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# **Grid-mapping the northern plains of Mars: using morphotype and distribution of ice-related landforms to understand multiple ice-rich deposits in Utopia Planitia**

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## Key Points:

- Distribution and morphotype of ice-related landforms were used to understand the permafrost cryolithology and evolution in Utopia Planitia.
- A mid-latitude, ice-rich, stratified permafrost (50–85% by volume) of 11 Ma age occurs from 38°N to 47°N.
- The latitude dependent mantle of about 1.5 Ma age is present from 47°N to 78°N.

**Keywords:** Mars, Permafrost, Periglacial processes, Ground-ice, Thermokarst

## Abstract

This work in Utopia Planitia is the first continuous regional mapping of ice-related landforms integrated into an effort to study the three main basins (Arcadia, Acidalia, Utopia Planitia) in the northern plains. The distribution and morphotypes of these landforms, SHARAD detections, and crater size-frequency distribution measurements (>50 m in diameter) were used to understand the permafrost cryolithology and its past evolution in relation to climate in Utopia Planitia. Three assemblages of landforms were identified based on their spatial correlation and correlation with MOLA surface roughness along a strip from 30°N to 80°N. At 30°–38°N, the assemblage is formed by km-scale polygons, high-albedo mounds, and thumbprint terrains. This assemblage is associated with a lobate deposit of 30 m in thickness with a crater retention age of 1 Ga. At 38°–47°N, the assemblage is comprised of large scallops, 100 m diameter polygons, pits, and mantled deposits. This assemblage is correlated with a deposit of 80 m in thickness containing excess ice (~50–85% by volume) with a crater retention age of about 10 Ma. At 47°–78°N, the assemblage is composed of mantled deposits, textured terrains, and 30 m diameter polygons. This assemblage is related to the ice-rich, debris-covered, latitude-

dependent mantle that has a crater retention age of about 1.5 Ma. Utopia Planitia appears to be a region of combined depositions of sediment and continuous cold climatic conditions that led to a complex distribution of ground ice.

### **Plain Language Summary**

The northern plains of Mars comprise several basins filled by sediments. The region has been proposed to have hosted an ancient ocean and currently contains ice in the ground even at latitudes where the ice is not stable. It is not known, however, what is the origin of the ice, whether it is related to the ancient ocean or, recent glaciations. The ages of different surfaces and landforms are also not well known. Improving the geological context of the northern plains will help constrain outstanding questions about evolution of the climate and geology on Mars.

An International Space Science Institute team project, composed of several European laboratories with expertise in mapping, has been convened to study targeted areas in the northern plain of Mars. In Utopia Planitia, we mapped and used the distribution of ice-related landforms and radar sounding to understand the distribution of ice over the region. Two young ice-rich deposits with different ice-content were identified over the mid and high latitude of the region. Similar to some regions in the Arctic on Earth, Utopia Planitia appears to be a region of combined deposition of sediment and continuous cold climatic conditions that partially preserved several ice rich deposits.

## 1. Introduction

The northern plains of Mars, topographically lower than the southern cratered highlands, comprise several large basins that are variably filled by materials stratigraphically younger than the highlands (e.g., Baker et al., 1991; Kreslavsky and Head, 2002a). Many landforms found in the northern plains are indicative of ground ice, including debris-covered glaciers referred to as viscous flow features (Carr, 2001; Squyres, 1978), scalloped depressions (e.g., Costard & Kargel, 1995; Morgenstern et al., 2007; Séjourné et al., 2011), thermal contraction polygons (e.g., Seibert & Kargel, 2001; Mangold, 2005), gullies (Malin & Edgett, 2000).. Presence of subsurface ground ice at high-latitudes has been confirmed by spectral measurements of hydrogen abundance (e.g., Feldman et al., 2004) and by direct observations from the Phoenix lander (Mellon et al., 2009). Its distribution is correlated with the current near-surface ground ice stability for latitudes  $>60^\circ$  (Haberle et al., 2003; Mellon & Jakosky, 1995). Ground ice at a depth greater than 1 m has been detected by direct observations of fresh craters (Byrne et al., 2009) and sublimation depressions (Dundas et al., 2018), as well as by radar sounding (Mouginot et al., 2012; Plaut et al., 2009).

However, no consensus has emerged as to what process was responsible for the presence of ground ice in the northern plains. During the Hesperian, these lowlands may have been the location of accumulation and later freezing of sediments from multiple episodes of outflow channel discharge, and perhaps even from a large ocean (e.g., Baker et al., 1991; Carr & Head, 2003; Clifford & Parker, 2001; Parker et al., 1993). During the Amazonian, significant variations in the orbital parameters are thought to have induced cyclic accumulations of ice in different areas of the northern plains. Variations in obliquity and eccentricity control the pattern of insolation and, thus, the climate (e.g., Chamberlain & Boynton, 2007; Haberle et al., 2003; Laskar et al., 2004; Mellon & Jakosky, 1995).

An ice-rich and geologically young (0.4–2.1 Ma) latitude-dependent mantle (LDM) is present and smooths the topography at latitudes  $>30^\circ$  in both hemispheres (Head et al., 2003; Mustard et al., 2001). Climatic models have simulated the formation of the LDM from ice deposition during periods of low or moderate obliquity suggesting a link with climate changes due to orbital cycles (Forget et al., 2006; Haberle et al., 2003; Head et al., 2003; Levrard et al., 2004; Mischna et al., 2003).

Moreover, along the dichotomy between the southern highlands and the northern plains (specifically from  $30^\circ$  to  $50^\circ\text{N}$ ), viscous flow features are interpreted to represent the remnants of glaciation formed during the Late Amazonian (Carr, 2001; Hauber et al., 2005; Head et al., 2006, 2010; Mangold, 2005). These glacial features could have been formed by ice deposition during moderate-obliquity periods (Head et al., 2006; Madeleine et al., 2009). Additionally, other ice-rich mantles were identified in multiple northern lowland locations and possibly predating the LDM (Kadish & Head, 2014; Kargel et al., 1995).

Improving the geological context of the northern plains will help improve our understanding of the recent geological evolution. Furthermore, although the spatial distribution of some ice-related landform types has been measured and correlated with latitude-controlled climatic processes, broad-scale heterogeneity in surface geomorphology exists within latitude bands, suggesting the potential importance of regional geology and climate. Therefore, the spatial distribution of landforms thought to be indicative of ground ice must be assessed in more detail and in a representative way, and the types of materials in which these forms occur must be determined.

A project of strategic, targeted mapping of water- or ice-related surface landforms was therefore undertaken to understand the geological and stratigraphic character of the martian northern plains, with particular regard to the role that ground ice has played in their evolutionary history. We used a grid mapping approach (e.g., Ramsdale et al., 2017) with the latest mid- to high-resolution images of Mars to study its northern large basins (Arcadia, Acidalia, Utopia Planitiae). The aims were to compare and to contrast the assemblages of landforms and their latitudinal distribution across these three regions, and to attempt to determine the origin of some of these materials (Orgel et al., this edition; Ramsdale et al., this edition). Grid mapping is an efficient method to deal with the problems of mapping small landforms over large areas by providing a consistent and standardized approach to spatial data collection (Ramsdale et al., 2017; Voelker et al., 2017).

Here, we describe our geomorphological mapping of western Utopia Planitia (UP) along a 300 km wide strip from 30°N to 80°N and centered on about 90°E (Figure 1a). The goals were to: (i) characterize the morphotypes and patterns of the spatial distribution of the landforms; (ii) identify their association with subtly expressed geological units and; (iii) interpret the cryolithology (ice content, sediment type) of the permafrost and environmental evolution in UP.

## **2. Geological background of Utopia Planitia**

UP centered at ~45°N and ~110°E (Figure 1a), is one of three major topographic basins in the northern hemisphere of Mars (McGill, 1989; Thomson & Head, 2001). The average elevation within the studied zone is about -4100 m (Figure 1b). The Vastitas Borealis formation (VBF) covers most of UP and is composed of the “Vastitas Borealis Interior” unit (ABvi) and the “Vastitas Borealis marginal” unit (ABvm) (Figure 1c; Tanaka et al., 2005). These units are interpreted to be Late Hesperian (~3.5 Ga) outflow channels sediments (Kreslavsky and Head,



2002a; Tanaka et al., 2005; Werner et al., 2011). The Late Hesperian “Utopia Planitia 2” unit (HBu2) in the south, was most probably formed by erosion and redeposition of Noachian highland material (Figure 1c; Tanaka et al., 2005). The VBF itself is overlain by the Middle to Late Amazonian “Astapus Colles” unit (ABa) which is thought to be an icy mantle emplaced by atmospheric precipitation, and possibly predating the LMD (Figure 1c; Skinner et al., 2012; Tanaka et al., 2005). In the north, the Olympia Undae unit (ABo) is formed by extensive dune fields (Tanaka et al., 2005).

### **3. Data and method**

#### **3.1 Grid mapping**

The geomorphological grid mapping was performed along a 300 km wide strip from 30°N to 80°N (from about 87°E to 93°E at 30°N; Figure 1b). To ensure a high spatial coverage of the study area and a good resolution, Context Camera images (CTX; 6 m/pixel) were used (Figure 1b; Malin et al., 2007). Complementary, High Resolution Imaging Science Experiment images (HiRISE; 25 cm/pixel; McEwen et al., 2007) helped in the study of some of the landforms. Mars Orbiter Laser Altimeter (MOLA) gridded global digital elevation data were used for regional topographic analysis (463 m/pixel), whereas detailed local topographic considerations were obtained through MOLA Precision Experiment Data Record (PEDR) laser shots analysis (vertical uncertainty of about ~1 m, horizontal precision ~100 m, with distance between individual shots on average 300 m; Smith et al., 2001, 2003).

