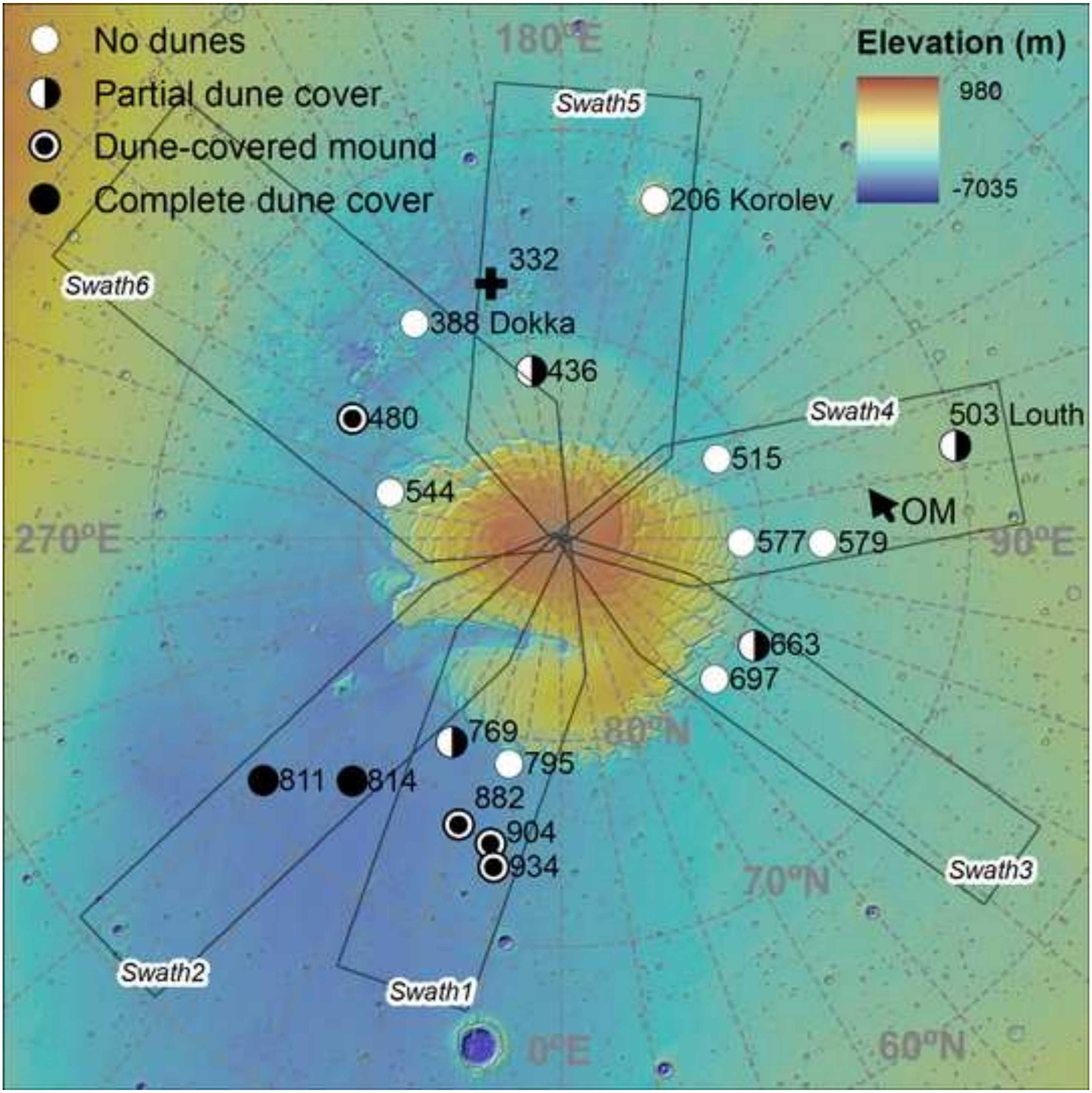

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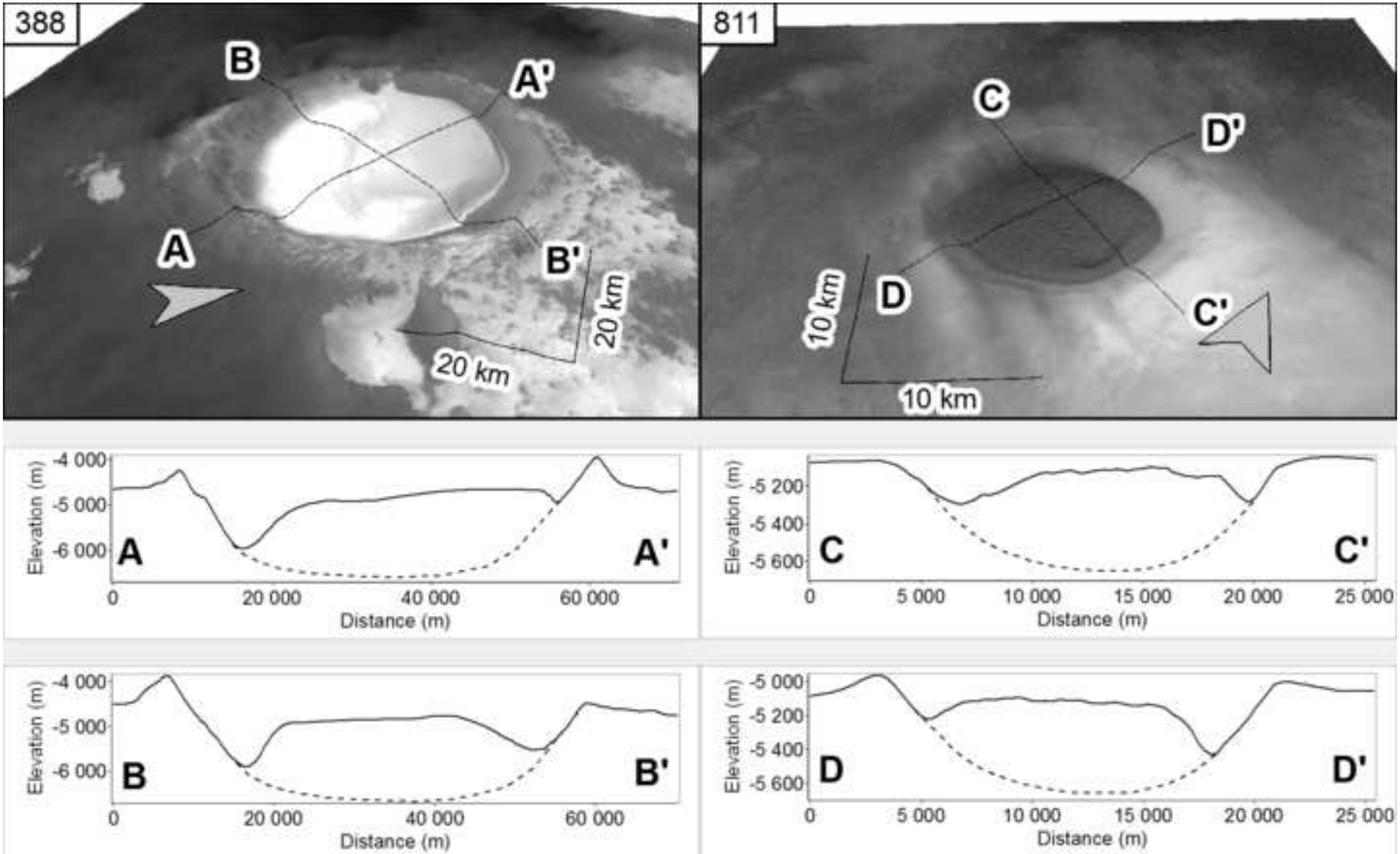

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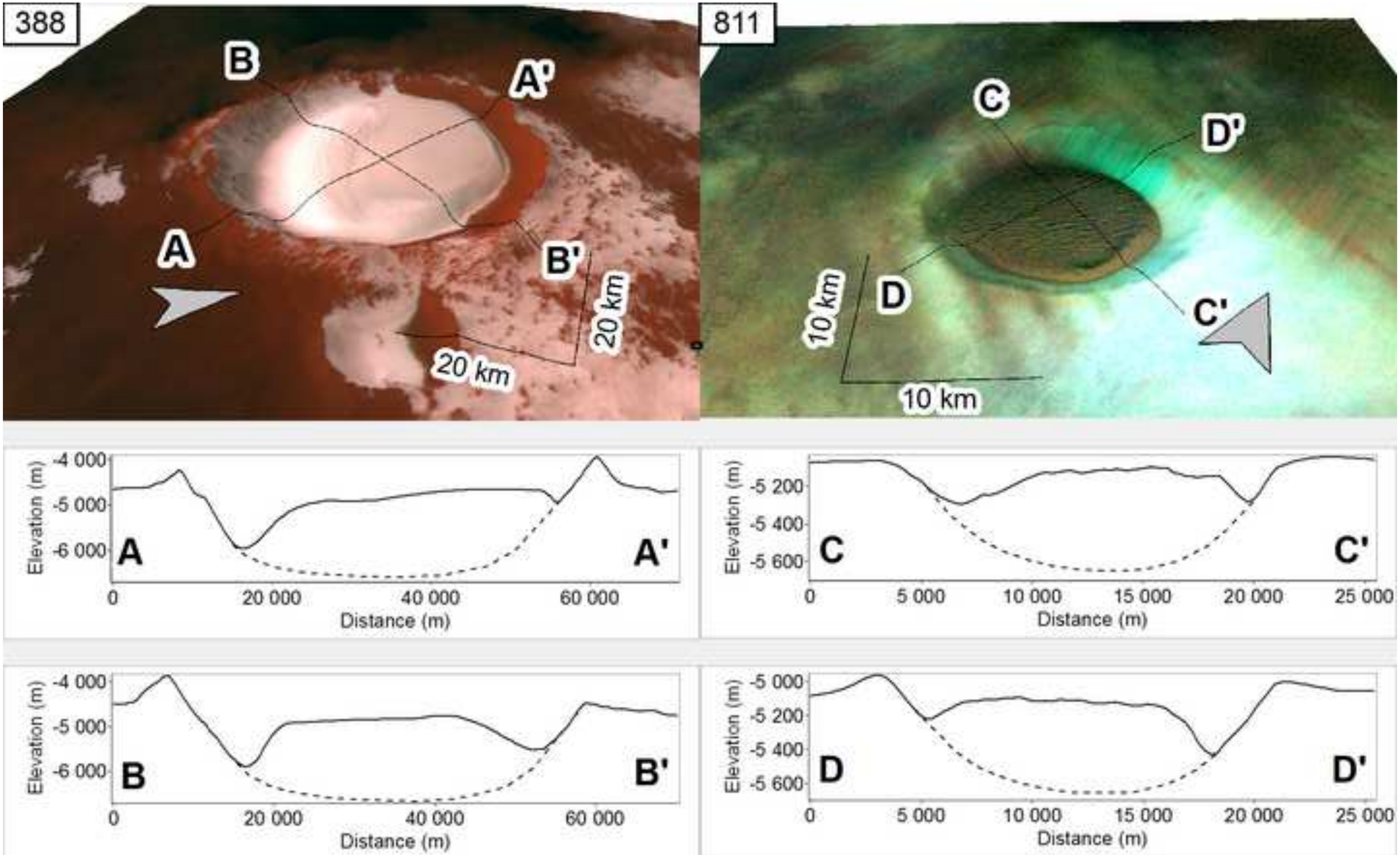

