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Possible ice-wedge polygonisation in Utopia Planitia, Mars

and

its latitudinal gradient of distribution

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ABSTRACT

On Earth, *ice complexes* are commonplace landscapes amidst the continuous permafrost of coastal or near-coastal plains in the Arctic. Formed by the freeze-thaw cycling of water, ice complex features include: hummocky (thermokarstic) terrain, inflated or deflated by the presence of absence of excess ice; thermokarst lakes (i.e. excess ice that has thawed and pooled); alases (i.e. thermokarst basins emptied of water); and, ice-wedge polygons, often characterized by raised (ice-aggraded) or lowered (ice-degraded) margins relative to the polygon centres.

The origin and development of these complexes is rooted in inter-or intra-glacial pulses of temperature that engender widespread thaw, meltwater distribution and migration through the soil column (sometimes to decametres of depth), and the freeze-thaw cycling of the meltwater.

The possible existence of ice-rich terrain on Mars revised by the freeze-thaw cycling of water dates back to the grainy Mariner-mission photographs of the 1960s and 1970s. However, absent of regolith samples from areas where this terrain is hypothesised, attempts to validate the ice-rich hypothesis often have ended abruptly, either with spectrometric inferences of water-equivalent hydrogen to one metre or so of depth or with “looks-like”, therefore “must-be” analogies derived of Earth-based ice-complexes.

In the case of small-sized Martian polygons with low- and high-centres, the similarities of form between ice and sand-wedge polygons on Earth has equivocated the reach of ice-wedge hypotheses on Mars.

Here, we show that:

1) The plains’ terrain of our study region in Utopia Planitia (40-50° N; 100-125° E) displays a statistically-significant and positive (linear) correlation between the ratio of low-centred to high-centred polygons (*lcps* vs *hcps*) and a poleward latitude of distribution.

2) This linear correlation would be expected, in as much as ground-ice stability increases with latitude, were the shoulders of higher-latitude *lcps* underlain by (aggraded) ice-wedges and those of lower-latitude *hcps* underlain by (degraded) ice-wedges.

3) The change of polygon morphology with latitude would not be expected were the *lcps* and *hcps* underlain by sand wedges, in as much as ground-ice stability is unrelated to their aggradation or degradation.

4) Crater counts of the polygonised terrain indicate that it is less youthful than previous studies have suggested, perhaps by an order of magnitude. This attenuates the possible inconsistency between the more temperate boundary-conditions required by the formation of ice-wedge polygons and the current constraints of extreme aridity, low temperatures and low atmospheric pressure.

1. Introduction

On Earth, *ice complexes* (also referenced as ice-rich terrain) at or near the surface or to decametres of depth are ubiquitous within the permafrost of Arctic coastal-plains, i.e. the Tuktoyaktuk region of northern Canada (e.g. Rampton and Bouchard, 1975; Rampton, 1988; Murton, 2005) and the Laptev peninsula of northeastern Siberia (e.g. Schirrmeister et al. 2002, 2017; Grosse et al. 2007; Morgenstern et al. 2013; Strauss et al. 2017). Some examples of this terrain are:

a) Hummocky thermokarst (ballooned by aggraded excess-ice and deflated where the excess ice has been lost by evaporation, drainage or sublimation).

b) Thermokarst lakes, i.e. pooled meltwater resulting from the thaw of excess ice.

c) Alases (thermokarst-lake basins depleted of water).

d) Ice-wedge polygons, typically ≤25 m in diameter. Morphologically, they can be distinguished by three principal archetypes. **1.** Low-centres relative to high margins/shoulders where the ice wedges have/are aggraded/aggrading; **2.** high-centres relative to low margins/troughs where the ice wedges have/are degraded/degrading; or, **3.** centres and margins that exhibit no discernable elevation difference, indicative of ice-wedge incipience or of a more mature and transitional stage between aggradation and degradation.

Pulses of ice-complex formation have been associated with inter-and intra-glacial mean-temperature rises and falls that reach back through the Holocene epoch (e.g. Rampton and Bouchard, 1975; Rampton, 1988; Murton et al. 2005; Grosse et al 2007; Morgenstern et al. 2013; Schirrmeister et al. 2013), to the Wisconsinan glacial stage (e.g. Rampton and Bouchard, 1975; Rampton, 1988; Murton et al. 2005; Grosse et al. 2007; Morgenstern et al. 2013; Schirrmeister et al. 2013) and, even into the earlier periods of the Pleistocene Epoch (e.g. Rampton, 1988; Kanevskiy et al. 2017).

Ice-wedge polygon networks are current as well as paleo-markers of temperature variances. The nascent conditions required for the ice-wedges to form are continuous permafrost, thermal-contraction cracking, meltwater, hoarfrost (to a much lesser degree), and seasonal/annual (mean) temperature fluctuations above and below 0° C (e.g. Lachenbruch, 1962). Where low-centred (aggraded) and high-centred (degraded) polygons are closely set, this marks a system actively in thermal disequilibrium and transition (Washburn, 1973; French 2007; also, Rampton and Bouchard, 1975; Rampton, 1988; Schirrmeister et al. 2002; Grosse et al. 2007; Morgenstern et al. 2013).

Mars-based *lcps* and *hcps* that are similar in shape and size to those observed on Earth and, on occasion have been reported widely in the literature (e.g. Lefort et al. 2009; Levy et al. 2009a, 2010; Séjourné et al. 2011; Soare et al. 2014, 2018). Questions concerning whether ice or sand underlies the margins of the Martian *lcps* and *hcps* have been and remain open-ended, for two main reasons (e.g. Lefort et al. 2009; Levy et al. 2009a, 2010; Séjourné et al. 2011; Soare et al. 2014, 2018).

First, and based solely on the plan-based observation of *lcps* or *hcps* on Earth and on Mars, the identification or discrimination of sub-marginal materials is not possible. This is due to the similarity of form and scale between ice-and sand-wedge polygons when viewed planimetrically, from above (Levy et al. 2009a, 2010; Hallet et al. 2011; Hauber et al. 2011; Soare et al. 2014, 2018).

Second, metres-wide ice wedges are not unusual amidst northern ice-complexes on Earth. Their development is incremental and episodically iterative, ranging through hundreds of seasonal/annual freeze-thaw cycles and meltwater fills (Lachenbruch, 1962). A similar or an even greater number of iterations would be required on Mars for similarly scaled ice-wedges to evolve. Current (mean) water-vapour pressures at or near the surface are as much as two orders of magnitude lower than on Earth (e.g. Renno et al. 2013; Trokhimovskiy et al. 2015) and are inconsistent with the freeze-thaw cycling of water at a region-wide scale. There are areas, however, where the triple-point of water is exceeded (e.g. Haberle et al. 2001; Hecht, 2002). What is unclear is whether the temporal extent of this meta-stability, current, past or long past, could have supported the iterative requirements of ice-wedge formation.

In this article we plot, describe, discuss and statistically evaluate the distribution of *lcps* and *hcps* in Utopia Planitia [*UP*] (40-50° N; 100-125° E) **(Fig. 1)**. The latter is one of the great

northern plains of Mars. We assume that here in our study region, as elsewhere on Mars, the stability of ground ice increases with its poleward latitude. This is because (mean) annual surface and near-surface temperatures tend to remain below the atmospheric frost-point with poleward proximity, mitigating ice loss by sublimation (e.g. Clifford, 1993; Mellon and Jakosky, 1993, 1995) Consequently, ice-aggraded polygons should show a higher density of distribution with poleward proximity than ice-degraded polygons and *vice versa*.

2. Methods

All of the High Resolution Imaging Science Experiment (*HiRISE)* images (*N*=135) in our study region (40-50° N; 110-125° E) were surveyed for the presence of low- or high-centered polygons **(Fig. 1)**. To ensure the consistency of our observations and interpretations, multiple co-authors evaluated the images independently and our data base (**Table 1**) integrates these evaluations.

The survey itself was based on a grid-mapping approach (Ramsdale et al. 2017). Eleven images were too blurry to be evaluated; sixteen images were not surveyed because they were fully overlapped by images with higher resolution, or with equal resolution but greater spatial extent. These images have been excluded from the study and are not reported in **Table 1**. The remaining sample comprised 108 images, divided into two groups: **1)** images whose resolution and/or quality support the unambiguous identification of *lcps* and/or *hcps*, if and when they are present (67 images, of which 41 have polygons); and, **2)** images whose resolution and/or quality are of a lesser order, making the identification of *lcps* and *hcps* more difficult (41 images, of which 35 have polygons (**Table 1**).

The *HiRISE* image-footprints were gridded into reference squares of 250,000 m^2 (37,639 squares) **(Figs. 2a-b)**. Some squares were not fully within the *HiRISE* footprints. We excluded

four-thousand two-hundred and thirty-seven of those squares from our data base since they comprised less than half the area of a reference square, i.e. 125,000 m². A further, 111 squares were discounted because they overlapped with data gaps in the *HiRISE* image. As such, 33,291 squares were evaluated.

If five or more candidate *lcps* or *hcps* were observed within each square they were recorded. For grid squares where polygonised terrain was observed but the terrain showed neither *lcps* nor *hcps* were observed, grid squares also were noted. Where fewer than five polygons were observed of any given type, they were not recorded for that grid square; highly localized if not stochastic boundary-conditions could have been responsible for the development of these small samples and would bias our evaluation of the larger population were they to be incorporated in our statistics. We are aware of and acknowledge that our five-unit (count/no count) threshold is arbitrary, but no more so than any other small unit-threshold that could have been used to differentiate meaningful samples from meaningless ones.