The mapping strip was divided into a grid of 15 x 150 squares, each approximately 20 km across (Figure 2). The Cassini projection (centered at 90°E longitude) was chosen to give minimal distortion along the central meridian, ideal for a long and narrow north-south strip. Using ArcGIS software, the grid system was set-up as a polygon shapefile, in which each grid

square was represented by a single square polygon object (more details in Ramsdale et al., 2017). Grid mapping is an efficient method to map small landforms over large areas (Ramsdale et al., 2017). Across the region, different water- or ice-related landforms were identified to understand the distribution and role of ground ice in Utopia Planitia (Figure 2). More details are provided in Ramsdale et al. (2017). The landforms were recorded in each grid square with a code as either: 1 for “present”, 1.5 for “extensive”, 2 for “dominant”, 0 for “absent”, or P for “possible” (Figure 2). The code N means an absence of CTX images. The “dominant” classification was used when a single landform type covered the entire grid square to such an extent that other landforms were obscured. The end result of the mapping was a raster-like data product of the distribution of landforms across the whole strip, hence providing a digital geomorphological map (Figure 2 and S1).

### 3.2 Crater size-frequency distribution measurements of small impact craters (>50 m in diameter)

Recently, small impact craters (10–100 m scale) and high-resolution images have been used to distinguish climate events of the Late Amazonian despite some uncertainties, such as the role of secondary craters or post-impact modification of crater rim (e.g., Daubar et al., 2013; Hartmann & Daubar, 2017; Hartmann & Werner, 2010).

We measured crater size-frequency distributions (i.e., numbers of craters accumulated on surface units) using CTX images to provide a crater retention age for the different geomorphologic units (e.g., Hartmann and Neukum, 2001; Ivanov, 2001). All visible impact craters (>50 m in diameter) were measured except the aligned secondary clusters and rays within the CTX resolution limit using the CraterTools ArcGIS add-in (Kneissl et al., 2011). Three zones comprising several grid cells were selected in places without extensive degradation and dust deposits following our interpretation of assemblages of landforms (black

squares on Figure 2a and Figure S3). Counting areas  $>1000 \text{ km}^2$  were considered, because they have been shown to have lower uncertainties than counting areas  $<100 \text{ km}^2$  (Warner et al., 2015).

Three morphological crater classes were selected to highlight potential multiple resurfacing events following a similar procedure described in Hartmann and Daubar (2017): “fresh”, “modified”, and completely “filled craters”. “Fresh craters” were defined as craters with an undegraded rim and bowl-shaped dark interior that does not show any evidence of sublimation (Figure S2a). “Modified craters” represent craters that have a smoothed rim and shallow interior (no clear bowl shape) due to mass wasting or some infilling (Figure S2b). Modified craters were distinguished from non-impact pits formed by sublimation by their regular, circular outline and bowl shape. “Filled craters” contain deposits completely filling the crater interior up to their irregular and low-relief rim (Figure S2c).

Then we follow a “three-count” procedure. Firstly, only the fresh craters were counted to estimate the date of last disturbance of the surface (i.e., pristine craters postdate the last resurfacing event). Secondly, for comparison, a second count that included the fresh and modified craters was performed to estimate the total survival time of the oldest, most modified craters that could correspond to another major resurfacing event. Finally, we counted separately the filled craters to give an estimation of the formation age of the buried underlying surface (Platz et al., 2013). This procedure gives insight into the different resurfacing and deposition that may have occurred and brackets a lower and upper temporal limit.

The three crater size-frequency distributions were plotted with Craterstats 2.1 software using a differential plot, similar to the incremental presentation used by Hartmann (Hartmann &

Neukum, 2001). The difference between the two plotting types is that the differential plot is normalized according to the choice of bin-width such that the plot is independent of the binning (aside from statistical effects; Michael, 2013). We chose a 10/decade binning to have a compromise between accuracy and smoothing of the data. The crater retention ages were calculated with Craterstats by visually selecting the portion of the crater size-frequency distributions that is parallel to some isochrons between steps of the distribution. The “Mars, Hartmann 2004 iteration” chronology system was used with the “Mars, Michael (2013)” chronology function and the “Mars, Hartmann (2005)” production function.

### 3.3 SHARAD subsurface reflections

We analyzed 129 observations from the SHallow RADar (SHARAD) instrument that cross the Utopia Planitia swath (Figure 2o). The aim was to detect radar reflections that represent an interface between contrasting materials: air-regolith, regolith-ice, regolith-basement. In this grid-mapping survey we searched for detection interfaces and postulated on the materials that were observed. Radar profiles were interpreted in several steps to avoid artifacts and reflections from surface topography (see more details in Smith & Holt, 2015; Figure S4). Material properties are discussed in the supplementary information.

## 4. Distribution of landforms in Utopia Planitia

### 4.1 Mantled deposits

Mantled deposits were defined as areas that appear subdued by a deposit over several kilometers. An example of this mantling would be filled impact craters (Figure 3a). We emphasize that these deposits do not necessarily correspond to the young LDM as other mantles were identified in multiple northern lowland locations. They are especially noticeable in a form of relief-softening over several grid cells (Figure 3b). Mantled deposits are observed at latitudes

>38°N up to the polar cap, where in turn they are covered by an extensive dune field around 75°N (Figure 2b-c).

#### 4.2 Textured terrains

At high-latitude, the surface can have a very distinctive decameter scale texture of evenly spaced knobs, referred to as “basketball terrain” (Figure 3b; Malin & Edgett, 2001). The knobs are made of accumulations of boulders (<1 m) that are locally concentrated in piles (3–5 m) and regularly spaced (20–35 m) (Levy et al., 2010a; Mellon et al., 2008). Rubble piles can be observed on top of smaller polygons (6–25 m in diameter). Formation mechanisms are unclear but may involve gravitational sorting of boulders into polygon troughs and differential sublimation (Levy et al., 2009a; Mellon et al., 2008). Textured terrains are present from ~47°N up to 78°N, where it is covered by an extensive dune field (Figure 2d).

#### 4.3 Viscous flow features

Viscous flow features (VFF) are interpreted as debris-covered/rock glaciers deformed by downslope flow, forming lineations (e.g., Carr, 2001; Head et al., 2010; Souness et al., 2012; Squyres, 1979). Different morphological types are recognized: lineated valley fill, lobate debris aprons, concentric crater fills, and glacier-like forms. In the study area, the VFF occur only as concentric crater fills (CCF) inside impact craters larger than 3 km in diameter and at latitudes of 30°–55°N (Figure 2e).

In UP, two morphological subtypes of CCF were defined: “classic” and “low-definition” (Levy et al., 2010b). The “classic” CCF are crater-interior units showing multiple concentric lineations formed by downslope flow in some part and typically display a “brain terrain” surface texture (Figure 3c). The “brain terrain” texture is interpreted to be due to differential sublimation through fractures in the debris-covered massive ice (Levy et al., 2009b; Mangold,

2003). According to our mapping, these “classic” CCF occur between 30°N and about 47°N (Figure 2e). The “low-definition” CCF have a muted undulated surface texture at broad scale (Figure 3d; Levy et al., 2010b). Smooth younger LDM overlies some portion of the CCF surface and has little evidence of deformation associated with CCF flow (Levy et al., 2009b). For latitudes >45°N, the brain terrain is progressively overlapped by this deposit until more or less complete burial at latitudes >48°N (Figure 2e).

#### 4.4 Gullies

Gullies are composed of an alcove feeding into a channel, terminating with a depositional apron (Figure 3a; Malin & Edgett, 2000). They may indicate recent liquid water (e.g., Malin et al., 2006); supplied by the melting of snow packs (Christensen, 2003) or near-surface ground ice (Costard et al., 2002); or they could result from “dry” mass movement processes (e.g., Pelletier et al., 2008; Treiman, 2003) related to seasonal CO<sub>2</sub> frost (Cedillo-Flores et al., 2011; Musselwhite et al., 2001).

In the studied area, the gullies are present from about 30°N to 67°N (Figure 2f). All identified gullies in this region are present on the inner walls of ten large impact craters (>8 km in diameter) (Figure 3a). Gullies are mostly present on the north-facing inner slopes (in seven out of ten craters), and the west-facing inner slopes of craters (in six out of ten craters).

#### 4.5 Scalloped depressions

Scalloped depressions in UP have circular to elliptical shapes with diameters ranging from tens of meters to several kilometers and depths up to several tens of meters (Figure 4; Costard & Kargel, 1995; Lefort et al., 2009; Morgenstern et al., 2007; Séjourné et al., 2011; Soare et al., 2005, 2007, 2008; Ulrich et al., 2010). They are characterized by a N-S asymmetric profile showing curvilinear steps; their pole-facing slopes are steeper (5°–20°) than their equator-

facing slopes ( $2^{\circ}$ – $5^{\circ}$ ; Lefort et al., 2009; Morgenstern et al., 2007; Séjourné et al., 2011). The scallops have been proposed to be the result of ground ice melting (Costard & Kargel, 1995; Soare et al., 2005, 2007, 2011); or sublimation of ground ice (Lefort et al., 2009; Morgenstern et al., 2007; Séjourné et al., 2011; Ulrich et al., 2010; Zanetti et al., 2010). The scalloped depressions therefore form by removal of excess ice (i.e., the amount of ice exceeding the natural pore water content in a non-frozen state) followed by collapse of the ground. Thermokarst denotes the processes and landforms associated with the volumetrically substantial loss of excess ice and resulting subsidence (French, 2007; Van Everdingen, 1998). In the paper, we will refer to “thermokarst” as degradation of excess ice either by melting or sublimation following French (2007).

The scalloped depressions are observed over the latitudes from  $\sim 38^{\circ}\text{N}$  to  $61^{\circ}\text{N}$  (Figure 2g). Four distinct zones can be distinguished based on differences in the diameter, shape, and concentration of scallops. Between the four zones, the change in morphology and concentration of scallops is relatively abrupt and is not associated with a difference in elevation or geological units (Figure 2g).