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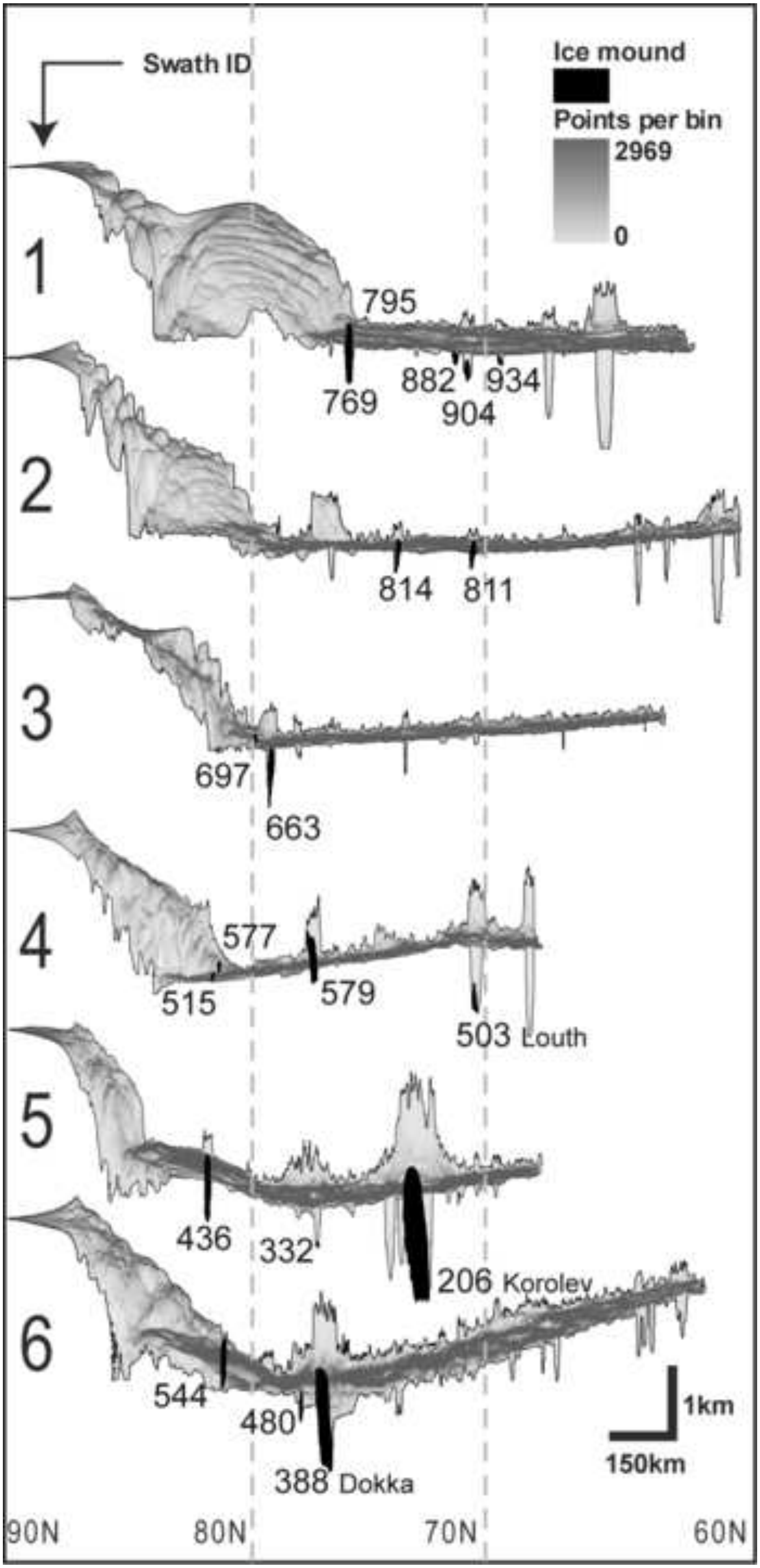

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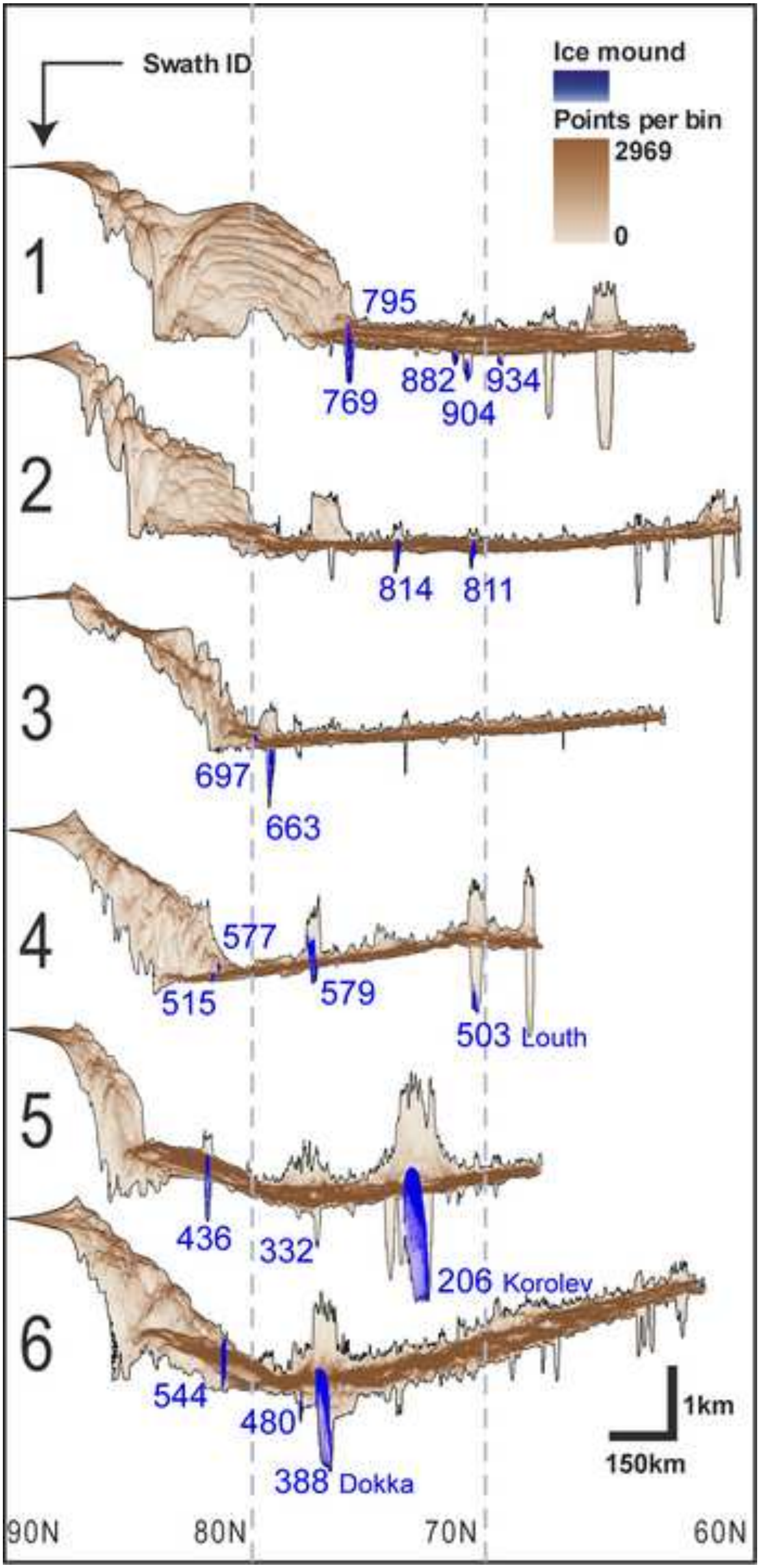

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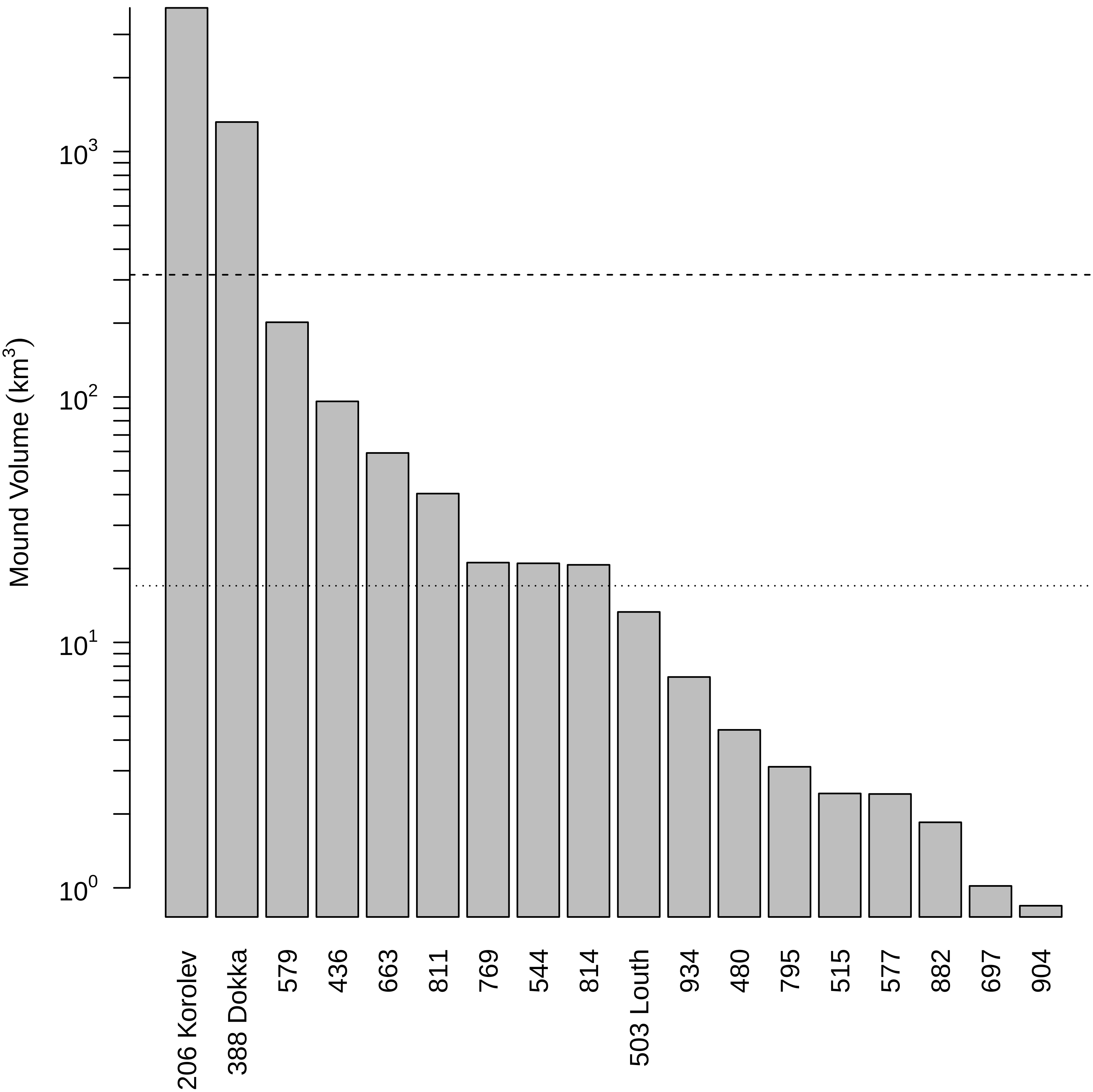