For each *HiRISE* image **(Table 1)** the number grid-squares containing *lcp* data-points was divided by the number of grid-squares containing *hcp* data-points to calculate the *lcp*/*hcp* ratio of each image. Also, each scanned grid-square was related to: **a)** its centre latitude; **b)** the presence/absence of *lcps* or *hcps*; and, **c)** one of two terrain types (**Type 1** and **Type 2**) by the location of their centre point. **Type 1** comprises the whole study region exclusive of **Type 2**. The latter comprises the walls, rims and interiors of craters ≥ 2 km (in diameter). These outputs were used to evaluate the possible correlation between polygon type (i.e. the *lcp*/*hcp* ratio) and latitude/terrain type using regression analysis.

Crater counts were conducted to derive information about the age of the polygonized terrain. *HiRISE* image ESP_026450_2270 (center longitude 110.880° E, center latitude 46.823°

N) was selected within the study region as it contained polygons throughout the image; counts were conducted on the entire image covering 150 km^2 at 0.5 m pix^{-1} resolution using the *Cratertools* plug-in for the *ERSI ArcGIS* to measure crater diameters (Kneissl et al. 2011). The crater-size frequency distribution (*CSFD*) was compared with modeled crater-retention age isochrons from Hartmann (2005) using *Craterstats* (Michael and Neukum, 2010; Michael et al. 2016).

3. Observations

Small-sized (~⩽25 m in diameter) and clastically non-sorted polygons are ubiquitous in our study region. Typically, the polygons are hexagonal and enclosed by metre-scale margins. The margins exhibit: **a)** slightly-elevated (metre-high) shoulders (low-centred polygons, *lcps*) **(Figs. 3a, c)**; **b)** depressed troughs (high-centred polygons, *hcps*), relative to the polygon centres **(Figs. 3b-c)**; or, **c)** variances of elevation that are small or non-existent **(Fig. 3c)**, at the highest resolution available (~25 cm/pixel, *HiRISE* image).

Here, as elsewhere in Utopia Planitia (e.g. Costard et al. 1995; Seibert and Kargel, 2001; Morgenstern et al. 2007; Soare et al. 2007, 2008; Lefort et al. 2009; Levy et al. 2009a; Ulrich et al. 2010; Séjourné et al. 2011, 2012), the small-sized polygons show distribution at or near the surface that is wide-ranging, often covering many square kilometres **(Figs. 2a-b)**, and varied: **1)** within rimless and sometimes scalloped depressions **(Figs. 3a-c)**; **2)** amidst inter-(impact) crater plains **(Figs. 2b, 3d-e)**; and, **3)** within impact craters **(Figs. 3f-g)**. Frequently, the *lcps* are proximal to the depression margins (e.g. Lefort et al. 2009; Séjourné et al. 2011). The *hcps*, by contrast, display no similar bias.

The depressions exhibit a variety of planforms: circular (**Fig. 4a**); sub-circular or elongated (**Fig. 4b**); and, scalloped on the margins (**Fig. 4c**). Their distribution is no less varied, being

observed individually (**Fig. 4a**), clustered or coalesced (**Fig. 4c**), or in wave-like forms (**Fig. 4d**). Orientation, defined by the loss of elevation relative to the surface datum, is longitudinal and equatorward (e.g. Morgenstern et al. 2007; Lefort et al. 2009; Séjourné et al. 2011, 2012). Depths range from metres to decametres (Morgenstern et al. 2007; Lefort et al. 2009; Séjourné et al. 2011, 2012). Inward-oriented and roughly equatorial tiers or terraces often demarcate the changes of elevation and are commonplace amidst the scalloped depressions (**Fig. 4c**).

4. Ice-complex constituents on Earth

4.1 *Thermal-contraction cracking and ice wedge polygons*

In low-lying and coastal regions of ice-rich permafrost such as the Tuktoyaktuk Coastlands small-sized (generally ≤25 m in diameter), clastically non-sorted, ice-wedge polygons are observed widely (Mackay, 1974; Rampton and Bouchard, 1975; Rampton, 1988; French, 2007). Here, mean monthly-temperatures (1971-2000) range from -26.3^{o} to +10.9^{o} C, and the mean annual-precipitation, inclusive of rain and snowfall, is 139.3 cm (water equivalent) (Environment Canada, 2010).

The polygons are the product of tensile stresses in frozen soil that arise from the sharp drop of sub-zero ^{0}C temperatures (Lachenbruch, 1962). This fracturing, or thermal-contraction cracking, initially forms shallow, narrow and vertical veins **(Fig. 5a)**. The type of fill depends on the ambient boundary-conditions and the availability of: **1)** meltwater derived of thawed snow or ice; **2)** winter hoarfrost; or, **3)** windblown sand, mineral-soil, or a mixture of the two (e.g. de Leffingwell, 1915; Péwé, 1959; Lachenbruch 1962; Washburn, 1973; Sletten et al. 2003; Hallet et al. 2011).

The diurnal and seasonal iteration of cracking and filling may grow the sub-centimetre to centimetre- scale veins into wedges with metre-scale depth and width (Lachenbruch, 1962). As the cracks intercept one another they form individual polygons and, eventually, consolidated polygon

networks (Washburn, 1973; Mackay, 1974; Mackay and Matthews, 1983). The latter may comprise hundreds of square kilometres of continuous or discontinuous coverage and represent hundreds of fill and crack-formation cycles (e.g. Black, 1954; Lachenbruch, 1962; Washburn, 1973; Mackay, 1974) **(Fig. 5b)**.

Thermal-contraction polygons exhibit low **(Fig. 5c)** or high centres **(Figs. 5d-e)** relative to their margins or centres, or margins that are relatively-equal in elevation **(Fig. 5b)**, depending upon their developmental stage. Ice-wedge aggradation is informed by the (seasonally) iterative infilling and freezing of meltwater at polygon margins, as well as by the active-layer expansion that moves sediment radially to the polygon margins from the centres. This uplifts the sedimentary overburdens, forms raised shoulders (Mackay, 1981; French and Guglielmin, 2000; French, 2007) and gives the polygons a distinctly low-centred appearance (Czudek and Dumek, 1970; Rampton, 1988; Hallet et al. 2011). Ice-wedge degradation follows from a sustained rise of thaw-tolerant mean temperatures. This deflates polygon-margin shoulders, perhaps sinking them below polygon-centre elevation, morphing low-centred into high-centred polygons (Czudek and Dumek, 1970; Rampton, 1988; Hallet et al. 2011). Regions or sub-regions where *lcps* and *hcps* are observed concurrently are in thermal flux (Czudek and Dumek, 1970; Rampton, 1988; Hallet et al., 2011).

Of note; some polygon centres and their icy margins may exhibit little or no difference in their relative elevations. This either means that the ice has aggraded insufficiently to form marginal shoulders or degraded insufficiently to form marginal troughs.

4.2 Sand-wedge, composite and sublimation-enhanced polygons

Sand-wedge polygons with low and high centres are observed in relatively-temperate and ice-rich landscapes where ice-wedge polygons are present (e.g. Mackay and Matthews, 1983; Murton et al. 2000) **(Fig. 5f)**, as well as in the McMurdo Dry Valleys *(MDVs*), Antarctica, where

they are absent from the landscape (e.g. Péwé, 1959; Berg and Black, 1966; French and Guglielmin, 2000; Marchant et al. 2002; Bockheim et al. 2002; Sletten et al. 2003; Hallett et al. 2011). In both cases the sand-wedge polygons are paleo-markers of climatic and hydrological regimes that are drier and/or colder than those associated with the origin and development of ice-wedge polygons (Washburn, 1973; Mackay and Matthews, 1983; Murton et al., 2000; Wolfe et al. 2018). For example, contrast the annual mean-temperatures and -precipitation at the Tuktoyaktuk coastlands (discussed above) with the mean annual air-temperatures and precipitation at Beacon Valley, one of the *MDVs*. At Beacon Vally, annual mean-temperatures are ~-34° C and soil temperatures possibly do not exceed ~0° C (Marchant et al., 2002). Precipitation is meagre (<50mm of water equivalent precipitation per annum) and is comprised solely of snowfall (Bockheim et al. 2002; Levy 2015). Boundary conditions consistent with the iterative formation of ice-wedge polygons are rare, perhaps not exceeding a few days per annum.

Sand-wedge polygons originate and develop by the same thermal stresses that engender the formation of ice-wedge polygons (e.g. Berg and Black, 1966; Péwé, 1959; French and Gugliemin, 2000; Marchant et al. 2002; Sletten et al. 2003; Hallet et al. 2011). Morphologically, the rise and fall of polygon margins follows a slightly different path. The latter deform upwardly, forming (raised shoulders) *lcps*, as sand and other surface minerals accumulate by the work of wind; downward deformation, i.e. the transformation of raised shoulders into marginal troughs, by aeolian erosion, creates *hcps* (e.g. Péwé, 1959; French and Gugliemin, 2000; Marchant et al. 2002; Sletten et al. 2003; Hallet et al. 2011). Sometimes, the evolution of *hcps* is enhanced or accelerated by the sublimation of underlying glacial ice (Marchant et al. 2002).

4.3 Thermokarst and ice-rich landscapes

Thermokarst refers to landforms developed by way of freeze-thaw cycling in permafrost composed of excess ice (e.g. Taber, 1930; Penner, 1959; Hussey, 1966; Hughes, 1974; Rampton, 1973, 1988; Harris et al. 1988; Osterkamp et al. 2009). Excess ice = "…the volume of ice in the ground that exceeds the total pore volume that the ground would have under unfrozen conditions" (e.g. Harris et al. 1988; also, see Taber, 1930; Penner, 1959; Rampton and Mackay, 1971; Washburn, 1973; Murton, 1996; Rampton, 1988; French, 2007).