1) A local occurrence of scallops around  $58^{\circ}$ – $60^{\circ}\text{N}$  is only observed on pedestal and lobate ejecta craters. We observed small to medium-sized depressions with a diameter ranging from 0.1–5 km (Figure 4a). 2) A northern zone from  $\sim 47^{\circ}\text{N}$  to  $54^{\circ}\text{N}$ , where small-sized shallow scallops (0.1–1 km in diameter) cover the plains and crater interiors (CCF) (Figure 4b). For latitudes  $>53^{\circ}\text{N}$ , the size and number of scallops decrease sharply northward (Figure 2g). 3) A central zone between  $\sim 45^{\circ}\text{N}$  and  $47^{\circ}\text{N}$ , where numerous coalesced, nested, medium-sized depressions (1–5 km in diameter) and large-sized deep scallops (20–50 km in diameter) are observed in high concentration (Figure 4c). In the grid mapping, medium-sized depressions are

coded as "1.5" while large-sized ones are coded as "2" (Figure 2g). We did not observe any depressions around 48°N/90°E because of extensive sand and dust deposits (Figure 2g). 4) A southern zone from ~40°N to 45°N, where small and medium-sized isolated scallops (0.5–5 km in diameter) are scattered on the plains (Figure 4d). Many depressions are observed around the exhumed ejecta of large impact craters (1–10 km in diameter). The number of scallops decreases southward down to 40°N. For latitudes <43°N, the ejecta of large impact craters with diameters larger than 3 km are completely exhumed while the crater interior is still partially filled.

#### 4.6 Small-sized polygons

Different small-sized polygonal patterned ground with morphology and scale consistent with thermal contraction cracking of permafrost occur in UP (Lefort et al., 2009; Mangold, 2005; Morgenstern et al., 2007; Seibert & Kargel, 2001; Séjourné et al., 2011; Ulrich et al., 2010). According to our mapping, two types of polygons are present in two distinct zones (Figure 2h). At high-latitudes between ~53°N and 75°N, the polygons have a rectangular to hexagonal shape and are ~20–50 m in diameter (Figure 5a). They occur both on the smooth relief of the plains and on buried impact craters. We observed that in some places, the polygons were better identified when frost deposited along the troughs underlines the network on CTX images (left side of the Figure 5a). Therefore, their distribution may be more extensive than currently mapped.

At mid-latitudes between ~38°N and 50°N, the polygons are rectangular with a larger typical diameter of about 100 m (Figure 2h and 5b). Over this area, these polygons form a continuous polygonized surface on the inter-crater plains with a regular N-S orientation. In areas with significant local relief (e.g., ridge or hill), the network of polygons shows a particular orientation, parallel or orthogonal to the local slope (black arrow on Figure 5b; French, 2007).



The southern limit of their distribution (about 38°N) is very sharp (Figure 2h). The northern limit (around 48°N) is uncertain and corresponds to areas where the polygons are progressively totally degraded by scallops and covered by sand deposits.

#### 4.7 Polygon-junction pits

The rectangular polygons of 100 m in diameter show pits exclusively along their N-S troughs (referred to as polygon-junction pits) (black arrow on Figure 5b; Seibert & Kargel, 2001). The pits are elongated to circular with a diameter ranging from 10 to 150 m for a depth of 5–40 m (Costard et al., 2016; Morgenstern et al., 2007; Soare et al., 2011). They were interpreted to be due to thermokarst degradation of the troughs either by sublimation, or surface or underground melting of ground ice (Costard et al., 2016; Morgenstern et al., 2007). They often cross-cut or incise the scalloped depressions suggesting a concomitant or subsequent formation (Lefort et al., 2009). The pit distribution and the polygon distribution overlap (38°–47°N; Figure 2i).

#### 4.8 Km-scale polygons

In southern UP, polygons of 5–20 km in diameter are irregularly shaped and occasionally not completely closed (Pechmann, 1980; Hiesinger & Head, 2000; Buczkowski et al., 2012). They may have been formed by desiccation or compaction of previously water-rich sediments (Buczkowski & McGill, 2002; Pechmann, 1980), possibly in combination with an elastic rebound after removing water or ice from the ground (Hiesinger & Head, 2000). They occur between 36°N and 40°N (Figure 2j). We observed only wavy shallow troughs of 100–200 m in width and 10–20 km in length (Figure 5c).

#### 4.9 High-albedo mounds

In the northern plains, several circular cones or mounds have been studied. They are usually interpreted to be rootless cones or cinder cones (Frey & Jarosewich, 1982; Plescia, 1990),

pingos (Burr et al., 2009; Soare et al., 2005), or mud volcanoes (Farrand et al., 2005; Ivanov et al., 2014; McGowan, 2011; Skinner & Mazzini, 2009). The mounds are high-albedo features with a domed top and sometimes a central pit (Figure 5d). Large mounds (500 m in diameter) are scattered over the plain, while smaller mounds (50–300 m in diameter) are generally grouped in large fields (Figure 5e). In some places, lobate deposits of several kilometers in length overlap the mounds (white arrow on Figure 5e). Interestingly, some small mounds can be coalesced forming a ridge along a SW-NE direction (black arrow on Figure 5e). They are present between 30°N and 40°N with a higher concentration for a 30°–35°N band (Figure 2k).

#### 4.10 Thumbprint terrains

Thumbprint terrains consist of sets of parallel curvilinear ridges or pitted cone chains, with individual segments up to several tens of kilometers in length (Figure 5f; Kargel et al., 1995). These features occur in various locations on the northern plains. Based on their morphology and analogy with Earth, they have been interpreted as mudflows (Jöns, 1986) or mud volcanoes (Farrand et al., 2005; McGowan, 2011). Recently, thumbprint terrains near Arabia Terra were observed in association with runup lobate deposits and interpreted as tsunami deposits (Costard et al., 2017; Rodriguez et al., 2016). In Utopia Planitia, the thumbprint terrains are present from 30°N to 40°N within a narrow zone oriented in a NW-SE direction (Figure 2l). The curvilinear ridges range from 1 to 5 km in length with a general convex shape toward the SW (Figure 5f). The mounds forming the ridges have a dome-like shape of several hundred of meters in diameter with a central pit.

### 5. Interpretation

#### 5.1 Three assemblages of landforms in Utopia Planitia

Assemblages of landforms can be defined based on the spatial association at regional and local scale of landforms and their formation under similar climatic and geomorphologic conditions,

hence constituting a coherent landscape. The range of occurrence of the landforms can highlight trends in their distribution and potential association. The distribution of landforms is correlated with latitude and MOLA roughness at 600 m scale but less with elevation (Figure 2a, m-n). Based on these relationships, three assemblages of landforms can be distinguished.

The high-latitude assemblage is composed of mantled deposits, textured terrains, and ~20–50 m diameter polygons that are correlated between ~47°N and 78°N (Figure 2). The surface of the mantled deposits is textured at local scale (Figure 2c-d and 3b). The mantled deposits show also polygons between 61°N and 75°N. The occurrence of this landform assemblage (47°–78°N) is correlated with a low MOLA topographic roughness at 600 m scale and 2400 m scale (smoothing of topography; Figure 2m-n). The subkilometer smoothing, extending to the northern hemisphere, of topography indicates the presence of the young LDM (Head et al., 2003; Mustard et al., 2001). From the observation of polygons and textured terrains, the LDM underwent thermal contraction, cryoturbation, and differential sublimation of the upper meter (Head et al., 2003; Kreslavsky & Head 2002b; Mellon et al., 2008). From 47°N to 61°N, the surface shows numerous small-sized scallops (Figure 2g). The LDM is thought to be degraded in the mid-latitude zones (30°–60°; and referred to as a dissected mantle), whereas at high-latitudes >60°, the mantle is intact (referred to as a continuous mantle; Mustard et al., 2001).

The mid-latitude assemblage comprises scallops, 100 m diameter polygons, pits, and mantled deposits that are correlated in a zone between ~38°N to 47°N (Figure 2). They are spatially associated at small scale, where the polygons occur on the mantled deposits. The polygons are degraded by the scallops, and by the pits (Figure 4d). Their spatial association indicates an assemblage of landforms (Morgenstern et al., 2007; Soare et al., 2007, 2011). By analogy with ice-rich permafrost landscapes on Earth that have a similar assemblage of landforms

(thermokarst lake and thermal contraction polygons), this zone is interpreted to be a degraded ice-rich permafrost (Lefort et al., 2009; Morgenstern et al., 2007; Séjourné et al., 2011; Soare et al., 2007). The association of these landforms was previously shown at local scale but not at regional scale. According to Kreslavsky & Head (2000), this zone in UP has a higher MOLA roughness at 600 m scale and 2400 m scale than other places in the northern plains (Figure 2g, m-n). The strong correlation with the 600 m scale roughness is due to the presence of numerous depressions of 250-1000 m in diameter, with a higher roughness being observed in the central zone (45°–47°N) due to a higher number of nested depressions (Figure 4c). The correlation with the 2400 m scale roughness is due to the presence of kilometric depressions (also higher number for the central zone) and exhumed impact crater with ejecta.

The low-latitude assemblage is composed of km-scale polygons, high-albedo mounds, and thumbprint terrains that occur mostly between ~30°N and 38°N (Figure 2). Their close spatial association and the similarity in their morphological characteristics point to an assemblage of landforms. This assemblage occurs at an elevation between –3600 m and –4000 m. Its distribution is correlated with a medium MOLA roughness at 600 m scale and 2400 m scale (Figure 2m-n). Collectively this assemblage of landforms was interpreted as mud flows from diapirism or compaction (Buczowski et al., 2012; Ivanov et al., 2014; McGowan, 2011; Skinner & Mazzini, 2009).

Gullies and concentric crater fills have a broader distribution and do not seem to be correlated with any assemblage of landforms in particular (Figure 2). The gullies are not observed above 67°N. In the northern plains, a decrease in gully frequency for latitudes >45°N is correlated with the poleward decline of the topographic roughness at 600 m scale and consequence lack of steep slopes (Figure 2f; Dickson & Head, 2009). Therefore, the absence of gullies may be

caused by lack of appropriate craters with steep enough slopes. The CCF are no observed at high-latitudes because they are progressively overlapped by the LDM until more or less complete burial at latitudes  $>47^{\circ}\text{N}$  (Figure 2e).