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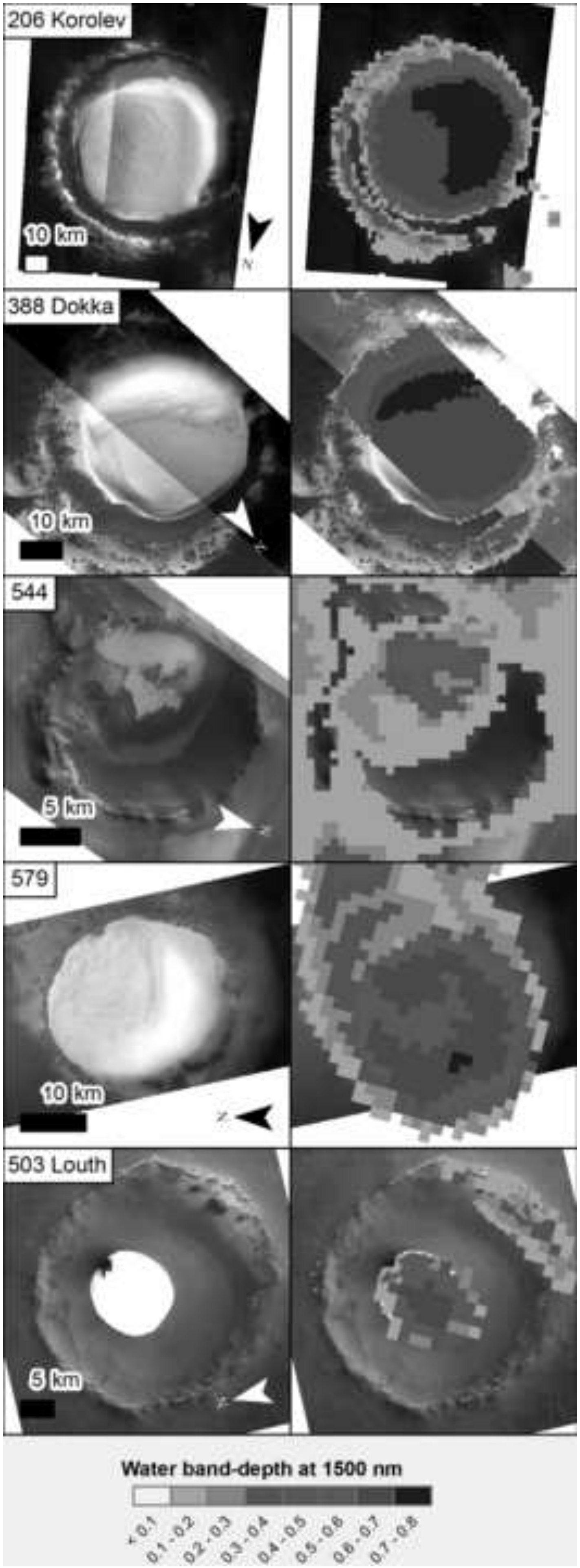

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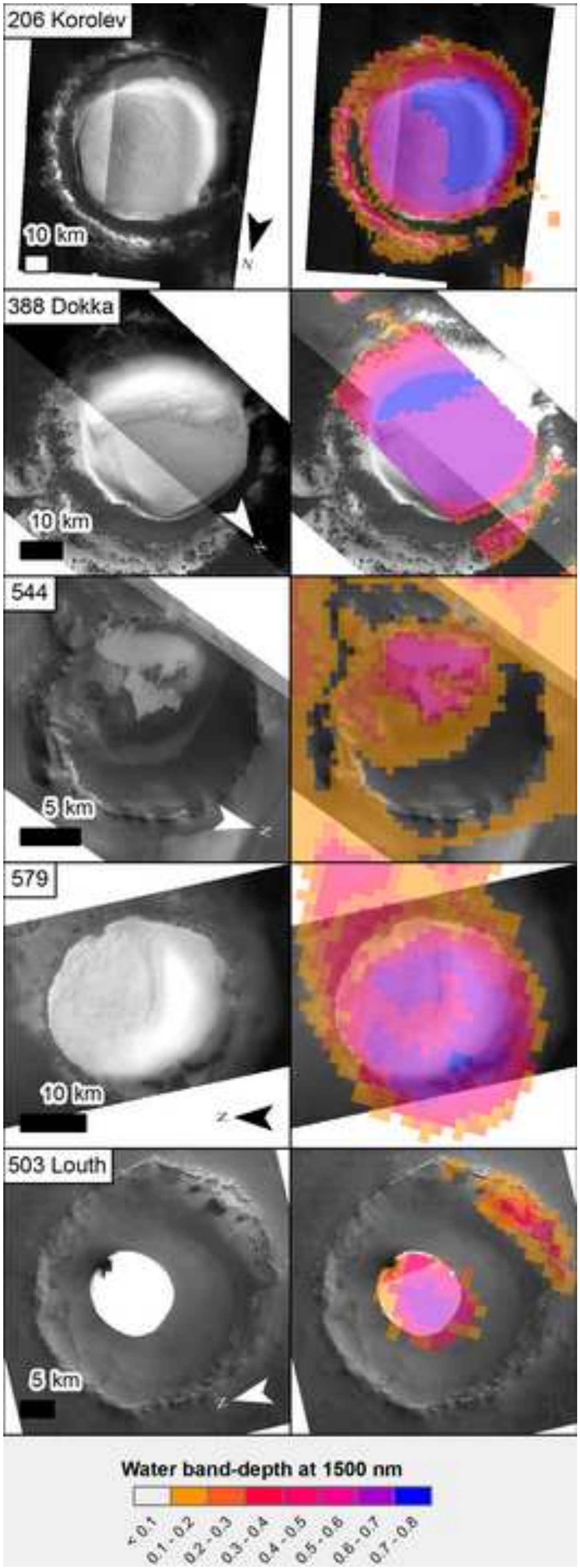

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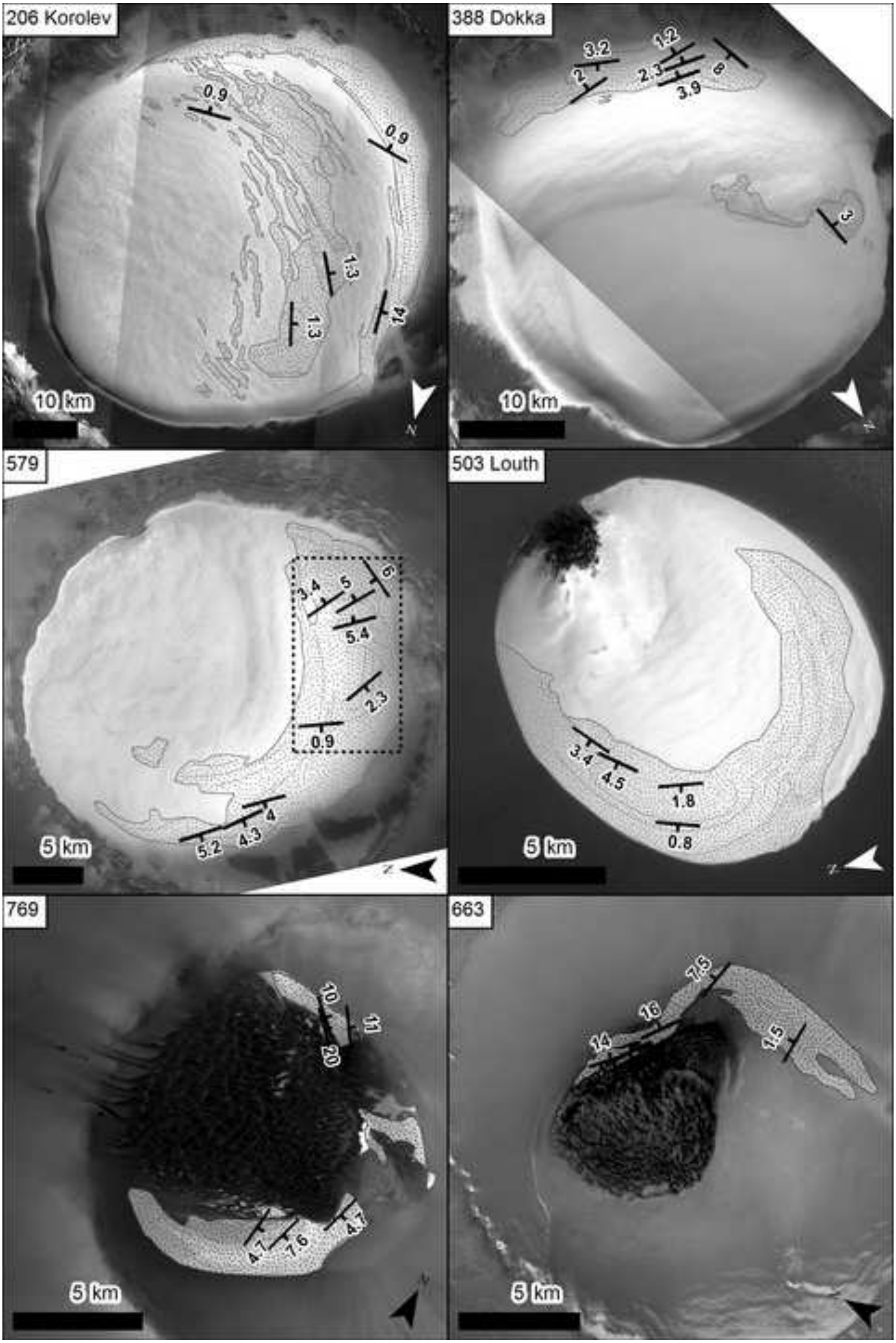