Excess ice comprises ice lenses, veins, wedges or larger masses of consolidated ice that are formed by ice segregation. Fine to medium-grained soils such as silts or silty clays are particularly adept at hosting segregation ice because they have relatively-small interstices (Taber, 1930; Black, 1954; Penner, 1959; Rampton and Mackay, 1971; Rampton, 1988; Sone at al. 2006; French, 2007; Abramov et al., 2008). Small interstices facilitate the migration of pore water to and the formation of discrete layers or lenses of segregation ice at the freezing front (Taber, 1930; Penner, 1959; Rampton and Mackay, 1971; Rampton, 1988; French, 2007). Were the grain-size too fine, pore-water migration would be overly constricted and ice lenses would not form; were the grain-size too coarse, pore-water migration would be overly expeditious and flush out of the system before segregation ice could form.

Where excess ice aggrades, frost heave follows (Taber, 1930; Penner, 1959; Rampton, 1988). Where excess ice thaws and degrades, the terrain subsides and settles under its own weight, until it reaches a consolidated state (Harris et al., 1988) **(Figs. 6a-b)**. If thaw-generated meltwater pools then a thermokarst pond or lake forms; were the pooled meltwater to be lost by evaporation or drainage, an emptied basin or alas would form (Washburn, 1973; Rampton, 1988; French, 2007). Were the loss of pooled water partial and episodic, inwardly-oriented benches or terraces

may develop on the alas margins; this imparts to the margins a distinctly-scalloped appearance (e.g. Brown et al. 1981; Soare et al. 2008, 2011).

Spatially, the dense or sparse distribution of heaved and subsided terrain, as well as of thermokarst lakes and alases, in ice-rich terrain delineates two things: **1)** the breadth and scale of thermal flux and disequilibrium in that region (e.g. Péwe, 1954; French, 2007; Grosse et al. 2007; Schirrmeister et al. 2013; Wetterich et al. 2014); and, **2)** a measure of the volumetric loss of ice in that region (e.g. Czudek and Demek, 1970; Rampton and Mackay, 1971; Washburn, 1973, Rampton, 1988; French, 2007). This is no less true of ice-rich landscapes where *lcps* and hcps are observed concurrently.

5. Results

5.1 LCPs/HCPs and poleward latitude **(Table 1)**

All of the possible *lcps* or *hcps*, regardless of the group of the image they are located in (i.e. **1.** images whose resolution/quality allows unambiguous identification of polygons; and, **2.** images whose lower resolution/quality make the identification of polygons more uncertain), show similar changes of morphology with their poleward latitude. As such, both datasets are combined in the following results. We identified 313 squares with *lcps* and no *hcps*, 3,325 squares with *hcps* and no *lcps*, 663 squares with both *lcps* and *hcps*, and 10,349 squares with flat polygons, generating a total of 14,650 polygonised grid-squares (**Table S1**). The study area (40-50°N, 100-125°E) comprised 23,380 grid squares classified as terrain **Type 1** and 9,911 classified as terrain **Type 2 (Table S1)**.

Overall, the ratio of low to high-centered polygons does not show any correlation with increasing latitude (R^2 ~0.0143) **(Fig. 7)**, despite the frequency of polygonised squares increasing with latitude. However, when we excluded polygons in **Type 2** terrain from the sample data, a

statistically-significant linear correlation appears between the *lcp*-to-*hcp* ratio and latitude (R^2 > 0.87, *p-value* <0.05 (**Fig. 8**). Note that **Fig. 8** and **Table S1** data include three latitudinal ranges without *lcps* (40°-<41°, 42°-<43°, 43°-<44°; *lcp/hcp* = 0), and one range without polygons (41°-<42°; *lcp/hcp* = no data).

Crater interiors (**Type 2**) host lower temperatures and slightly-higher atmospheric-pressures than the exterior terrain populated by them; these cold traps enhance the stability and preservation of volatiles (e.g. Conway et al. 2012; also, Ingersoll et al. 1992; Dickson et al. 2010). Low-centred ice-wedge polygons, formed under antecedent boundary conditions consistent with the freeze-thaw cycling of water but cold-trapped by recent or current conditions, would be able to survive and avoid degradation at much lower latitudes than similar polygons exposed to greater variances of temperature and pressure in the surrounding inter-crater terrain **(Type 1)**. Excluding the crater-based polygons in **Type 2** terrain from our regression analysis prevents our results from being biased by these geographical outliers.

Sixty-eight percent of the squares with *lcps* also have *hcps*. This increases to 83% if **Type 2** grid squares are excluded from the count and decreases to 37% when **Type 2** grid-squares alone are counted.

5.2 Surface-age estimates

Crater counts conducted on the polygonized terrain in our study region provide information about the surface age. A total of 168 craters were identified in a 150 km^2 area of the polygonised terrain in *HiRISE* image ESP_026450_2270 (**Fig. 9**). The crater-size frequency distribution (*CSFD*) of the crater population has a much shallower power-law slope than predicted by production functions (e.g. Hartmann, 2005) and thus does not represent a formation age, but rather a “crater retention age” regulated by a destruction rate that in general preferentially removes

smaller craters (e.g. Öpik 1965; Hartmann 1971; Smith et al. 2008; Kite and Mayer 2017; Williams et al., 2018; Palucis et al. 2020).

The emplacement of these terrain materials and other mid-latitude mantling deposits has been suggested to have occurred very late in martian history as a consequence of recent climate excursions based on a general paucity of craters observed in lower-resolution MOC images (e.g. Head et al., 2003; Morgenstern et al. 2007). However, more recent observations of craters on the polygonized terrains within *UP* using higher-resolution HiRISE images, suggests a more complicated history. Craters with a fresh appearance, unmodified by polygon formation, were used by Levy et al. (2009a; 2009b) to estimate a minimum age for the cessation of emplacement of mantling material ~1.5 Ma at this latitude (~45°) and *CSFDs* on polygonised terrain measured by Soare et al. (2020) suggested a more protracted history of emplacement and modification for the polygonised materials in *UP* with a modelled surface exposure age at least ~100 Ma.

The number of craters $D > 50$ m in our count area suggests the material was emplaced and exposed to impact cratering for at least 10 Myrs, possibly longer, and precludes a surface as young as the most recent obliquity variations (Laskar et al., 2004). The CSFD suggests the surface has experienced substantial degradation. This is supported by the morphologies of the craters in various states of degradation; many of them exhibit shallow hollows. Moreover, this demonstrates that the polygonised surface has undergone erosion, the likely agent responsible for the removal of many of the craters in a size-dependent manner (e.g. Hartmann, 1971; Smith et al. 2008; Williams et al. 2018).

6. A geochronology of periglaciation at the mid-latitudes of Utopia Planitia

Here, we present an integrated geochronology that comprises a baseline of generally-accepted ideas and reasoning within the community complemented by our own observations, deductions and statistical findings.

6.1 The origin and development of ice-rich terrain

Stage 1a: Atmospheric water-vapour (the Martian equivalent of winter hoar-frost on Earth) diffuses and charges the pore space of near-surface regolith (top few metres) with ice at all latitudes during one or more periods of high obliquity during the Late Amazonian Epoch (e.g. Clifford, 1993; Mellon and Jakosky, 1993, 1995; Douglas and Mellon, 2019). In concert (or not) with this diffusion-driven ice-charging of near-surface regolith, high obliquities engender enhanced periods of H_2O-ice precipitation. The latter accumulates at the surface and forms an icy mantle (e.g. Head et al. 2003; Forget et al. 2006; Madeleine et al. 2009, 2014).

Stage 1b: Sub-aerial temperatures sufficient to host the freeze-thaw cycling of the icy mantle and near-surface regolith occur, perhaps for a few hours per sol at the appropriate solar longitude (Rivera-Valentin et al. 2020) but through sufficient annual iterations for excess ice and ice-wedge polygons to form (Soare et al. 2014, 2017, 2018). The presence of brine-bearing sediments could/would have facilitated the onset and iteration of this process because salt depresses the freezing point of water (e.g. Brass 1980; Fairén et al. 2008; Rennó et al., 2009; Möhlmann and Thomsen, 2011; Martinez et al., 2013; Travis et al. 2013; Toner and Catling, 2018; Rivera-Valentin et al. 2020). The recent work of Rivera-Valentin et al. (2020) points to brines possibly being present today in our study region.

Further facilitation could have been provided, as is the case in ice-rich permafrost regions on Earth, by the presence of relatively fine-to medium-grained regolith (Soare et al., 2015; based on: Taber, 1930; Black, 1954; Penner, 1959; Rampton and Mackay, 1971; Rampton, 1988; Sone

at al. 2006; French, 2007; Abramov et al., 2008). The latter would have been delivered to our study region by aeolian (Bristow and Moller, 2018) or volcanic surface processes (Hoppington and Leverington, 2014).

Stage 2: As high obliquity/ies wanes/wane, near-surface ground ice at the lower mid-latitudes of our study region loses stability and sublimates (Mellon and Jakosky, 1993, 1995). Thermokarst or alas-like basins form where the volatile loss occurs (Morgenstern et al. 2007; Lefort et al. 2009; Ulrich et al. 2010; Séjourné et al., 2011, 2012; Mangold, 2011; Soare et al. 2014; Dundas et al. 2015; Dundas, 2017). The episodic or intermittent return to higher obliquities, much like glacial inter-stades on Earth (e.g. Rampton and Bouchard, 1975; Rampton, 1988), could/would have replenished and recharged the volatile-depleted terrain with ice. Against this backdrop, the rimless and sometimes scalloped depressions in our study region, and throughout the mid-latitudes of *UP* might be indicative of the net loss of ice-richness through the Late Amazonian Epoch to the present day. Their shape, size, depth, and location possibly reflect random differences in the distribution and depth of ground ice as well as in the reach of thermal disequilibrium (Soare et al. 2018). In this regard, the inwardly-oriented steps or tiers often observed within the depressions could be markers of an episodic loss of ice, much in the same way as tiered thermokarst basins in the Tuktoyaktuk region point to the episodic loss of water (Soare et al. 2007, 2008; 2017; also, Séjourné et al. 2011, 2012).