## 5.2 Geomorphological units

We identified two deposits that are associated with the low- and mid-latitude assemblages of landforms forming geomorphological units. The southwestern limit of the low-latitude assemblage of landforms is correlated with several lobate deposits (Figure 6a). This limit corresponds to the southern limit of the ABvi unit (Figure 2a; Kargel et al., 1995; Ivanov et al., 2014; Jöns, 1986; Tanaka et al., 2005). This landform assemblage is thus correlated with this deposit, which covers the southeastern portion of the study area. Based on MOLA laser tracks, the thickness of the lobes was estimated to be 20–30 m (Figure 6b). The lobes each have a ridge at the front (white arrow on Figure 6b). These lobes share similar morphologic characteristics with the runup lobate deposits that were recently interpreted as tsunami deposits (Costard et al., 2017; Rodriguez et al., 2016). All of these features superimpose the VBF and collectively suggest that there was a large (1000 km diameter) reservoir of water/mud in the deepest portions of the Utopia basin (Ivanov et al., 2014; Skinner & Mazzini, 2009).

The southern limit of the mid-latitude assemblage of landforms takes the form of a polygonized deposit with clearly defined, gently sloping limits (Figure 6c). In some places, the deposit partially covers hills (dashed lines on Figure 6c, compare with the hill on Figure 5b). The deposit covers a surface that is lower in altitude (grey symbols) and partially buries some impact craters (black arrow). This underlying rough surface contains large high-albedo mounds of the previously mentioned assemblage of landforms (white arrow). The estimated thickness of the deposit is about 40–80 m, based on several MOLA laser tracks along the limit (Figure 6d). The 100 m polygons occur on the front slope of the deposit (Figure 6c). Locally, at higher

latitude (around 41°N), similar hills and an underlying rough surface were observed from the degradation of the polygonized surface by scallops (Morgenstern et al., 2007; Soare et al., 2012, 2013). Therefore, we suggest that the mid-latitude landform assemblage is correlated with this deposit that covers the low-latitude assemblage. This deposit was not recognized in previous geological maps (e.g., Tanaka et al., 2005, 2014). It is important to note that the scallops are absent along the limit of the deposit, indicating a non-degraded boundary. Additionally, the remarkable continuous morphology of the 100 m polygons over the surface indicates that they necessarily formed after the deposition of the sediment. For example, in places of local relief (ridge or hill), the network of polygons follows the local slope (Figure 5b).

In summary, a geomorphological unit is composed of the mid-latitude assemblage associated with a deposit of 80 m in thickness. This unit overlaps another geomorphological unit composed of the low-latitude assemblage associated with a lobate deposit of 30 m in thickness.

### 5.3 Crater size-frequency distribution measurements of the different geomorphological units

We measured crater size-frequency distributions (>50 m in diameter) in the zone of the three landform assemblages at 35°N, 42°N and 58°N (black squares on Figure 2a). For the high-latitude assemblage, fresh and filled craters were counted (Figure 7a). No modified craters were observed. We observe several steps in the crater size-frequency distribution of the complete crater count (black squares). Craters between ~320 m and 1 km, all of which are filled, give a best-fit age of  $1.1 \pm 0.1$  Ga. All craters <316 m (bin edge) were partially erased by resurfacing processes after 1 Ga. The fresh and filled craters with diameters between 63 and 126 m (bin edges) are aligned along the 1 Ma isochron. Fresh craters that post-date the last deposition show a similar tendency at diameters between 79 m and 160 m (bin edges), with a best-fit age of  $1.0 \pm 0.5$  Ma (red crosses; Figure 7a).

For the mid-latitude assemblages, filled, modified and fresh craters were counted on the 100 m polygonized surface (Figure 7b). The distribution of filled craters for diameter  $>500$  m is aligned along the 1 Ga isochron with a best-fit age of  $1.2 \pm 0.2$  Ga (blue circles). The count of fresh and modified craters shows a step between the 1 Ma and 10 Ma isochrons (black squares). The craters with diameters  $>100$  m are aligned along the 10 Ma isochron and give a best-fit age of  $11 \pm 1$  Ma. They correspond to the last major deposition in the zone. For craters smaller than 79 m (bin edge), a resurfacing event occurred after 11 Ma, partially removing them. The fresh crater distribution fits the 1 Ma isochron with a best-fit age of  $1.7 \pm 0.2$  Ma (red crosses; Figure 7b). Given the morphology (shallow interior and smoothed rim) of modified craters (Figure S2), this resurfacing event between 11 Ma and 1.7 Ma could be due to partial filling and/or modification by sublimation of the smallest craters.

For the low-latitude assemblage, filled and fresh craters were counted together for the crater count (Figure 7c). No modified craters were observed. The craters between 250 m and 1 km are aligned along the 1 Ga isochron with a best-fit age of  $1.0 \pm 0.1$  Ga (black squares). The crater size-frequency distribution has a step for craters  $<251$  m (bin edge) suggesting that a resurfacing event after 1 Ga partially removed them. Almost all craters appear to be partially filled in the area, so deposition of some material has clearly occurred since 1 Ga (Figure 5c).

In summary, the three crater counts of the three zones provide insight into the different resurfacing and deposition events that have occurred in UP. Using small craters and relatively small areas for dating leads to some uncertainties (e.g., Hartmann, 2005; Hartmann & Werner, 2010), but this method allows the relative comparison of the two crater retention ages. According to the filled crater counts over the three zones, we deduce a best minimum model age for the older cratered basement at  $\sim 1.1$  Ga. Over the mid-latitudes, between 1.1 Ga and 11

Ma, the relative lack of filled craters <500 m in diameter can be explained by the deposition of a layer of about 70–100 m in thickness, assuming depth-to-diameter ratios of about 1:5 to 1:7 for simple bowl-shaped primary crater (Pike, 1977). This interpretation is consistent with our previous estimation of the thickness (Figure 6c) and distribution of mantled deposits over these mid-latitudes (Figure 2c).

The fresh craters that post-date the last major resurfacing event give a similar crater retention ages (about ~1.5 Ma) at mid- and high-latitudes. This age of the high-latitude landform assemblage corresponds to the time period of LDM deposition (0.4–2.1 Ma; Head et al., 2003; Levy et al., 2010b; Mustard et al., 2001; Schon et al., 2012). This age is also in agreement with the age of non-deformed LDM covering the low-definition CCF type (Figure 3d) for latitudes >48°N (~1.5 Ma; Levy et al., 2009b). However, no noticeable difference in topography or clear examples of overlapping deposits are observed around 48°N. At the mid-latitudes, the resurfacing event at ~1.7 Ma may be related to deposition a thinner LDM or modification by sublimation. The deposition of a thinner LDM at mid-latitudes in UP is consistent with the observation of non-deformed LDM progressively overlapping the older CCF for latitudes >45°N.

#### 5.4 Ground ice content and distribution in Utopia Planitia: a particularly ice-rich area between 38°N and 47°N

Analysis of the distribution and morphotypes of ice-related features is useful to understand the cryolithology of the permafrost and its past evolution in relation to climate. The mid- and high-latitude assemblages of landforms between 38°N and 78°N are indicative of ground ice and its degradation, but the estimation of ice content is challenging. For example, the mid-latitude 100 m polygons are interpreted to have been formed into an ice-cemented permafrost with variable ice content (Levy et al., 2009a; Séjourné et al., 2011; Soare et al., 2011; Ulrich et al.,



2010, 2011). Instead of focusing on polygons that have the same morphology from plan view, we suggest that thermokarst landforms such as scallops can be used to get an estimation of the amount of excess ice, although they represent a degradational feature that can be challenging to interpret.

In our previous work, we classified the depressions from small-sized to large-sized scallops to highlight their morphological development with increasing sublimation of ground ice (Séjourné et al., 2011). Based on these considerations and changes in the morphology of the scallops in the mid-latitude assemblage (Figure 4), we can classify the thermokarst in the three zones. 1) For latitudes between  $\sim 47^{\circ}\text{N}$  and  $54^{\circ}\text{N}$ , the occurrence of small-sized shallow scallops indicates that the thermokarst has been of relatively low intensity and shallow. 2) For latitudes between  $\sim 45^{\circ}\text{N}$  and  $47^{\circ}\text{N}$ , the presence of numerous large and medium-sized, deep, coalesced scallops and a higher number of polygon-junction pits indicates that the thermokarst has been significant and deep. 3) For latitudes between  $\sim 38^{\circ}\text{N}$  and  $45^{\circ}\text{N}$  and locally on crater ejecta at  $58^{\circ}$ – $60^{\circ}\text{N}$ , the observation of isolated medium-sized scallops and polygon-junction pits indicates that the thermokarst has been moderate.

On Earth, the morphology and distribution of thermokarst landforms depend on two main factors: thermal destabilization (cause and duration) and cryolithology (amount, structure and spatial distribution of ground ice; e.g., Jorgenson et al., 2010; Rampton, 1974). Permafrost degradation is always initiated by deepening of the active layer because of an increase in subsurface temperature (French, 2007; Soloviev, 1973). Regional causes include a long-term increase in air temperature and/or precipitation, whereas local causes encompass runoff or ponding of water, local erosion, forest fire, and/or human interaction with the environment. The ice content can vary depending on the type of ice (pore, segregated, or massive ice) and

the sedimentary characteristics of permafrost (granulometry, origin; e.g., French & Shur, 2010). Therefore, based on this comparison with Earth, we suggest that the distribution and variation of thermokarst in UP indicate a difference either in 1) sublimation or, 2) ice content of the permafrost in UP.