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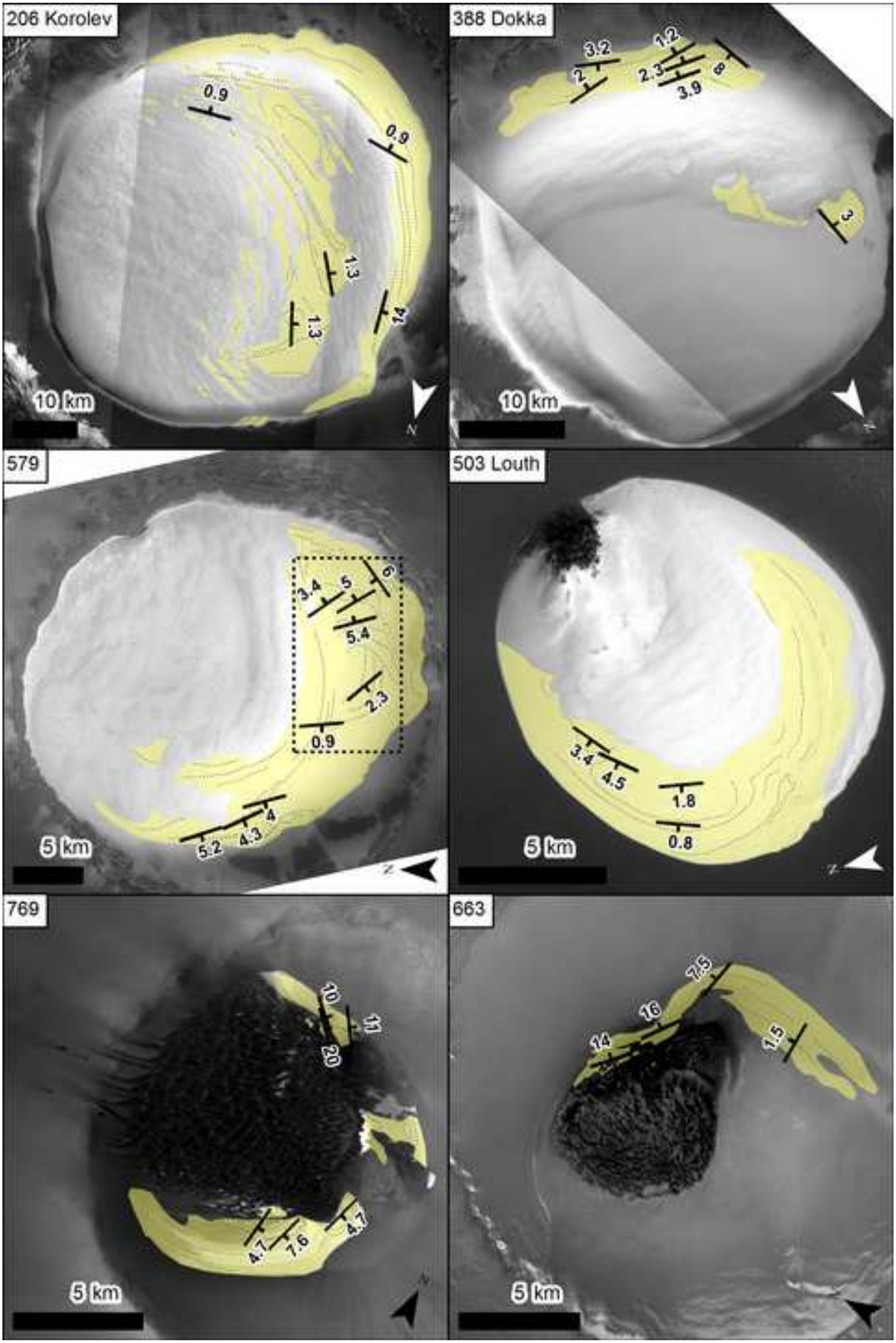

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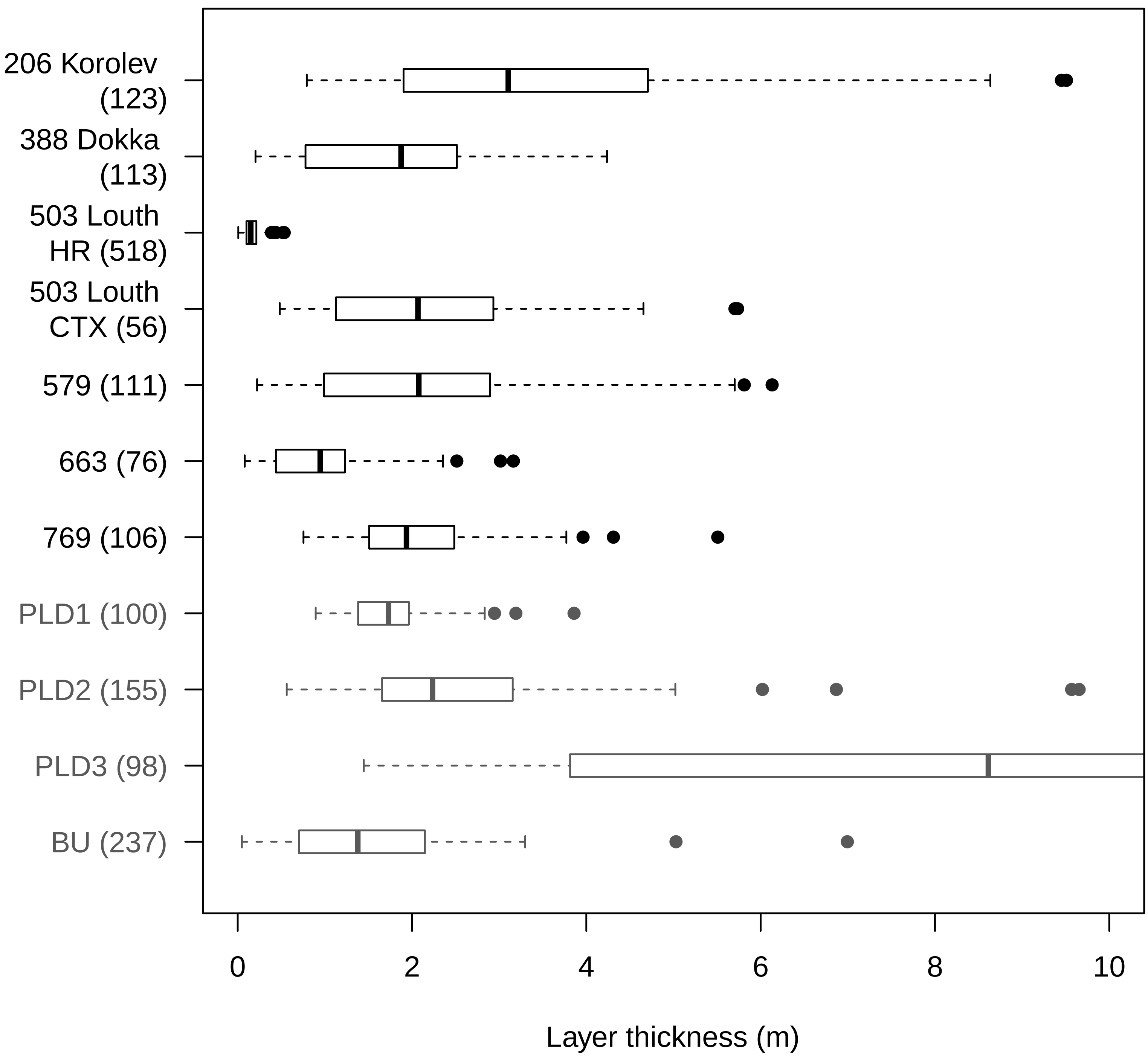

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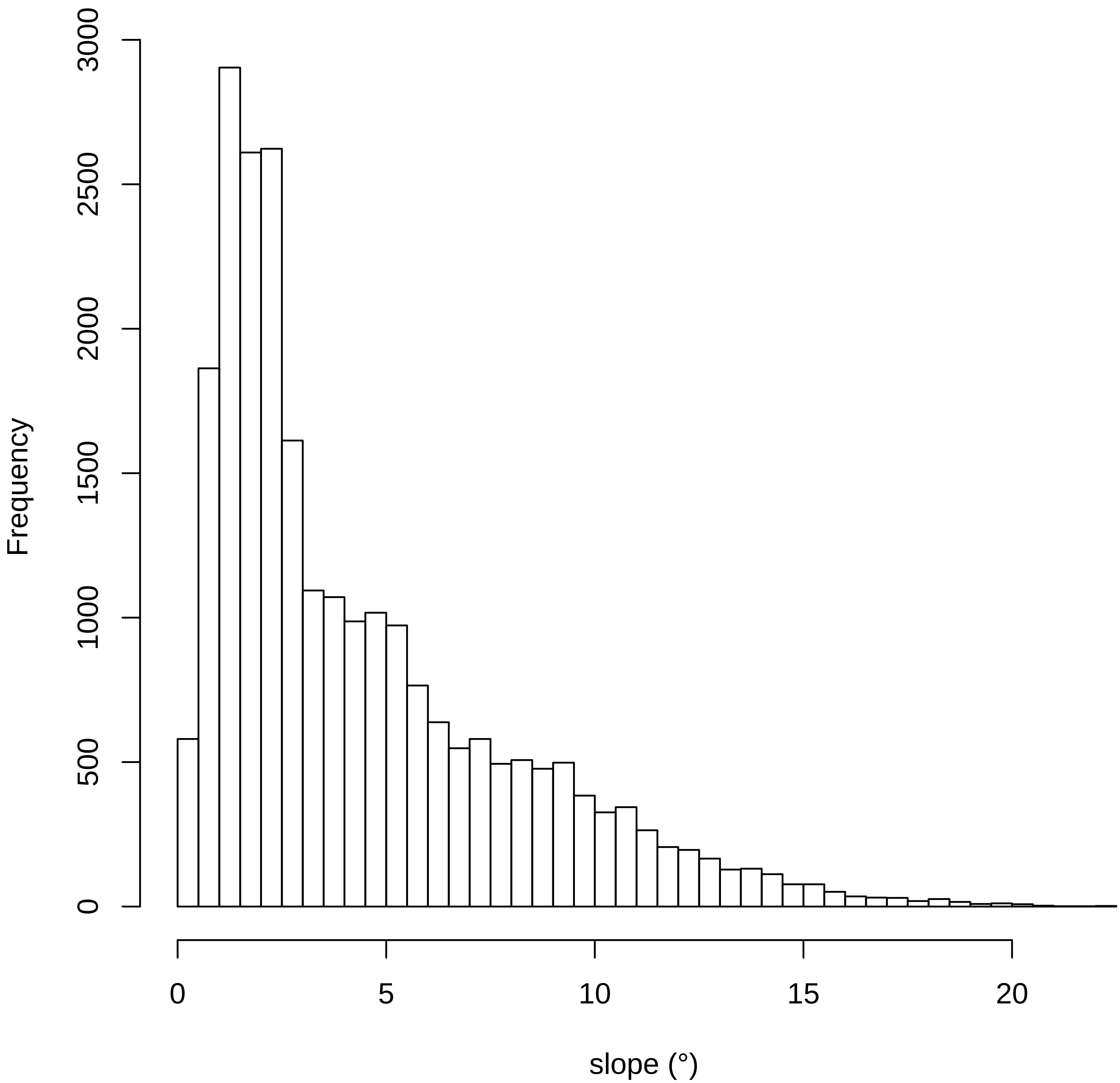