6.2 The origin and development of ice-wedge polygons

Stage 3: As we suggested in **Section 5.2**, thermal-contraction cracking polygonises the terrain within the depressions, perhaps 100 Mya. Additionally, we propose that thaw-derived meltwater is produced, whenever local/regional boundary conditions are fitting, with the possible assistance of near-surface brines, and maybe in conjunction with higher obliquities. The meltwater

fills the polygonised cracks episodically. As the ice aggrades the regolith margins rise above the elevation datum of the polygon centres. This forms low-centred polygons (Séjourné et al. 2011, 2012; Soare et al. 2014; 2018). As earlier, recharge by the atmospheric diffusion of water vapour could be of assistance in maintaining the volume and structural integrity of the ice wedges (Clifford, 1993; Mellon and Jakosky, 1993, 1995).

Later and lower obliquities, whereby the stability of ground ice falls, sublimate/degrade the ice-wedges preferentially at lower latitudes. This transforms high-shouldered margins into low-shouldered troughs, revises *lcps* into *hcps*, and decreases the ratio of *lcps* to *hcps* at these latitudes. Sand-wedge polygons would not be expected to fit the pattern of latitudinal distribution associated with the *lcps* and presented in **Section 5.1**, nor would the linear correlation between the *lcp/hcp* ratio and poleward latitudes.

Stage 4a: In conjunction with more recent high-obliquities a metres-deep icy and lightly-toned dusty mantle(s) accumulates in our study region (e.g. Mustard et al. 2001; Head et al. 2003; Forget et al. 2006; Madeleine et al. 2009). The mantle buries or mutes the thermokarstic (polygonised) depressions and the polygonised terrain incised by the depressions (Soare et al. 2017).

Stage 4b: As Mars moved cyclically back to lower obliquity(ies) the surface mantle began to sublimate and erode, uncovering and exhuming the terrain populated by the thermokarst-like depressions and possible ice-wedge polygons (Soare et al. 2017). Mirroring the antecedent and pre-mantling losses of ice, polygons at the lower latitudes show a higher observed-distribution of low shoulders or troughs (*hcps*) than polygons at higher latitudes.

7. Discussion & Conclusion

7.1 What if…

Questions concerning the possible existence of ice-rich terrain on Mars date back to the grainy Mariner-mission photographs of the 1960s and 1970s (e.g. Belcher et al. 1971; Sharp et al. 1971). In the aftermath, data-based validation of this hypothesis has been fleeting. “Looks-like” therefore “must-be” inferences of thermokarst-like landforms or of possible ice-wedge polygons are suggestive, even beguiling, but inadequate.

Validation, however, requires ground-truth, i.e. direct evidence of near-surface ground ice by a human or a robotic shovel at the latitudes of our study or, possibly, a *GPR* operating at frequencies sufficient to identify discrete and heterogeneous emplacements of ground ice from the near-surface to depth. Current inferences of widespread water-equivalent hydrogen (possible near-surface ground-ice) derived from orbital spectrometry have been less than satisfying. The data have a relatively short/shallow skin-depth (≤ 1 m) and a kilometre-scale footprint (e.g. Boynton et al. 2002; Feldman et al. 2002, 2004; Mitrofanov et al. 2002).

More recent observations of Utopia Planitia (~70-90° E), slightly to the west of our study region and derived of the ShallowRadar (*SHARAD*, Mars Reconnaissance Orbiter), point to the possible presence of a large and near-surface but decametres-deep body of admixtured ice, air and dust (Stuurman et al. 2016). Whether this buried body comprises massive ice, excess ice or something else remains open-ended, as does the time-frame of its origin, i.e. late Noachian Epoch, early or mid-Amazonian Epochs, etc, and the process by which it came to be, i.e. glacial, fluvial, marine, periglacial, a combination thereof, or something else.

North of our study region and substantially to the east (68.22^{o} N, 234.30^{o} E), shallow trenches dug by the Phoenix Lander revealed patches of light-toned ice that could be the work of ice/soil segregation (Mellon et al. 2009). The light-toned near-surface ice is thought to be geologically young, perhaps no older than 100 kyr (Mellon et al. 2009), much younger than the

periglacial terrain discussed by us above. The depth and distribution of this segregated ice, at this location and possibly elsewhere, is unknown.

7.2 Poleward latitude, ice-wedging and the freeze-thaw cycling of water

1) Our study region in Utopia Planitia (40-50° N; 100-125° E) displays a statistically-significant and positive (linear) correlation between the ratio of low-centred to high-centred polygons (*lcps* vs *hcps*) and a poleward gradient of *lcp/hcp* distribution. As near-surface ground ice is more stable at higher latitudes than at lower ones, the increased (observed) occurrence of *lcps* at poleward latitudes would be expected were their margins underlain by ice-wedges. Also expected would be the lower prevalence of *lcps* at the equatorward latitudes where ice is less stable than at poleward ones.

2) This latitudinal gradient of *lcp/hcp* distribution would not be expected or predicted were the polygon margins underlain by sand wedges. The gains and losses of elevation in the margins of sand-wedge polygons are driven by aeolian accumulation or erosion or, in relatively unique periglacial contexts, by the sublimation of underlying glacial-ice. Variances in the stability of ground-ice, expressed on a local, regional or latitudinal scale, exercise no positive or negative effect on the relative elevations of sand-wedge polygon centres or margins.

3) The morphometric/statistical validation of ice-wedge polygonization along a latitudinal gradient of distribution at the mid-latitudes of *UP* accomplishes two things. First, it enhances the empirical/observational depth of excess-ice hypotheses and, second, it underpins the possibility that excess-ice could be present, even today, at lower than poleward (Phoenix-like) latitudes and closer to the surface in the northern plains than might otherwise be thought.

4) Equally important is a return to questions concerning the time-scale required to form metre-high ice-wedges on Mars. Ice-wedges on Earth may originate and evolve into mature landforms on relatively short scales of time. For example, within a few years of a thermokarst lake in the Tuktoyaktuk region having been drained of its water artificially, the newly-exposed basin underwent permafrost aggradation; thermal-contraction cracking also was observed, as were nascent ice-wedges (Mackay and Burn, 2002). Although hundreds of seasonal/annual freeze-thaw cycles are required to evolve mature and metre-scale ice wedges from sub-centimetre ice veins, even this somewhat extended period is relatively short, geologically speaking.

On Mars, the areas where liquid water is meta-stable today could have been much more long-standing and wide-ranging in the past under more favourable obliquity solutions, especially if surface or near surface brines were present then as they appear to be now. This would have depressed the freezing point of water and lowered the temperature constraints on its freeze-thaw cycling.

5) The crater counts (see **Section 5.2**) of the polygonised terrain in our region indicate that the latter is substantially older than previous studies of similar terrain in Utopia Planitia have suggested, perhaps by an order of magnitude. This attenuates the possible inconsistency between the more temperate boundary-conditions required by the formation of ice-wedge polygons, under higher obliquities perhaps, and the current constraints of extreme aridity, low temperatures and low atmospheric-pressure. Additionally, slightly-higher temperatures in a volatile-rich environment could have induced an accompanying rise of atmospheric vapour-pressure. This would have shortened the period of time and the number of iterative (freeze-thaw) cycles required to originate and grow ice wedges.

6) Attention also ought to be cast on the possibility that disparate transformational-processes might be responsible for ice-wedge aggradation and ice-wedge degradation on Mars. The freeze-thaw cycling of water may be the most plausible means by which wedge ice, or even segregation ice, aggrade, on Mars as on Earth. On the other hand, the degradation of Martian wedge or segregation ice could and most probably is the work of sublimation. Reconciling the two end-processes may be no more difficult than ascribing a less than recent origin or aggradation period to the ice wedges and a possibly very-late Amazonian Epoch and much more recent degradation period to them.

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Figures

Fig 1. View of the study region in Utopia Planitia (Lambert conformal conic-projection). The footprints of the 119 *HiRISE* images scanned for this study displayed with different colours according to the type(s) of polygons observed therein: blue, low-centred polygons (*lcps*); red, high-centred polygons (*hcps*); blue/red strips: *lcps* and *hcps*; white: no polygons; and, black: non-analysable image. Background: *MOLA* (Mars Orbiter Laser Altimeter) elevation data overlapping *THEMIS* (Thermal Emission Imaging System) daytime-*IR* mosaics. *MOLA* data credit: *MOLA* Science Team; Arizona State University; *THEMIS* image credit: Arizona State University.

Fig. 2. a) *HiRISE* image ESP_055038_2250 gridded by 500x500 metres squares (background: *MOLA* elevation data overlapping a *THEMIS* daytime-*IR* mosaic). **b)** Zoom on four 500x500 metres-squares that illustrate the mapping method. Red and blue dots respectively locate *lcps* and *hcps*. North is up on all tiles. *HiRISE* image credit: *NASA*/*JPL*/University

of Arizona; *MOLA* data credit: *MOLA* Science Team; Arizona State University; *THEMIS* image credit: Arizona State University.