In UP, our first hypothesis proposes that the more severe thermokarst between  $\sim 38^{\circ}\text{N}$  and  $47^{\circ}\text{N}$  than in other areas is due to a higher sublimation of ground ice. This hypothesis is supported by the observation that the boundaries between the three thermokarst areas seem to be controlled by latitude rather than geological units or elevation; therefore, a climatic control is likely. At regional scale, sublimation can be initiated by an increase in surface temperature or a decrease in air humidity (Chevrier et al., 2008; Law & Van Dijk, 1994; Mangold, 2011). Models at this scale show ground ice in UP is currently stable at depths  $<0.1$  m for latitudes  $>60^{\circ}$ – $65^{\circ}\text{N}$  and unstable at depths  $>1$  m for latitudes  $<50^{\circ}$ – $55^{\circ}\text{N}$  (Chamberlain & Boynton, 2007; Mellon et al., 2004). The distribution of thermokarst corresponds to the latitudes of current instability of ground ice. It does not necessarily mean that the thermokarst was formed during the current obliquity period on Mars, but that the ice degradation is consistent with current conditions. With increasing obliquity, the stability of ground ice shifts toward the equator (Chamberlain & Boynton, 2007; Mellon & Jakosky, 1995;). From these models, the sublimation of ground ice should be more extensive equatorward. However, this trend is the opposite of the observed decrease in size and number of scallops for latitudes  $<45^{\circ}\text{N}$  (Figure 4d). At local scale, sublimation may also depend on surface properties (e.g., surface albedo) and climatic parameters such as the insolation (surface temperature increase) or wind speed (i.e., decrease in air humidity) or removal of an insulating layer (Van Dijk & Law, 1995). The development of scallops is thought to have been controlled by insolation (Lefort et al., 2009; Morgenstern et al., 2007; Séjourné et al., 2011). However, the distribution of the high

concentration of scallops from 45°–48°N is not correlated with slopes where enhanced insolation could have increased sublimation. Additionally, there is no evidence of enhanced eolian activity (e.g., dunes, eolian erosional grooves such as yardangs, or eolian deflation hollows) or local difference in albedo that would increase the sublimation rate.

The second hypothesis suggests that the more severe thermokarst between ~38°N and 47°N than in other areas is due initially to a higher content of excess ice. In this latitude range, the 100 m polygons show a remarkably similar morphology (shape and size) and degradation level (polygon junction pits), suggesting that the ice content of wedges was consistent over a large area. On Earth, the depth of thermokarst lakes is controlled by the amount and distribution of ground ice in the substrate (Jorgenson et al., 2010; Soloviev, 1973). Based on this comparison, the largest scallops between ~45°N and 47°N (which are up to 70 m deep) show that the ice-rich zone must be of similar thickness and of relatively high excess ice content (Dundas et al., 2015; Séjourné et al., 2012). This interpretation is in agreement with the observation of a deposit 80 m thick (Figure 6c). At latitudes <48°N in UP, the subsurface ice concentration detected by neutron and gamma-ray data is significantly lower (<4% per weight of ice) than at other latitudes (Feldman et al., 2004; Maurice et al., 2011). However, both of those methods detect hydrogen present at depths less than 1 meter, so if the ice is buried deeper, then it would not be detected.

Previous SHARAD studies suggested the presence of an ice-rich layer about 100 m thick (estimated to be 50–85% water ice by volume) in a zone (70°–90°E and 40°–48°N) in the southwestern portion of our study area (Stuurman et al., 2016). We used SHARAD radar detections for our mapping survey (blue and green tracks on Figure 2o). At mid-latitudes (39°–45°N), we found a thick deposit of material that corresponds to the ice-rich zone highlighted

by our mapping. By using the estimated thickness of 80–100 m, we determined that the material has a dielectric constant between 3.4 and 5.3 (Figure S4). These values are consistent with a material that is primarily ice (~50–85% by volume) with some lithic components. We selected a zone where no extensive sublimation occurred to better estimate the ice content, but a better estimation of the maximum thickness of the deposit would improve that range. Radar reflections are not detected in the zone of significant and deep thermokarst (numerous large and medium-sized scallops at 45°–47°N; Figure 2o). This absence is either because the physical interfaces do not provide enough dielectric contrast for SHARAD to detect them, or in many locations because the ice was sublimated with formation of scallops. It is interesting that we do not see more radar reflections at high-latitudes where the LDM is supposedly widespread and visually thick (Figure 2o). We infer either that the radar is unable to penetrate deeply enough to find the interfaces with the underlying basement or that the top and bottom of the ground ice is too close to the surface. SHARAD has a minimum first detection of ~15 m, so the ice must be at depths greater than 15 m to be detected. One other deposit that we detected is associated with the low-latitude assemblage of landforms at about 32°N, but we did not constrain its dielectric properties (Figure 2o).

MARSIS also detected ground ice at depths down to 60–80 m, probably in the form of excess ice (estimated to be 60% water ice by volume) in UP for latitudes >40°N (Mouginot et al., 2012). Therefore, all of these observations are consistent with an interpretation of a thick deposit (at least 80 m) containing excess ice (~50–85% by volume) between ~38°N and 47°N. Presence of ground ice at depth is possible at those latitudes if covered by a dry insulating layer (Bramson et al., 2017; Chamberlain & Boynton, 2007; Mellon et al., 2004).

For latitudes  $>47^{\circ}\text{N}$ , we suggest that the shallow thermokarst may indicate either a smaller content of excess ice of this mid-latitude deposit or, a more recent and less extensive degradation of the LDM than at mid-latitudes. The deposition of a thinner LDM at these mid-latitudes ( $>47^{\circ}\text{N}$  from the observation of low-definition CCF, Figure 3d) in UP may have prevented extensive sublimation. Textured terrains are thought to indicate debris-covered, ice-rich LDM (Kreslavsky & Head 2002b; Mellon et al. 2008). In this context, the occurrence of scallops only on pedestal and lobate ejecta craters around  $58^{\circ}$ – $60^{\circ}\text{N}$  can be interpreted as the excavation of deep ground ice by impact. It is unclear whether the impacts have excavated the ground ice from the LDM, the deeper ice-rich deposit, or both.

## **6. Discussion: Utopia Planitia, a case study of multiple depositions of ice-rich sediments over the Late Amazonian**

Our results show that UP is covered by an ice rich deposit (40–80 m in thickness) of about 1.1 Ma ages from  $\sim 38^{\circ}\text{N}$  to  $47^{\circ}\text{N}$  which, in turn, is covered by the ice-rich LDM of  $\sim 1.5$  Ma age for latitudes  $>47^{\circ}\text{N}$ . The origin of this mid-latitude ice-rich deposit can be tied to obliquity-driven climate changes like the young LDM. The crater retention age indicates a formation during a period of high and moderate obliquity ( $35^{\circ}$  to  $45^{\circ}$ ; Laskar et al., 2004). Formation related to mid-latitude glacial landsystems along the dichotomy, which is also thought to have formed during moderate ( $\sim 35^{\circ}$ ) obliquity periods (Head et al., 2006; Madeleine et al., 2009), has been previously suggested (Séjourné et al., 2012; Soare et al., 2012; Stuurman et al., 2016). The crater retention ages of these glacial features range between  $\sim 60$  Ma and 1 Ga (Berman et al., 2015; Head et al., 2006; Mangold, 2003; Morgan et al., 2009), and the associated CCF in UP are estimated to be  $\sim 60$ – $300$  Ma in age (Levy et al., 2009b, 2010b). The difference in crater retention age between these features and the youngest deposition in UP dismisses a related origin. Given the 1.1 Ga age of the older cratered basement, which is a lower bracket on deposition in UP, a related origin cannot be ruled out for older, deeper deposit.

We suggest that this mid-latitude ice-rich deposit has the geological characteristics of a stratified sediment, not a debris-covered ice-sheet such as the LDM. The scallops reveal a complex stratification of the deposit whose origin can be fluvial or eolian (Séjourné et al., 2012). The relative smoothing of topography at mid-latitudes (Morgenstern et al., 2007) and absence of layers on crater walls and flow-related stratigraphic landforms support an eolian origin (Soare et al., 2015).

Finally, the surface material in UP may be composed of different sedimentary deposits with various excess ice contents and ages, similar to the ice-rich sedimentary permafrost (Yedoma deposits) on Earth. Underneath the mid-latitude ice-rich sediments, the VBF may also contain ice (Clifford & Parker, 2001; Mouginit et al., 2012). Formation of the thick ice-rich Yedoma deposits in the Arctic resulted from the deposition of sediments of different origins (alluvial, eolian, lacustrine) in a stable non-glaciated region under cold-climate conditions over a long period of time (French, 2007). UP appears to be a region of combined depositions of sediment and continuous cold climatic conditions that led to a complex distribution of ground ice due to morpho-climatic legacies (marine, fluvial, glacial, or eolian deposition) of the region.

## **7. Conclusion**

Our mapping highlighted three assemblages of landforms based on their association at regional and local scale as well as correlation with latitude and topographic roughness at 600 m scale:

- 1) A high-latitude assemblage composed of mantled deposits, textured terrains, small-sized scallops, and 20–50 diameter polygons between  $\sim 47^{\circ}\text{N}$  and  $78^{\circ}\text{N}$ . This assemblage is related to the ice-rich, debris-covered, latitude-dependent mantle and about 1.5 Ma in age.

2) A mid-latitude assemblage including scallops, 100 m diameter polygons, polygon-junction pits, and mantled deposit that is present from  $\sim 38^{\circ}\text{N}$  to  $47^{\circ}\text{N}$ . This assemblage and SHARAD detection provide evidence for an ice-rich stratified sedimentary permafrost of 80 m in thickness containing excess ice ( $\sim 50\text{--}85\%$  by volume) and about 11 Ma in age.

3) A low-latitude assemblage formed of km-scale polygons, high-albedo mounds, and thumbprint terrains that occurs between  $\sim 30^{\circ}\text{N}$  and  $38^{\circ}\text{N}$ . This assemblage is associated with a lobate deposit of 30 m in thickness with an age of 1.1 Ga.

Utopia Planitia appears to be a region of combined deposition of sediment and continuous cold climatic conditions that led to the formation of thick, stratified, ice rich deposits, many of which are partially preserved. Future work will focus on extending the grid mapping longitudinally in Utopia Planitia.