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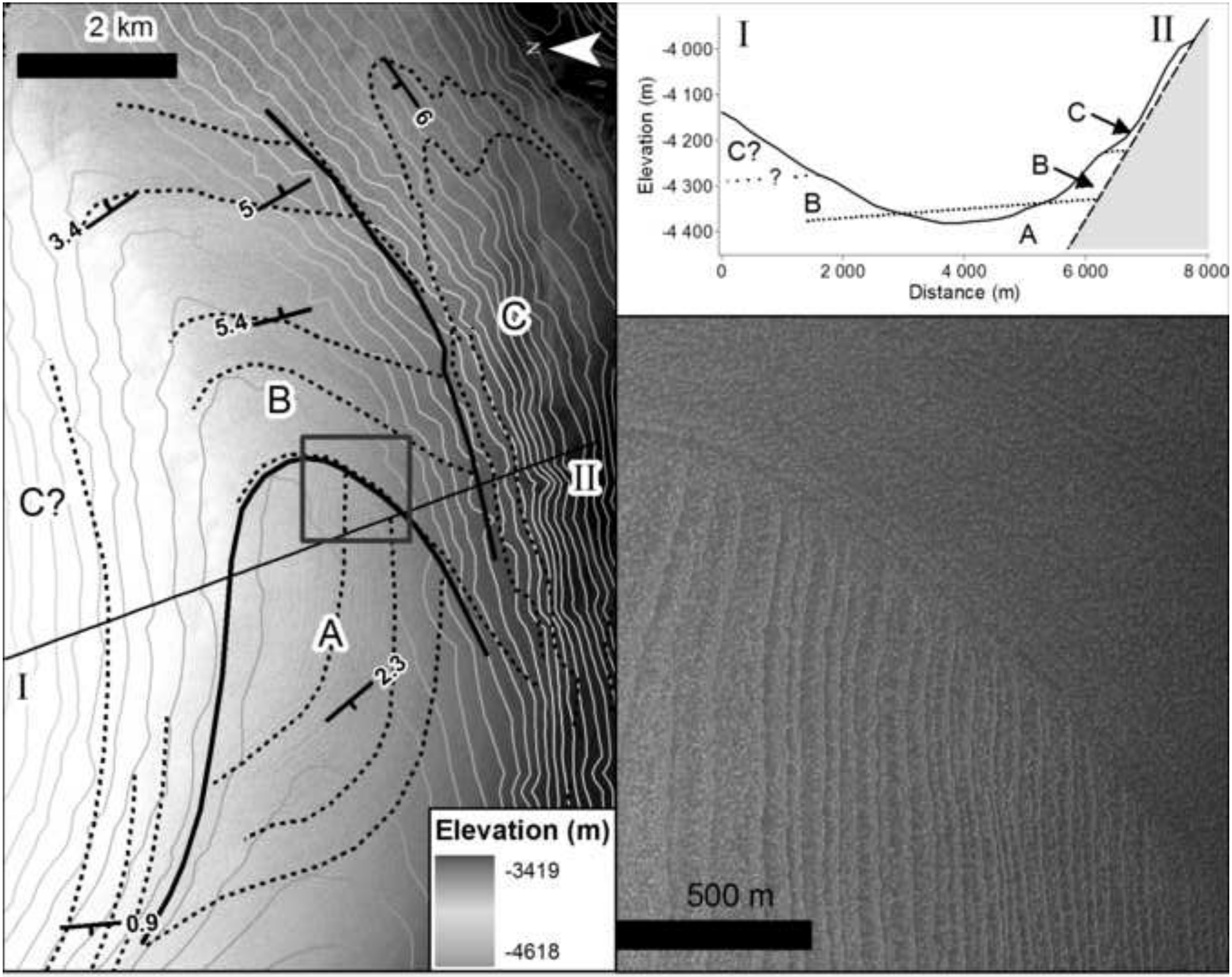

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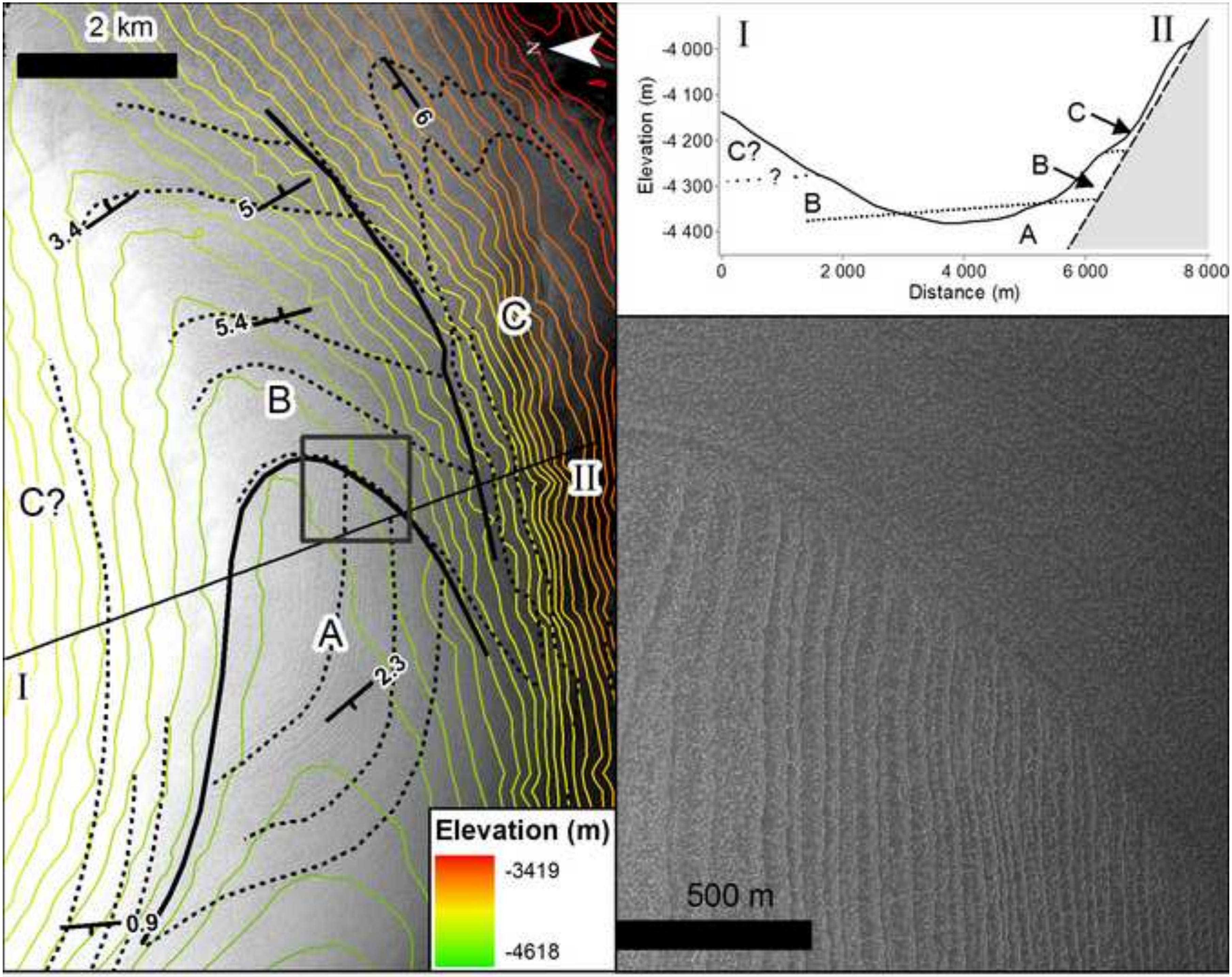

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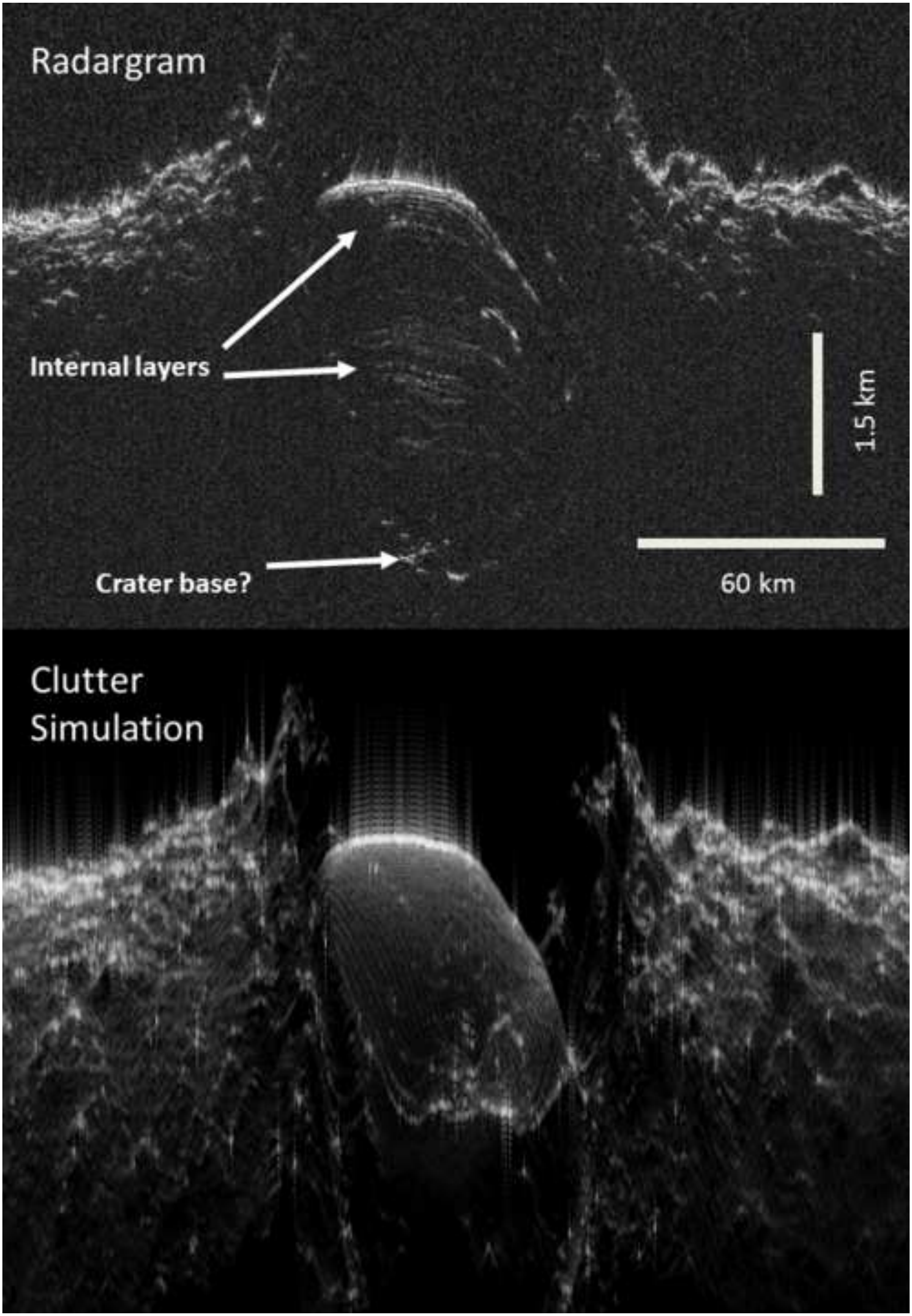