Fig. 3. Examples of the three polygon-types, categorised by the relative elevation-differences between polygon centres and margins, and disparate locations, in our study region. **a)** Scalloped depressions overprinted completely by *lcps* (*HiRISE* image PSP_006908_2215, 41.178° N, 124.130° E, 25cm/pixel). **b)** Ubiquitous distribution of *hcps* throughout scalloped terrain (*HiRISE* image ESP_046467_2280, 47.594° N, 119.679° E, 25cm/pixel). **c)** Proximal *lcps* (black arrows) and *hcps* (white arrows) (*HiRISE* image ESP_055038_2250, 44.757° N, 113.888° E, 50cm/pixel). **d)** Polygons with no observable elevation-difference between centres and margins (*HiRISE* image ESP_026450_2270, 46.823º N, 110.880º E, 50cm/pixel). **e)** Larger-scale view of panel **d)** within the inter-crater plains. **f)** *Lcps* (black arrows) and *hcps* (white arrows) in close proximity (*HiRISE* image ESP_036366_2235, 42.939º N, 115.428º E, 50cm/pixel). **g)** Panel **f)** at a larger scale. Note: these polygons are in **Type 2** terrain (background: *THEMIS* day-*IR* controlled mosaic). North is up in all images. *HiRISE i*mage credits: *NASA/JPL*/University of Arizona.

Fig 4. a) Circular and individual rimless depressions *(HiRISE* image ESP_053205_2235, 43.350° N, 117.223° E, 25cm/pixel). **b)** Elongated rimless-depression *(HiRISE* image ESP_026094_2250, 44.657° N, 111.415° E, 25cm/pixel). **c)** Coalesced rimless depressions; note the scalloping of margins *(HiRISE* image ESP_026094_2250, 44.657° N, 111.415° E, 25cm/pixel). Note the inward-facing tiers (black dashed lines). **d)** Wave-like depressions (*HiRISE* image ESP_035667_2295, 49.377° N, 115.622° E, 25cm/pixel). North is up on all tiles. *HiRISE* image credits: *NASA*/*JPL*/University of Arizona.

Fig. 5. a) Early-stage polygonization of continuous permafrost, by way of thermal contraction

cracking (Tuktoyaktuk, Northwest Territories, Canada). **b)** Ubiquitous coverage of continuous permafrost by small-sized, possibly ice-wedge, polygons (Husky Lakes, [*HL*] midway between Tuktoyaktuk and Inuvik, Northwest Territories). **c)** Low-centred (ice-wedge) polygons, with marginal but elevated troughs hosting wedge-derived meltwater (*HL*). **d)** High-centred (ice-wedge) polygons, with marginal troughs filled with melt-water (*HL*). **e)** Ice-wedge exposure in scarp eroded by Beaufort Sea wave-action (Peninsula Point, ~6 km southwest of Tuktoyaktuk, Northwest Territories, Canada). Note the surface depressions above the large ice-wedge on the right and the dangling ice-wedge remnants to its left. The depressions, were they observed planimetrically, would comprise the marginal troughs of high-centred (degradational) polygons. The brownish sediments in whose midst the ice-wedges lie are silty clays. **f)** Exposed sand wedge (Tuktoyaktuk Coastlands) (Murton et al., 2000). Image credits for all panels other than **f)**: R. Soare.

Fig. 6. a) Hummocky thermokarst, featuring thermokarst lakes, polygonised alases and beaded streams linking thaw pools of degraded ice-wedge polygons (Husky Lakes, Northwest Territories). **b)** Panoramic and cross-sectional view of an ice complex comprised of ice wedges (on the right-hand side) and massive ice. The small, surface depressions flanking the largest ice wedge (on the right-hand side, black arrows) mark the location of bilateral polygonal troughs associated with a field of high-centered polygons. In a wet permafrost-environment high-centered, ice-wedge polygons form when and if local or regional (mean) temperatures rise to the point of inducing thaw and melt at or near the surface where the ice wedges occur. Grayish-beige material on the right and throughout the floor of the terrace comprises silts and clays that have slumped to the floor from the near surface as retrogressive thaw has eroded this Beaufort Sea headland. The oblong block on the right-

hand side of the terrace floor points to the recentness of thaw-based detachment as there is a gap in the overhang of the terrain immediately above the block location (Peninsula Point, Northwest Territories). Image credits: R. Soare.

Fig. 7. Low-centred vs high-centred polygon latitudinal ratios for all terrain types. Data for all images have been binned for each degree of latitude excluding the upper bound. The black dashed line represents the linear correlation between *lcp*/*hcp* and latitude, its R^2 value specified above it. No correlation is observable. Error bars are calculated by assuming one in 15 squares could be erroneously classified and are not visible if they fall within the points.

Fig. 8. Low-centred vs high-centred polygon latitudinal ratios (plains, excluding craters). Data for all images have been binned for each degree of latitude excluding the upper bound. The black dashed line represents the linear correlation between *lcps*/*hcps* and latitude. The R^2 value is specified above it. A significant correlation is observed (*p-value*: 6.65×10^{-5}). Error bars are calculated by assuming one in 15 squares could be erroneously classified and are not visible if they fall within the points.

Fig. 9. a) Portion of *HiRISE* image ESP_026450_2270 within the study region (centered on 46.823° N, 110.880° E) showing impact craters on the polygonized terrain. **b)** Cumulative unbinned and **c)** log-incremental root-2 binned crater size-frequency distribution for the entire *HiRISE* image (Area: 150 km^2) with absolute model age isochrons for 1 Ma, 10 Ma, and 100 Ma surfaces shown from Hartmann (2005). *HiRISE* image credit: *NASA*/*JPL*/University of Arizona.

Fig. 1

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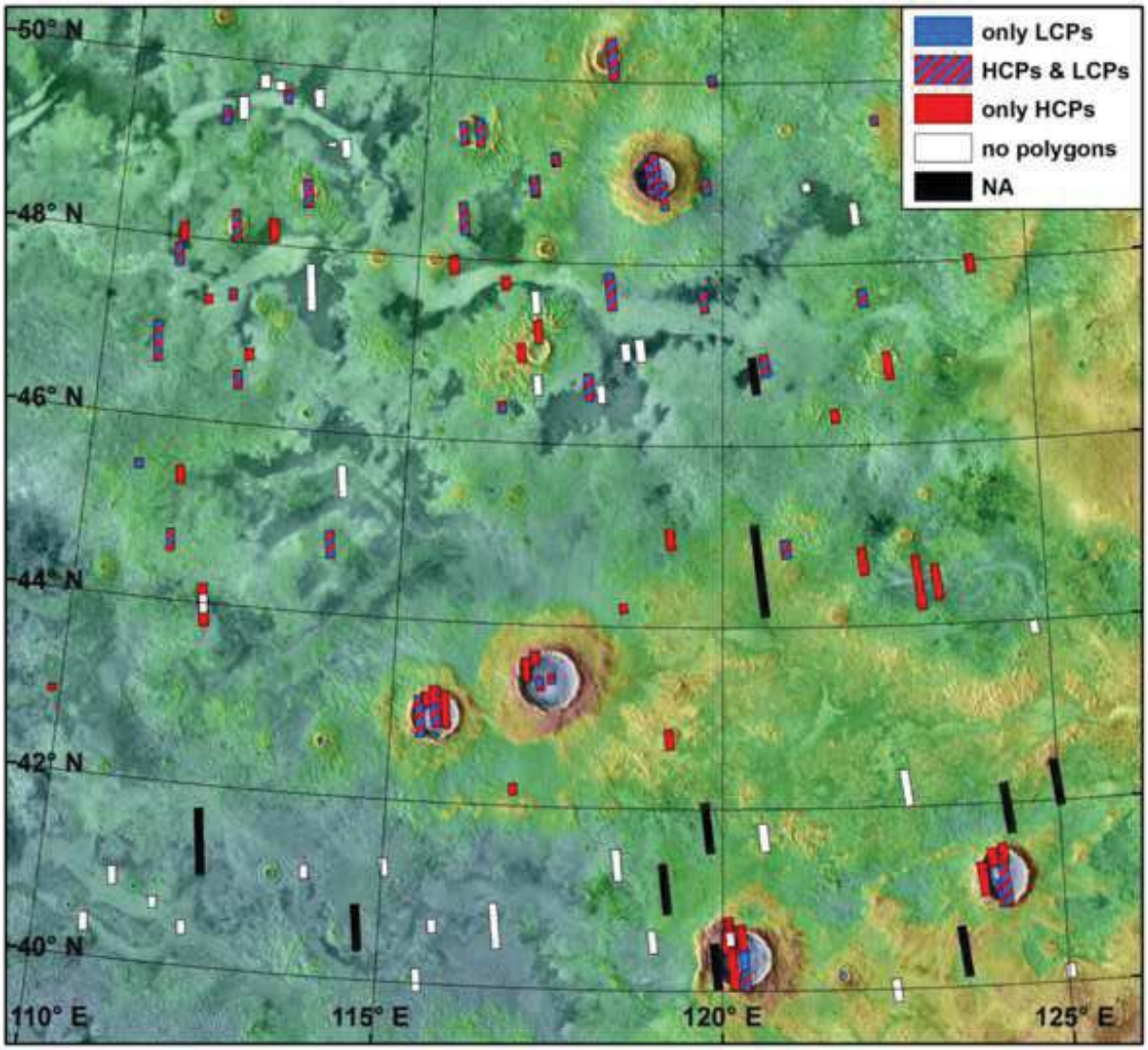