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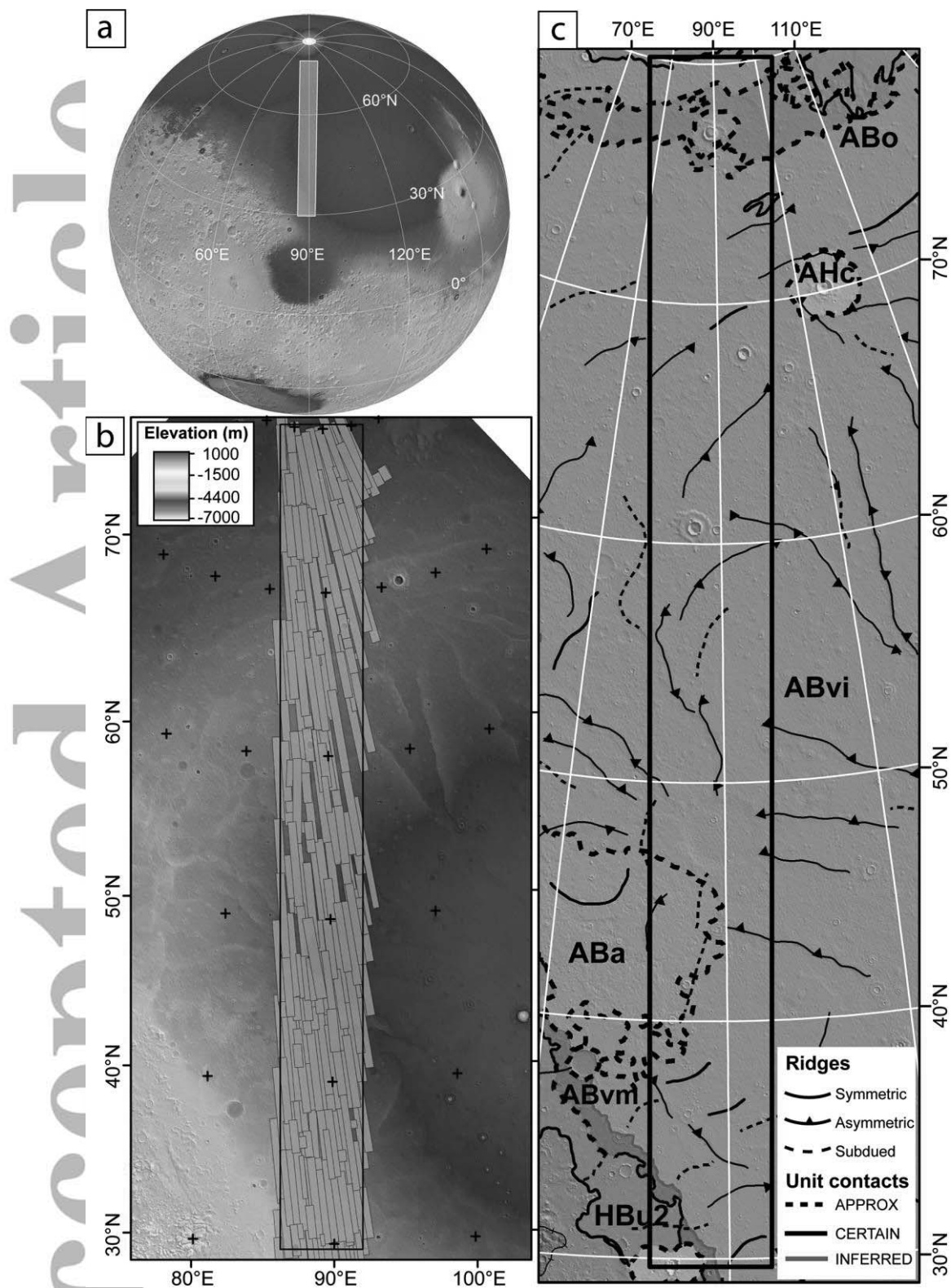
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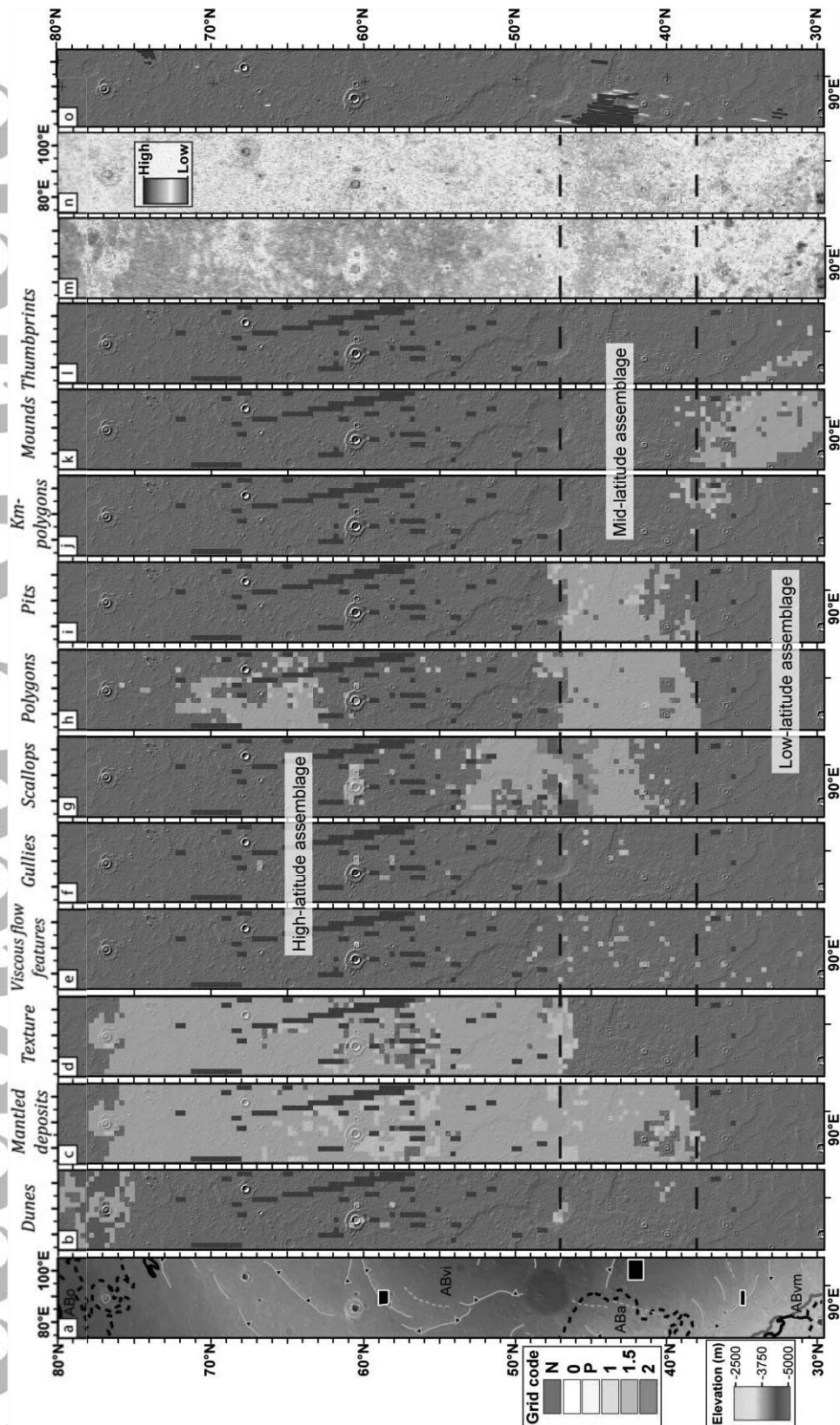
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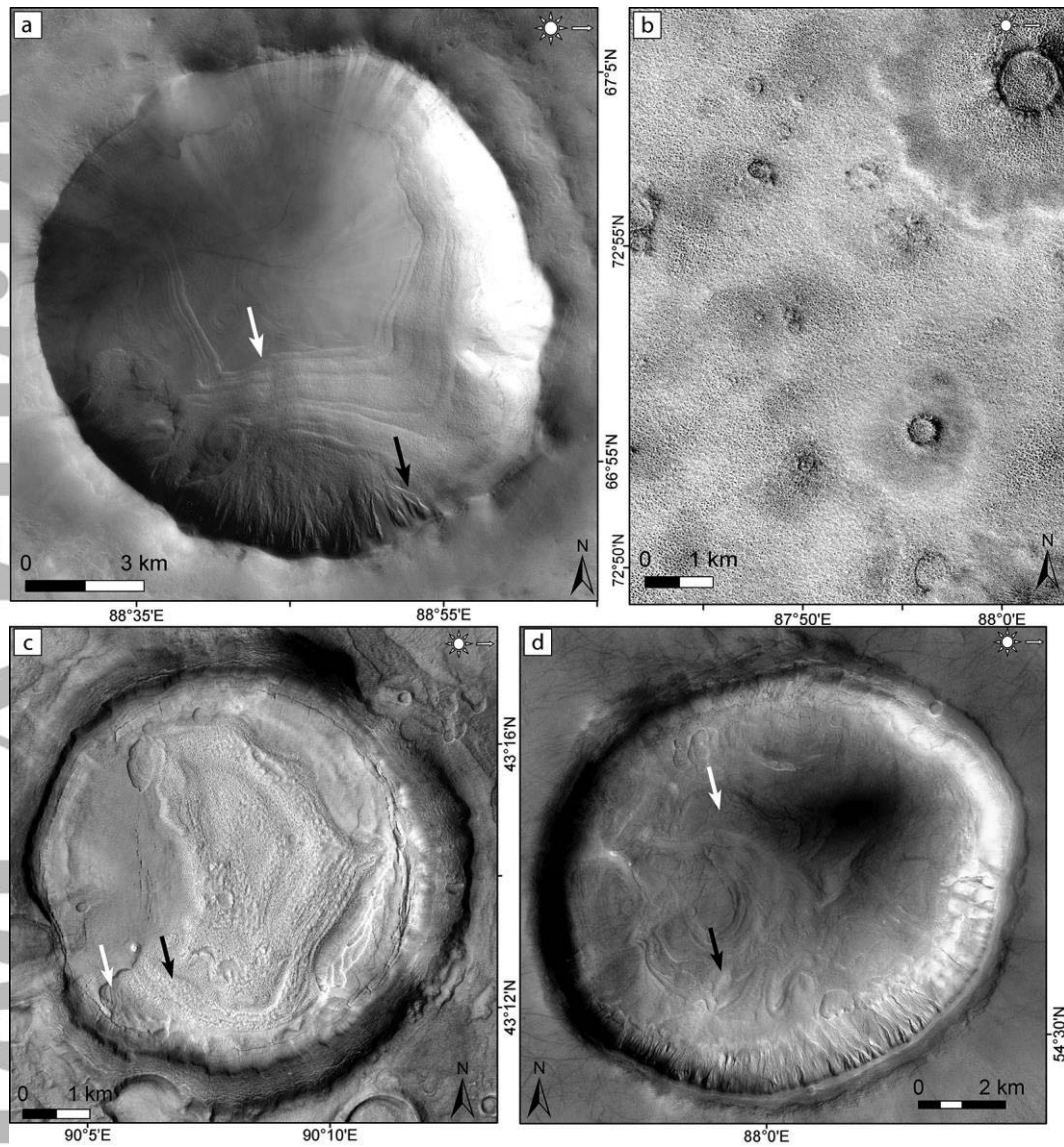
**Figure 1.** Grid mapping in western Utopia Planitia. a) Localization of the strip in the northern plains (Mars Orbiter Laser Altimeter (MOLA) digital elevation background). b) CTX image footprints used in this study overlaying MOLA map. c) Study area over the geological map of

the northern plains from Tanaka et al. (2005). ABvi: Vastitas Borealis interior unit; ABvm: Vastitas Borealis marginal unit; HBU2: Utopia Planitia 2 unit; ABa: Astapus Colles unit; ABo: Olympia Undae unit; AHc: Crater unit. See more details in the text.