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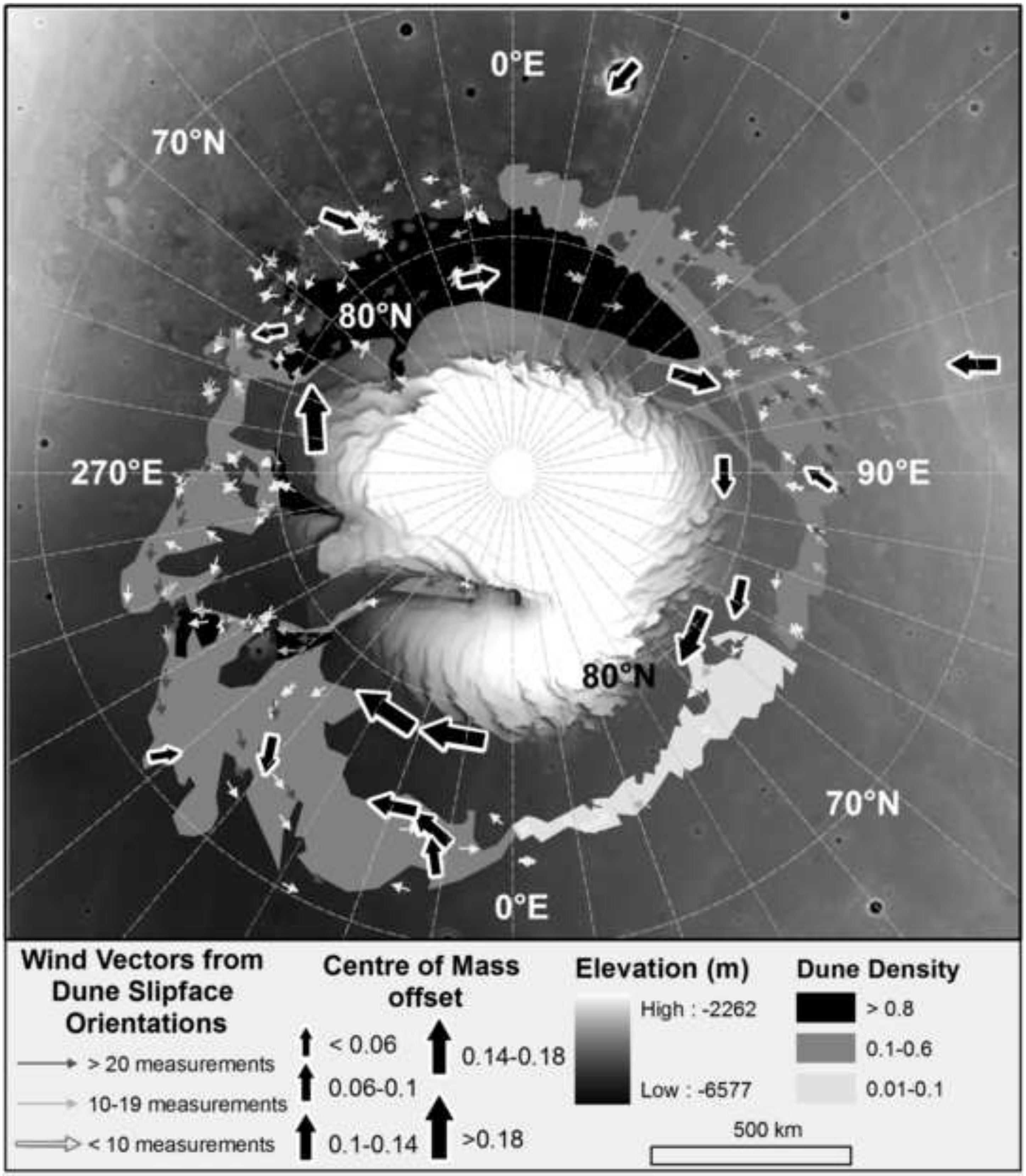

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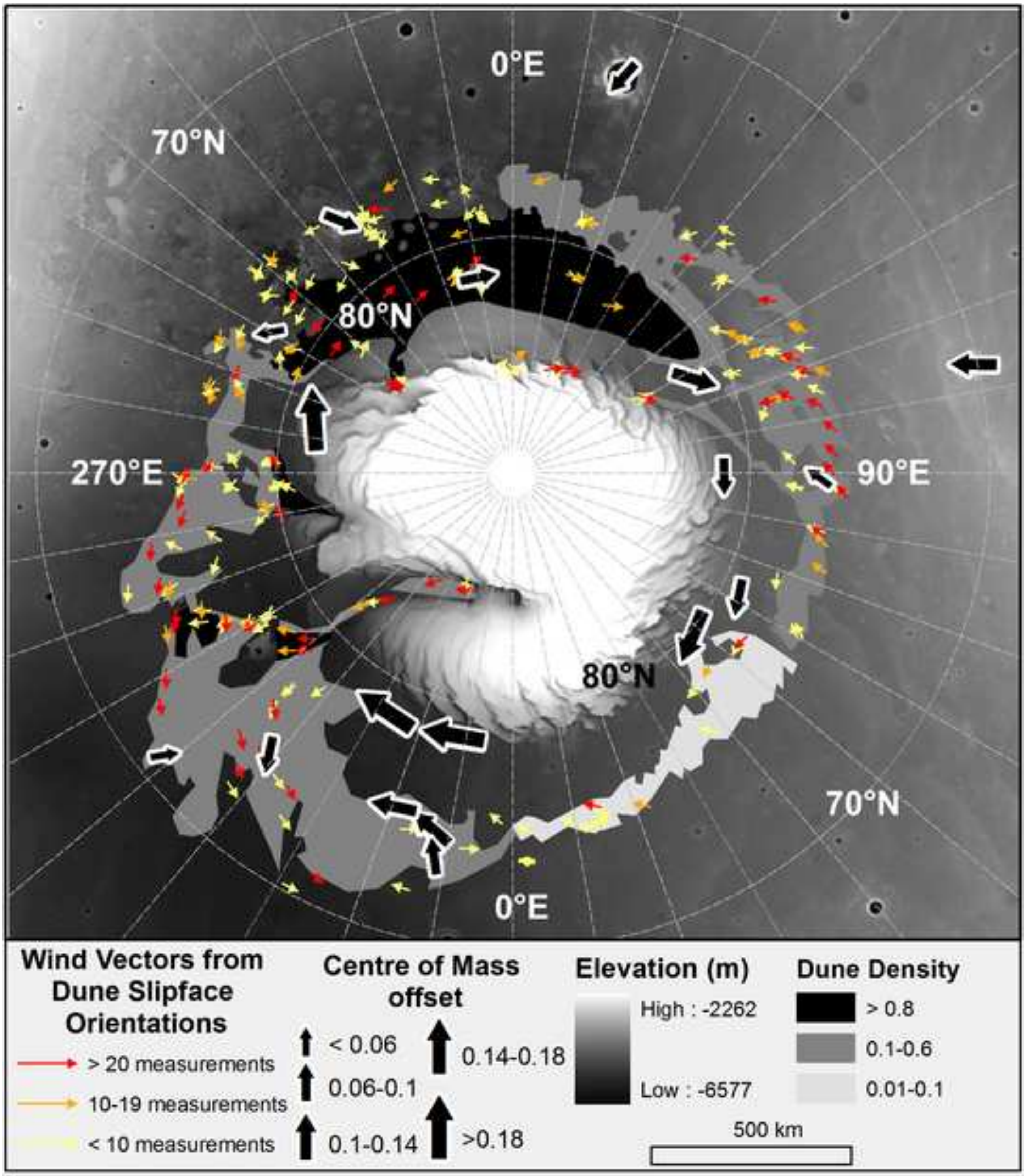