Fig. 2

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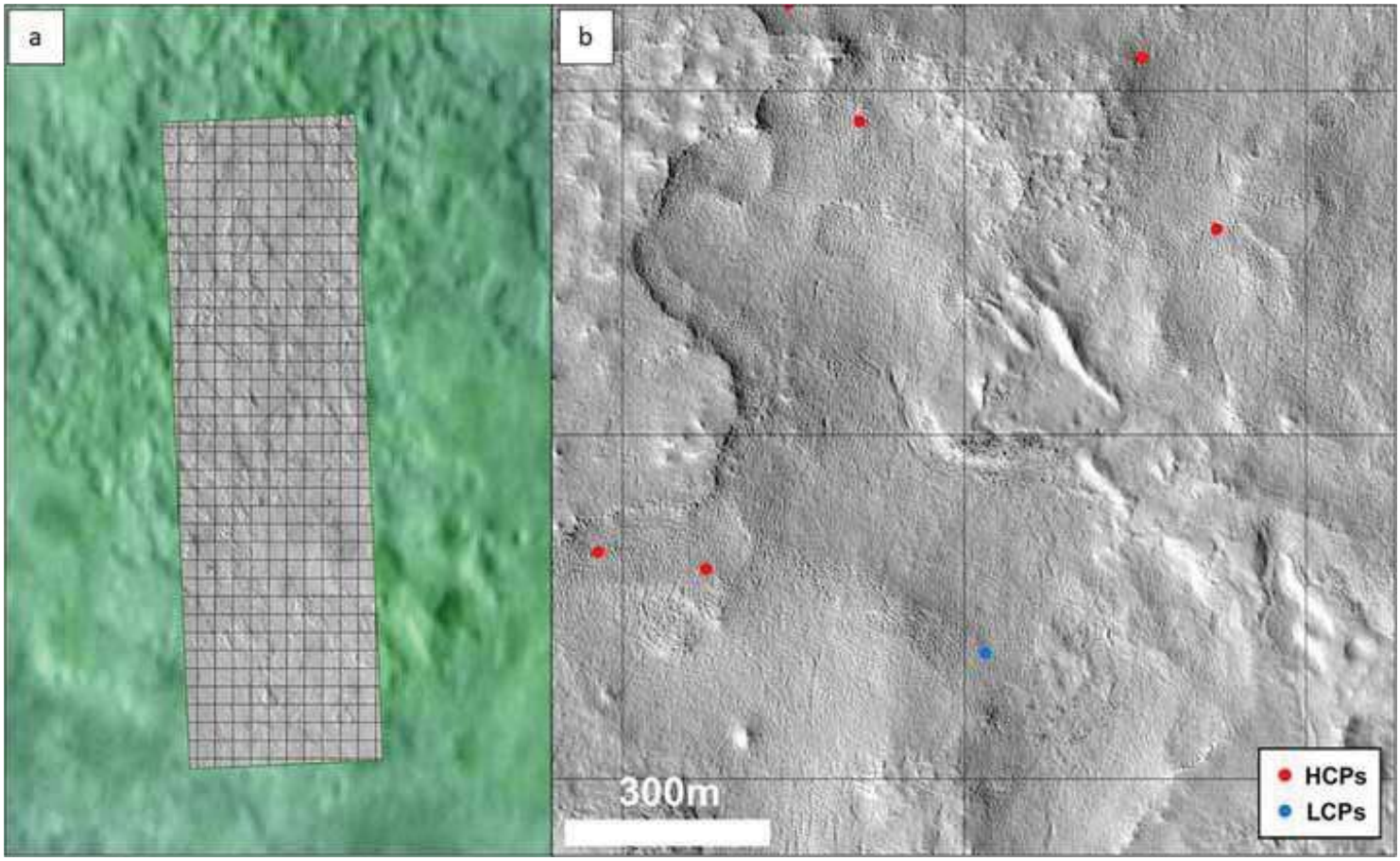

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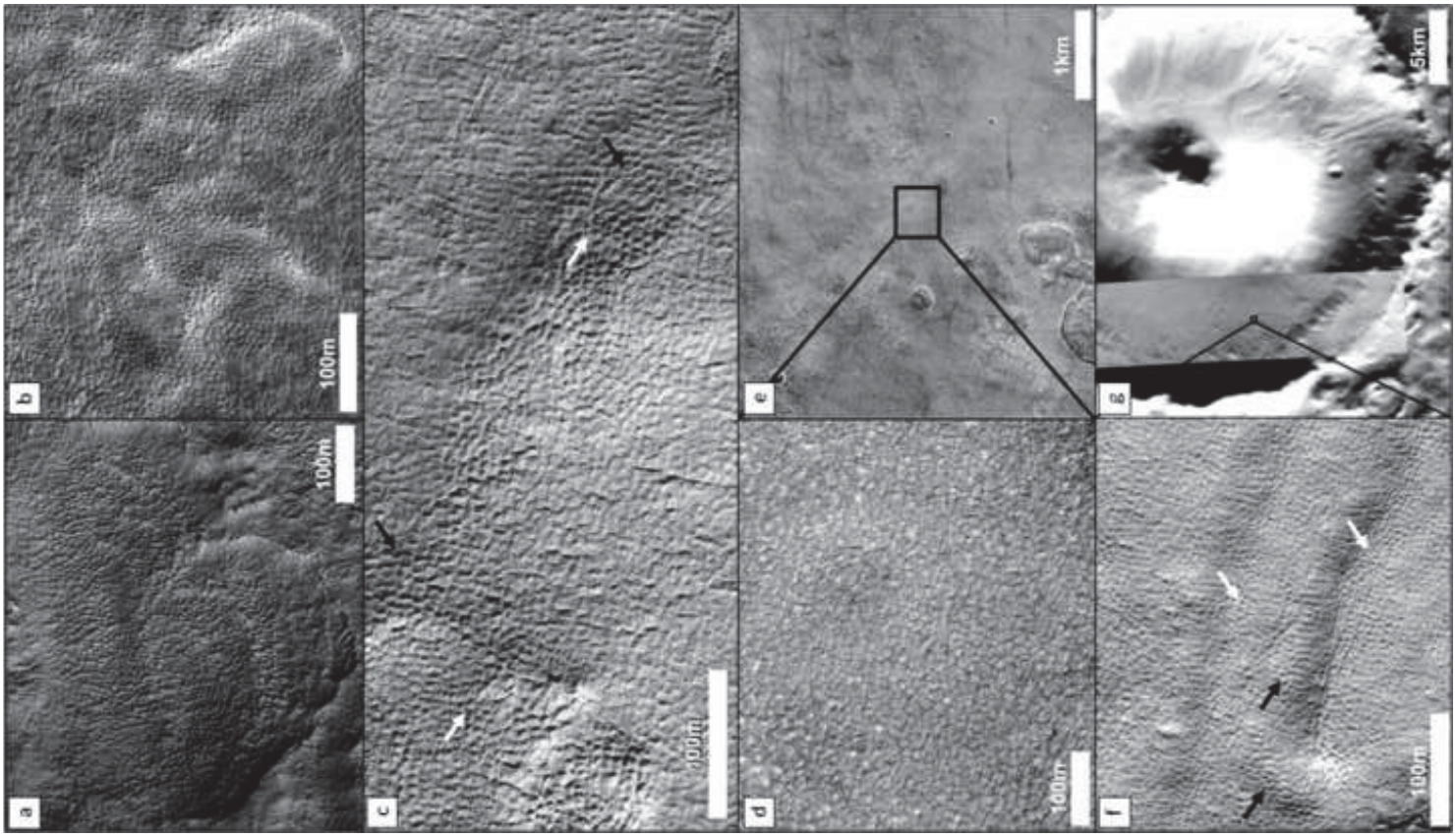

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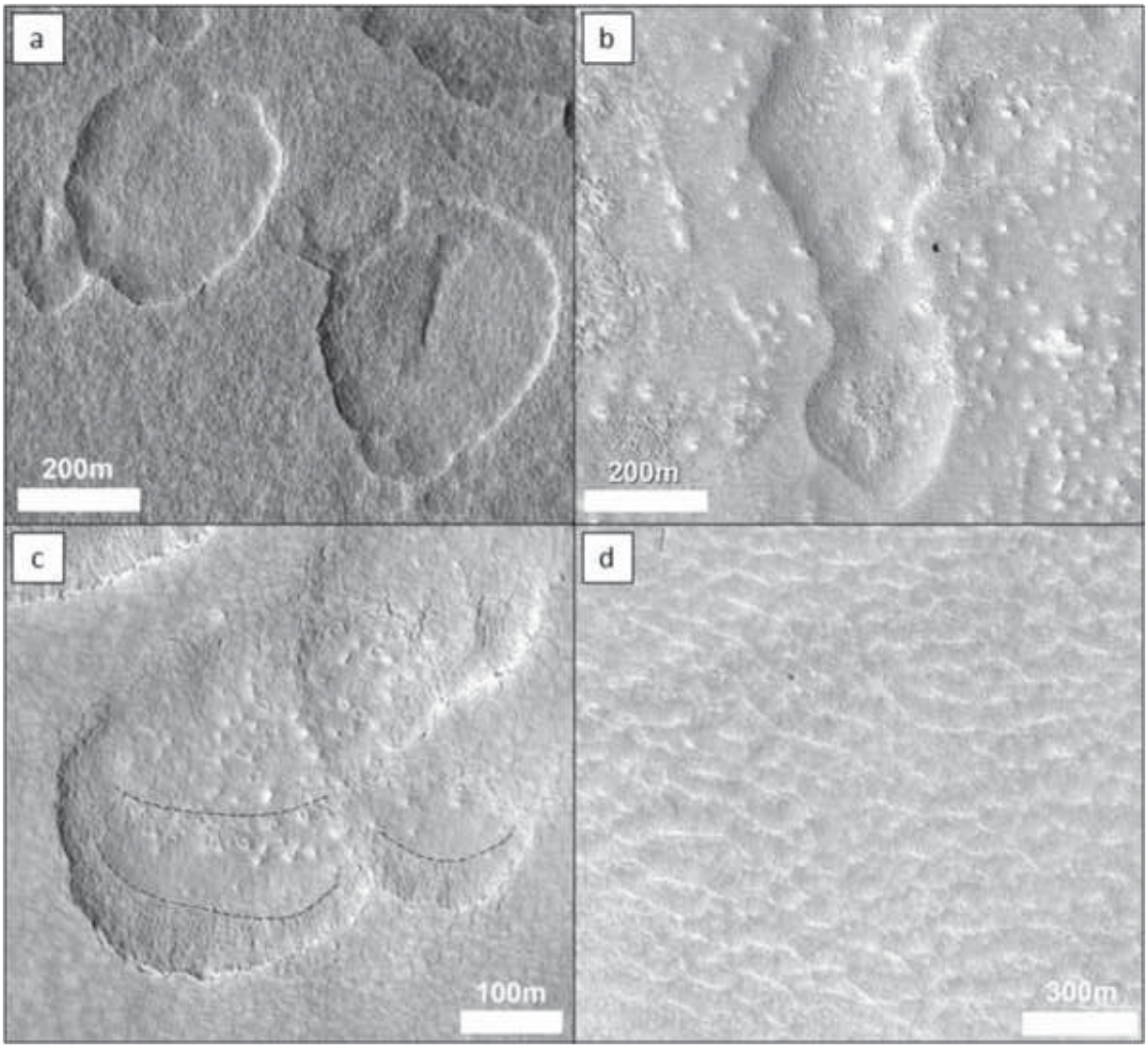