**Figure 2.** Three assemblages of landforms can be distinguished in Utopia Planitia based on their associations at regional and local scale as well as correlation with latitude and topographic roughness at 600 m scale. a) MOLA topographic map with geological units of Tanaka et al.

(2005). See more details in the text and in Figure 1. Crater count areas for geomorphological units (black squares). b) to l) Distribution of landforms within the grid mapping with a MOLA hillshade background. The landforms were recorded in each grid square (20 km across) as either: “absent 0”, “possible P”, “present 1”, “extensive 1.5” or “dominant 2”. N means an absence of CTX images. MOLA topographic surface roughness represented at m) 600 m and n) 2400 m scales, modified from Kreslavsky & Head (2000). o) SHARAD subsurface reflections (blue) and possible reflections (green) associated with ice-regolith or regolith-basement interfaces.

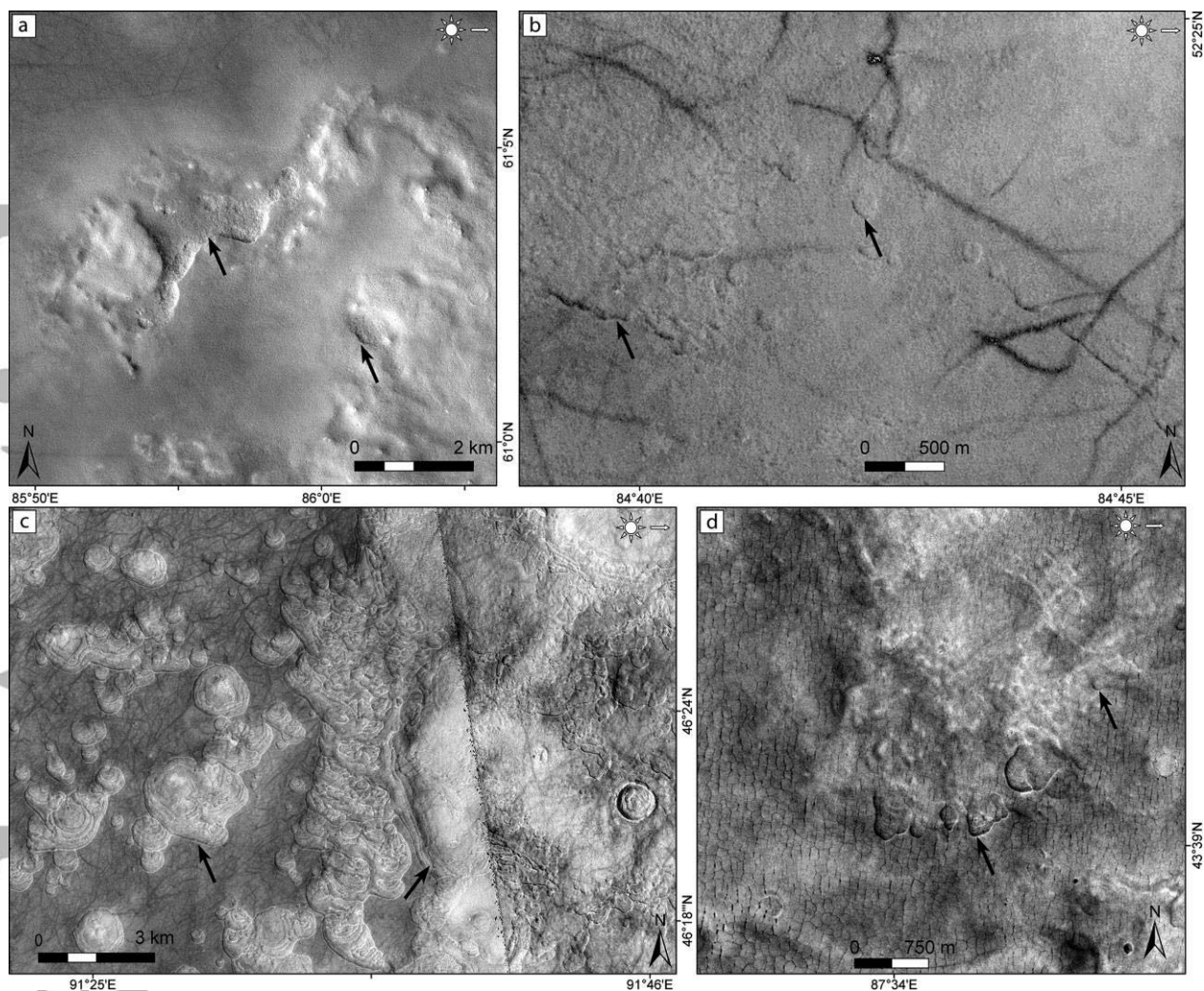


**Figure 3.** Ice-related landforms smoothing the topography. a) Mantled deposits with layers (white arrow) covering the interior of an impact crater. Gullies occur on the north-facing slopes eroding the mantled deposits (black arrow). CTX F02\_036670\_2450\_XN\_65N270W. b) The basketball terrain type consists of evenly spaced knobs that are ubiquitous over the smoothed relief of mantled deposits. G01\_018763\_2538\_XN\_73N273W. c) “Classic” concentric crater fills (CCF) with the typical brain terrain texture, concentric deformation (black arrow), and local degradation by scallops (white arrow). B17\_016126\_2348\_XN\_54N272W. d) “Low-definition” CCF instead have a muted, undulated surface texture at broad scale (black arrow).

Smooth latitude-dependent mantle overlies some portion of CCF surfaces (white arrow).

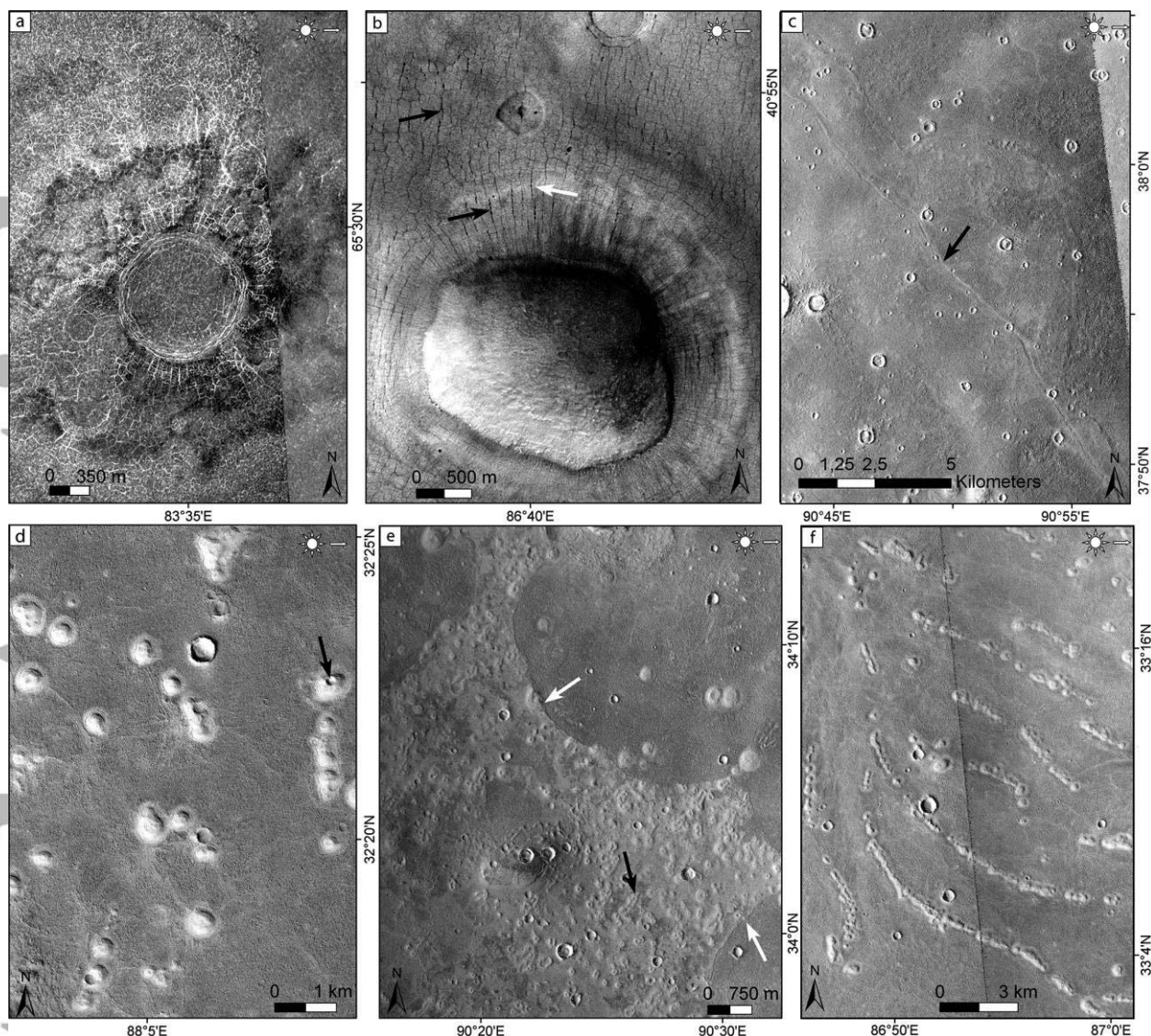
P13\_006184\_2233\_XN\_43N269W.