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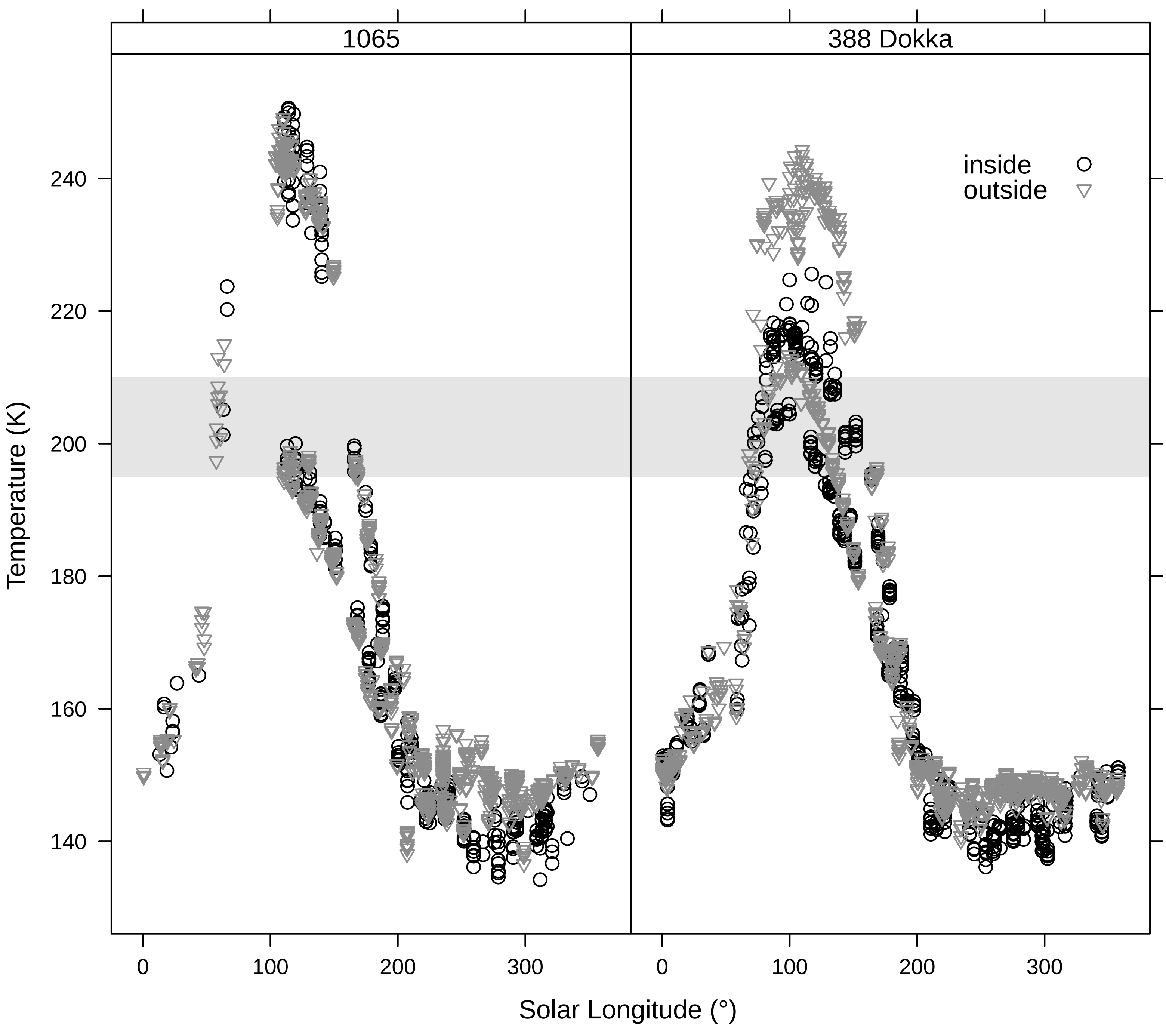

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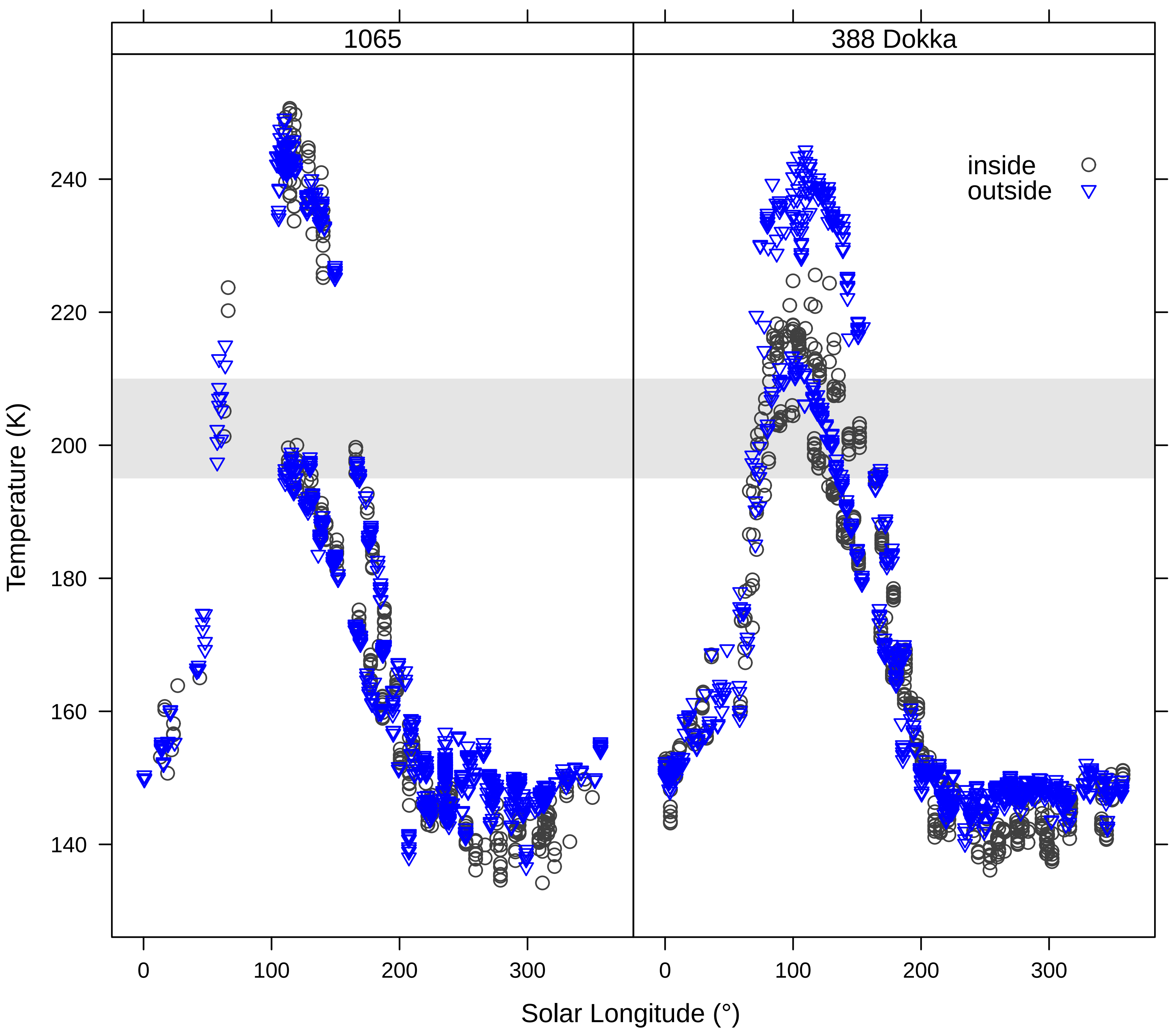
1065
388 Dokka
inside
outside
Temperature (K)
Solar Longitude (°)
240
220
200
180
160
140
0
100
200
300

Figure 14 (black and white)
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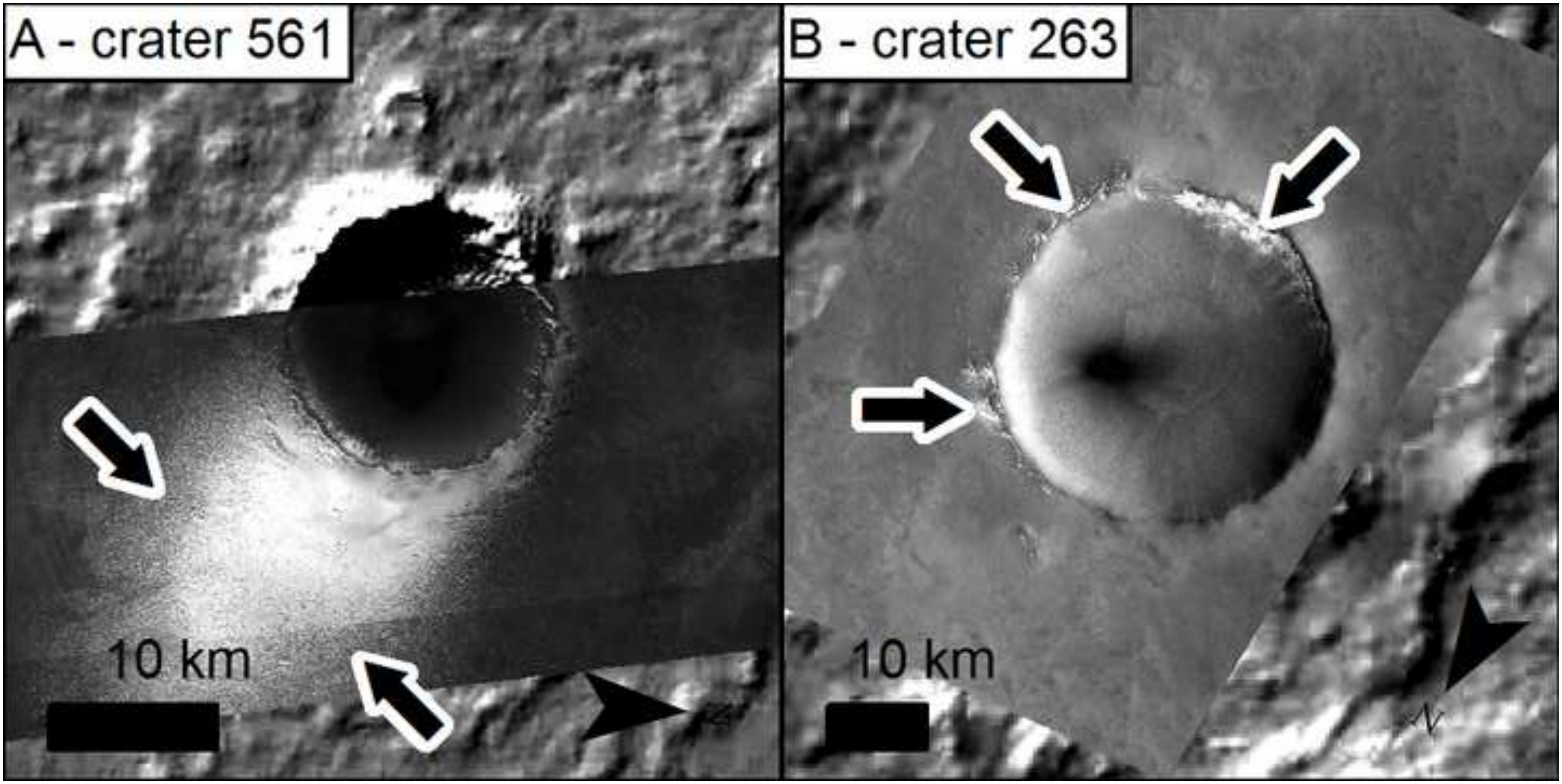