Fig. 5

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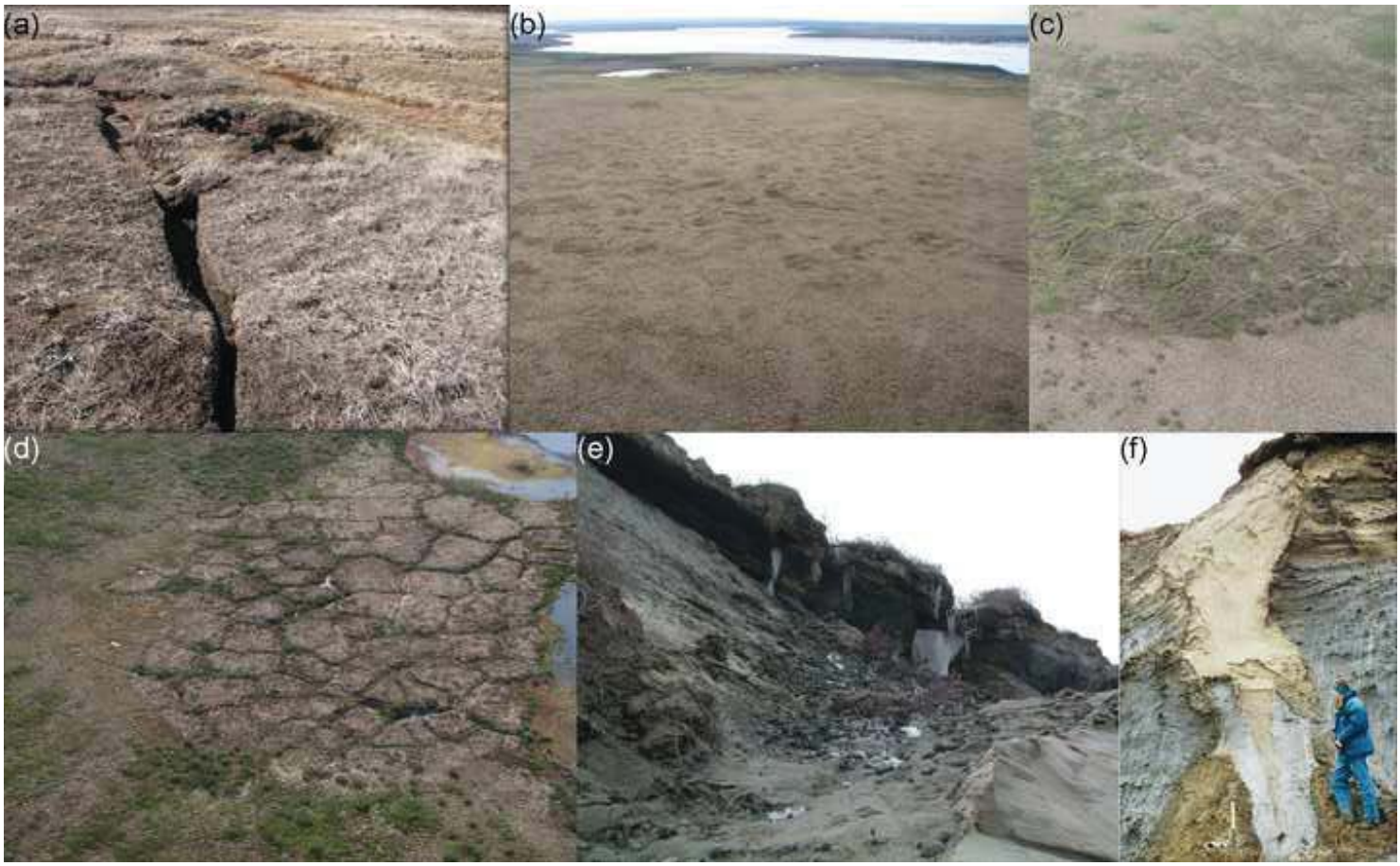

Fig. 6

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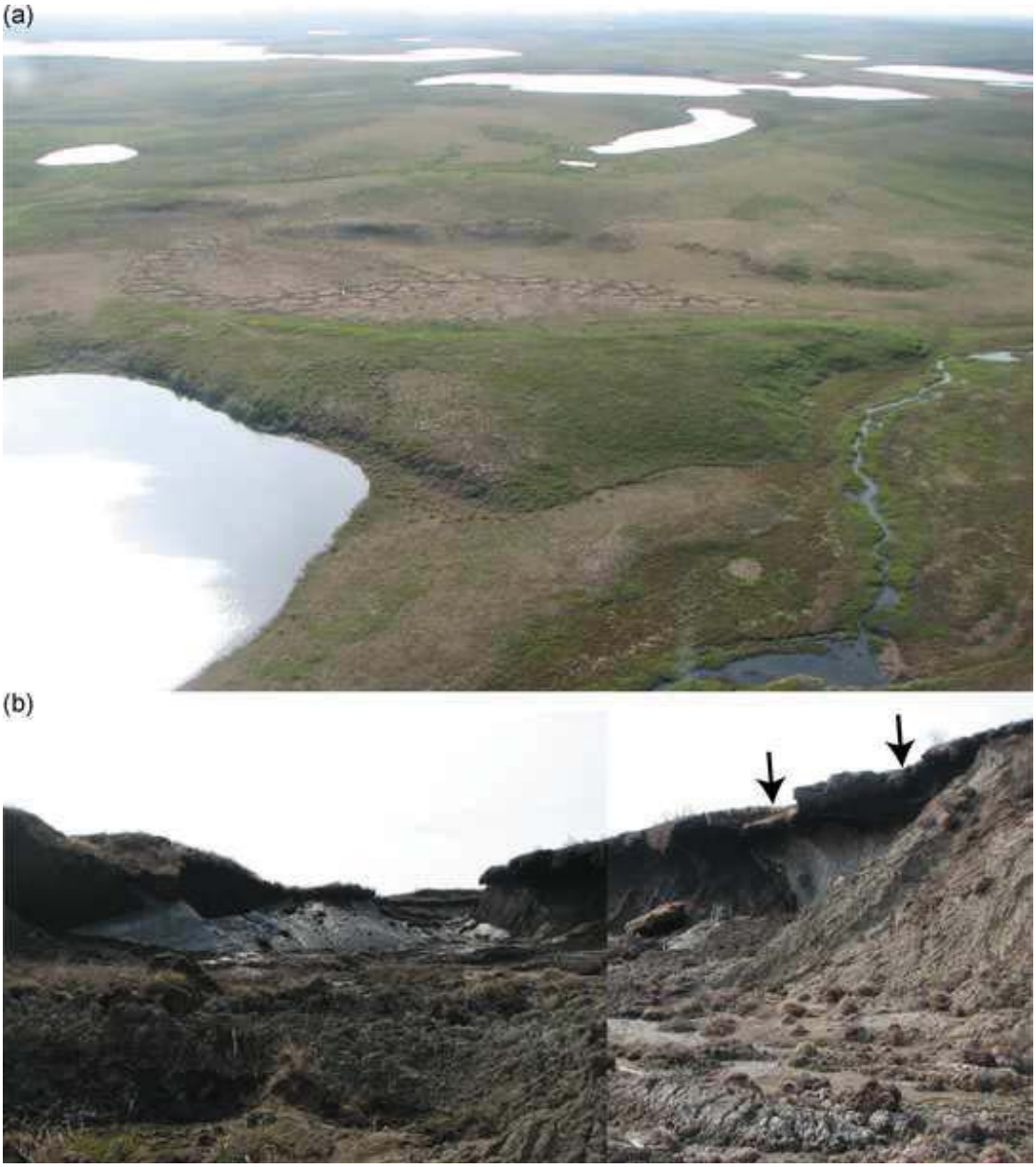
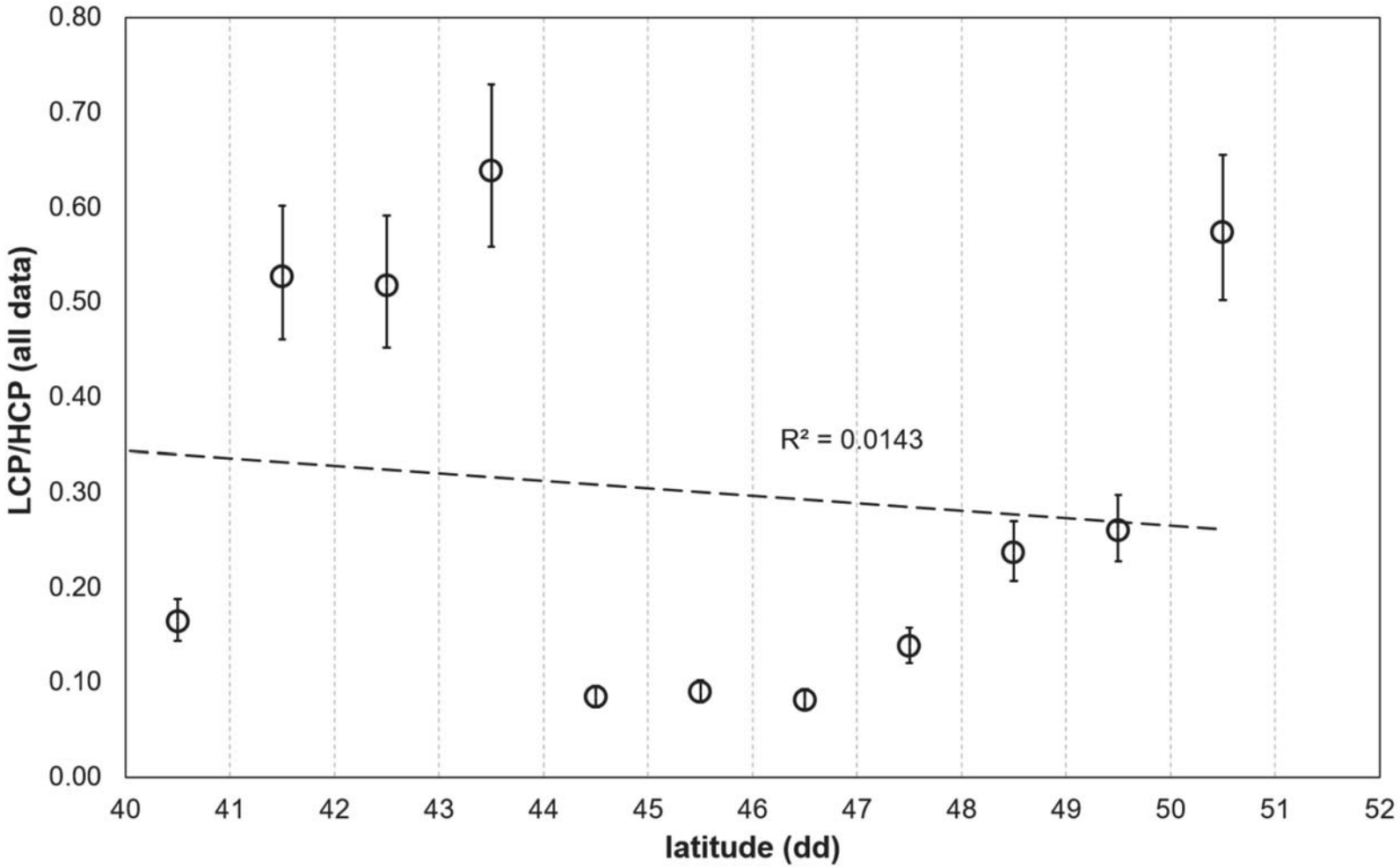