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**Figure 4.** Four zones are distinguished based on the morphologies and the concentration of scalloped depressions. a) Medium-sized (0.1–5 km in diameter, coded as 1 in the grid) scallops (black arrow) on ejecta of a lobate craters. CTX B01\_010179\_2408\_XN\_60N274W b) A northern zone with small-sized (0.1–1 km in diameter, coded as 1) shallow scallops (black arrow). G23\_027163\_2305\_XN\_50N274W. c) A central zone with numerous medium-sized scallops (1–5 km in diameter, coded as 1.5 in the grid) and large-sized deep ones (20–50 km in diameter, coded as 2). P19\_008478\_2246\_XI\_44N268W. d) A southern zone with scattered medium-sized (0.5–5 km in diameter, coded as 1 in the grid) single scallops (black arrow) that are scattered on the plains <43°N. B18\_016759\_2236\_XN43N272W.

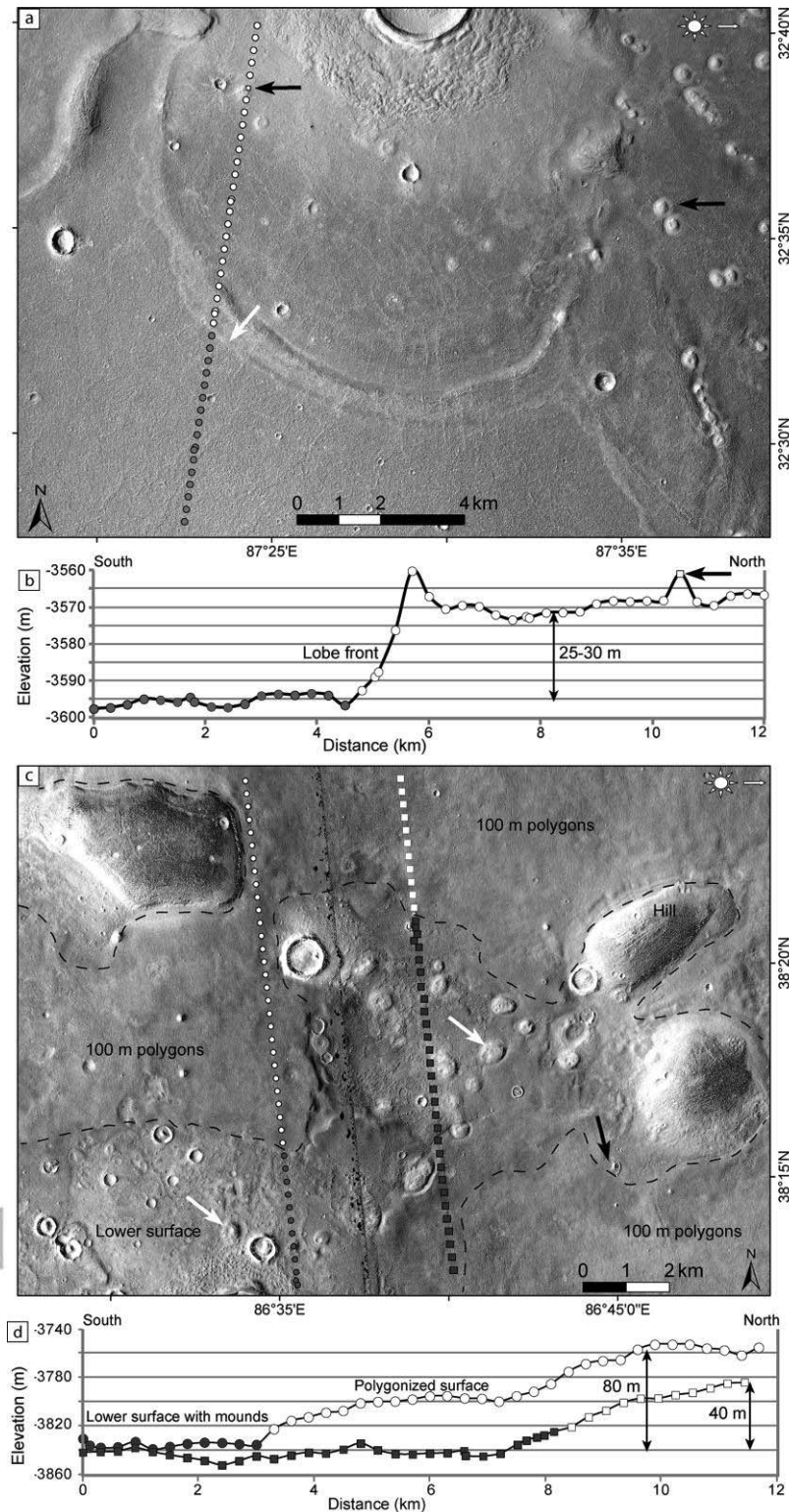




**Figure 5.** Polygons and mounds in UP. a) High-latitude polygons of 20–50 m in diameter occur on the plains and on filled impact craters. CTX D18\_034112\_2459\_XI\_65N275W/D17\_033967\_2459\_XI\_65N277W. b) Mid-latitude rectangular polygons of 100 m in diameter have a N-S elongation. In areas of local relief, the long axis of polygons is oriented along the local slope (white arrow). Semicircular pits (black arrow) are only located along the troughs of these polygons. P16\_007384\_2216\_XN\_41N273W. c) Km-scale polygons are only observed in a form of wavy shallow troughs of 10–20 km in length (black arrows). F01\_036235\_2161\_XN\_36N269W. d) High-albedo mounds composed of large pitted domes (500 m in diameter; black arrow) and smaller mounds (50–300 m in diameter) generally concentrated in large fields. They can be



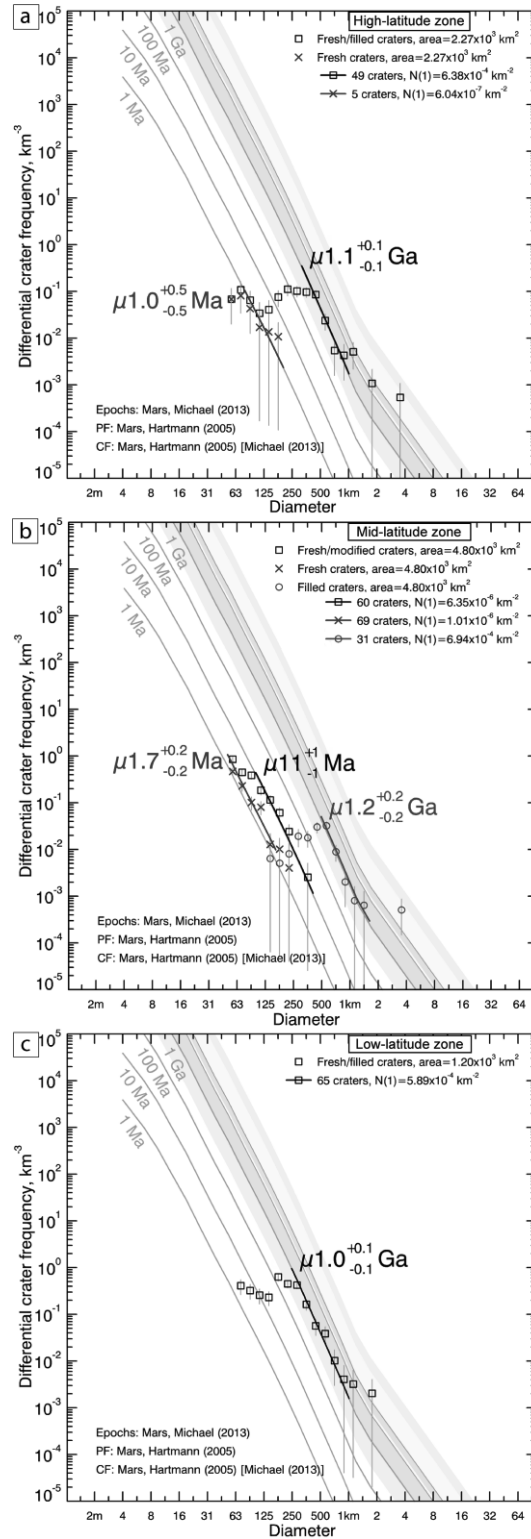
coalesced and aligned along a SW-NE direction (black arrow). Lobate deposits (white arrow) locally overlap some mounds. P12\_005815\_2128\_XI\_32N272W and B17\_016126\_2126\_XI\_32N269W. e) Thumbprint terrains form parallel and curvilinear ridges composed of aligned coalesced mounds or chains of individual ones. The curvilinear ridges range from 1 to 5 km in length with a convex shape indicating a general SW direction. B05\_011656\_2137\_XI\_33N273W.



**Figure 6.** Deposits associated with the low- and mid-latitude assemblages of landforms. a) Lobate deposit associated with the low-latitude assemblage. Large high-albedo pitted (black arrows) occur on the lobate deposit. The deposit has a ridge at the front (white arrow). b) The thickness of the deposit is about 25 m (white circles). The height of one mound is estimated to

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be about 10 m (black arrow). Data from MOLA laser shots. CTX D18\_034099\_2111\_XI\_31N272W. c) Deposit associated with the mid-latitude assemblage. The deposit that has 100 m polygons (dashed lines), partially covers some hills, high albedo mounds (white arrow), and impact craters (black arrow). d) The deposit (white symbols) is 40 and 80 m in thickness along the two shot tracks shown. Data from MOLA laser shots. CTX D21\_035378\_2184\_XN\_38N273.



**Figure 7.** Crater size-frequency distributions of the three assemblages of landforms. a) The high-latitude assemblage where textured terrains occur at  $58^\circ\text{N}$ . b) The mid-latitude assemblage where a continuous and undegraded 100 m polygonized surface is present at  $42^\circ\text{N}$ . c) The low-latitude assemblage where high-albedo mounds occur at  $35^\circ\text{N}$ .