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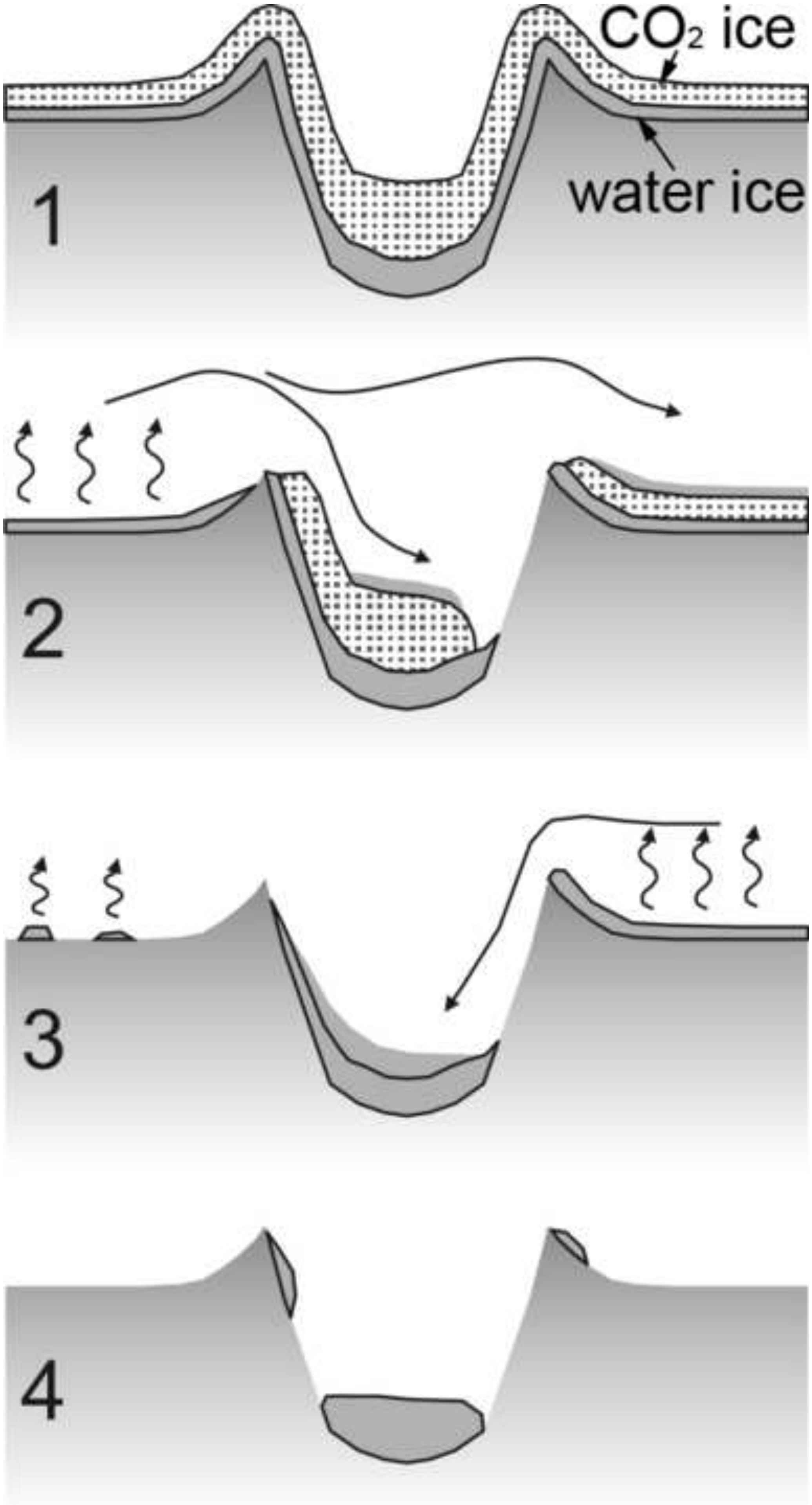

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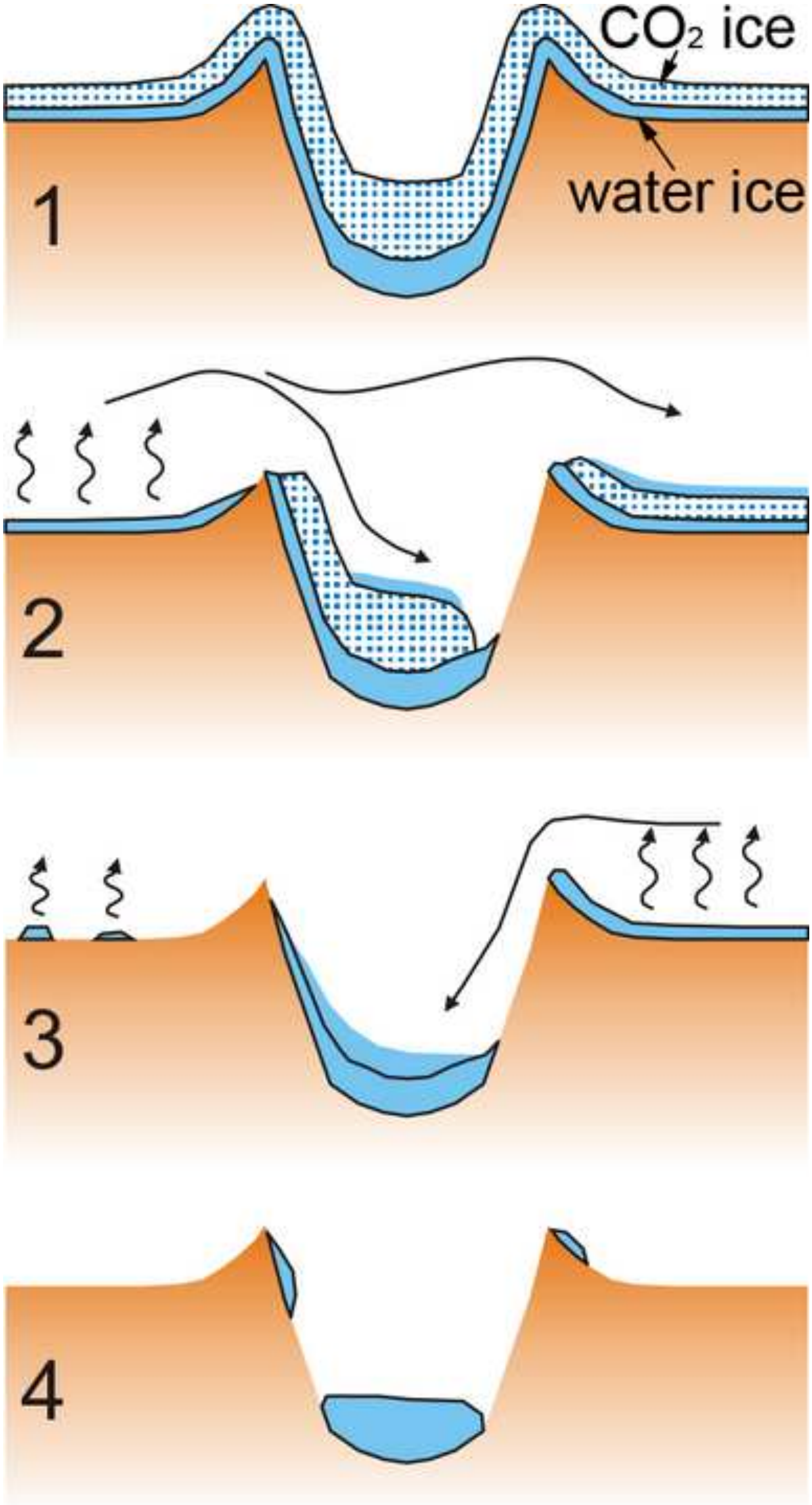

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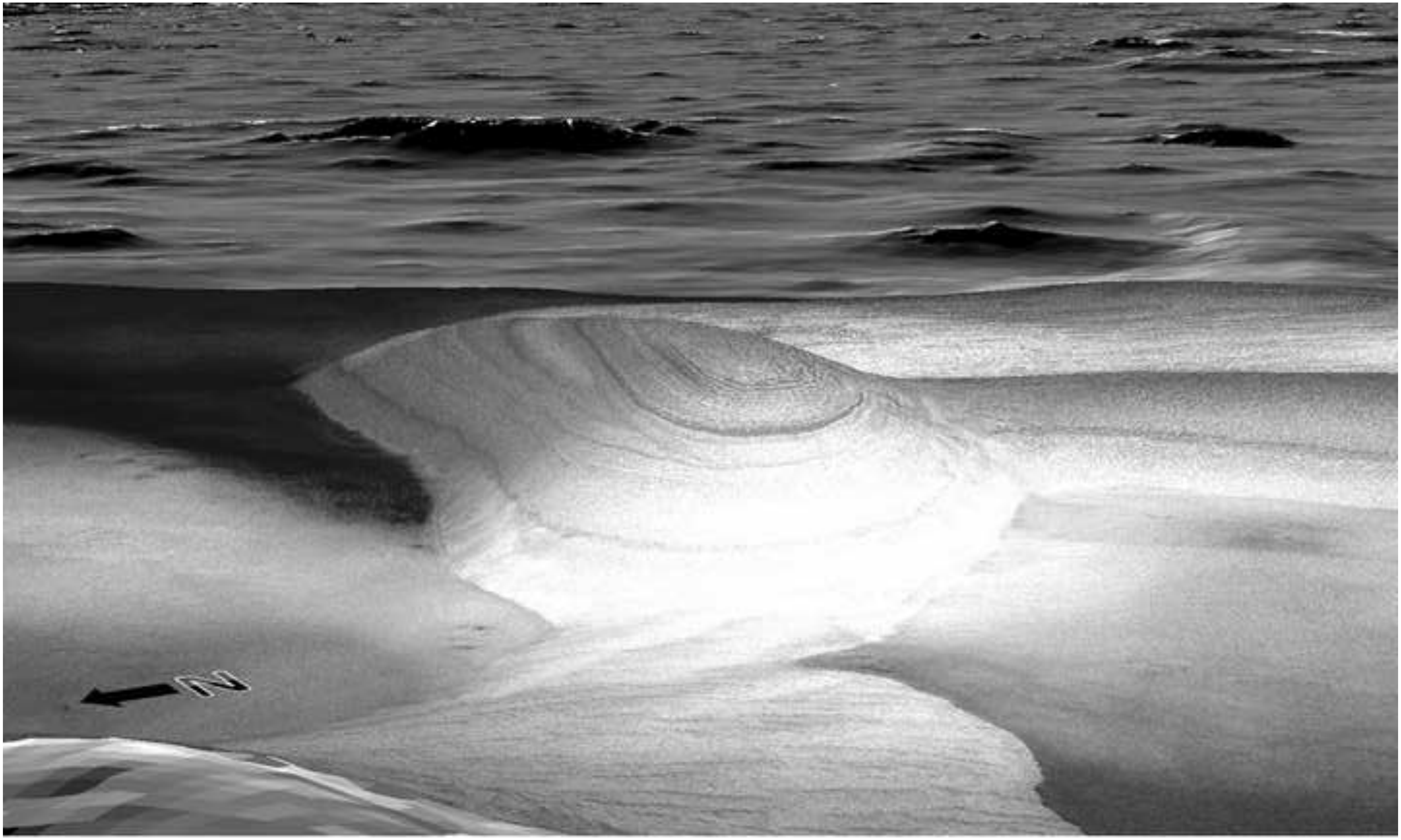

Sup. Mat. Fig. S1