Fig. 7


0.80
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LCP/HCP (all data)
R² = 0.0143
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latitude (dd)

Fig. 8

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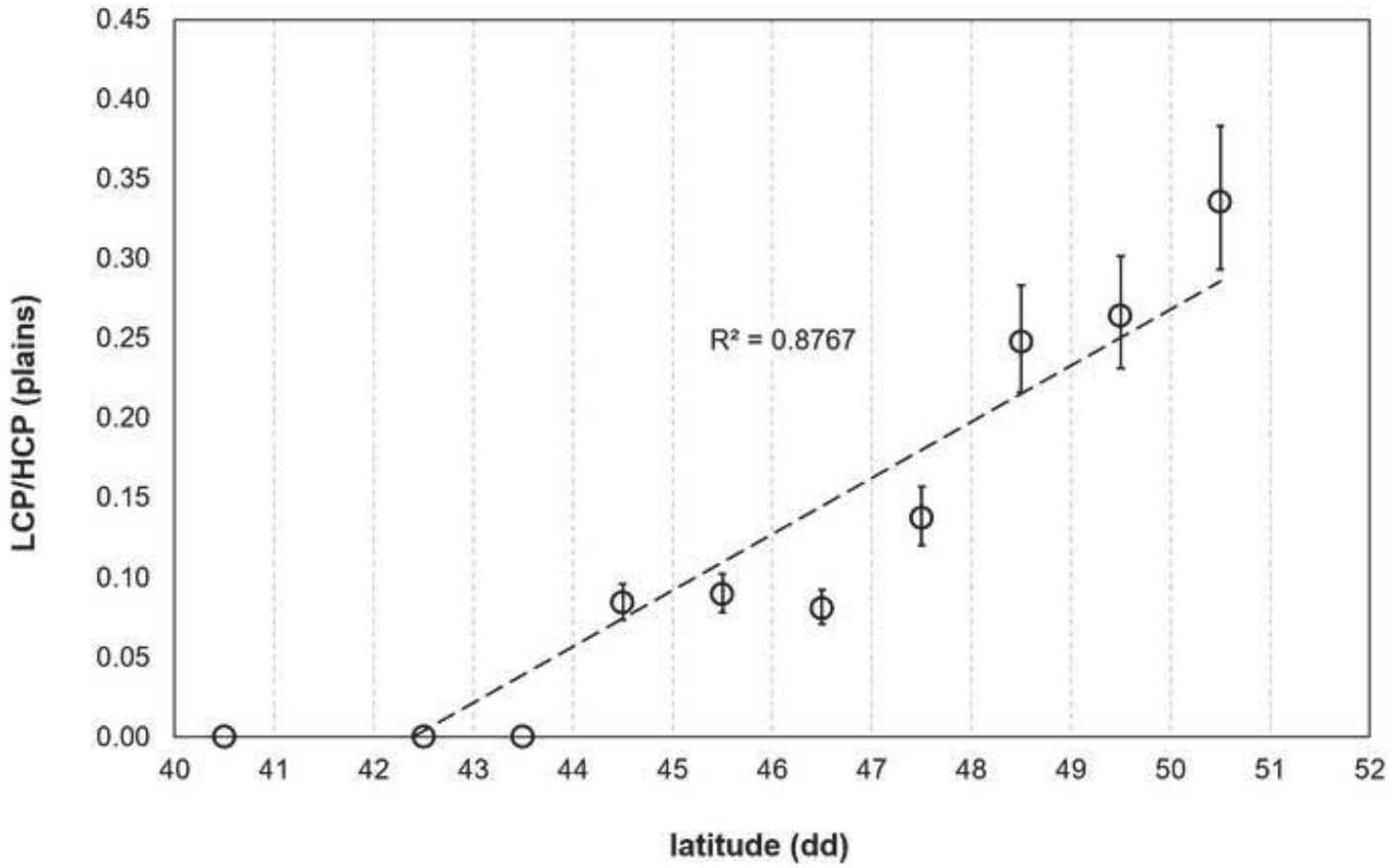

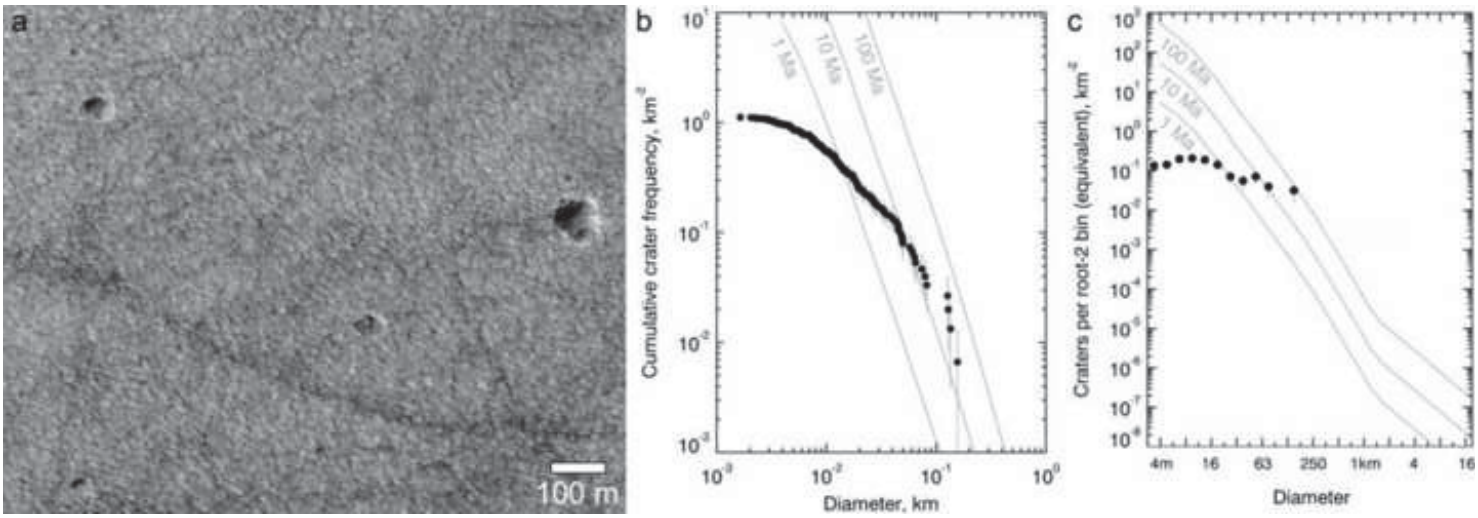
a
100 m
b
Cumulative crater frequency, km^{-2}
1 Ma
10 Ma
100 Ma
Diameter, km
c
Craters per root-2 bin (equivalent), km^{-2}
100 Ma
10 Ma
1 Ma
4m
16
63
250
1km
4
16
Diameter