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2 **Possible ice-wedge polygonisation in Utopia Planitia, Mars**
3 **and**
4 **its latitudinal gradient of distribution**
5

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32 ABSTRACT

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

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

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

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

51 Here, we show that:

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

- 55 2) This linear correlation would be expected, in as much as ground-ice stability increases with
56 latitude, were the shoulders of higher-latitude *lcps* underlain by (aggraded) ice-wedges and
57 those of lower-latitude *hcps* underlain by (degraded) ice-wedges.
- 58 3) The change of polygon morphology with latitude would not be expected were the *lcps* and
59 *hcps* underlain by sand wedges, in as much as ground-ice stability is unrelated to their
60 aggradation or degradation.
- 61 4) Crater counts of the polygonised terrain indicate that it is less youthful than previous
62 studies have suggested, perhaps by an order of magnitude. This attenuates the possible
63 inconsistency between the more temperate boundary-conditions required by the formation
64 of ice-wedge polygons and the current constraints of extreme aridity, low temperatures and
65 low atmospheric pressure.

66 1. Introduction

67 On Earth, *ice complexes* (also referenced as ice-rich terrain) at or near the surface or to
68 decametres of depth are ubiquitous within the permafrost of Arctic coastal-plains, i.e. the
69 Tuktoyaktuk region of northern Canada (e.g. Rampton and Bouchard, 1975; Rampton, 1988;
70 Murton, 2005) and the Laptev peninsula of northeastern Siberia (e.g. Schirrmeister et al. 2002,
71 2017; Grosse et al. 2007; Morgenstern et al. 2013; Strauss et al. 2017). Some examples of this
72 terrain are:

- 73 a) Hummocky thermokarst (ballooned by aggraded excess-ice and deflated where the
74 excess ice has been lost by evaporation, drainage or sublimation).
- 75 b) Thermokarst lakes, i.e. pooled meltwater resulting from the thaw of excess ice.
- 76 c) Alases (thermokarst-lake basins depleted of water).

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

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

91 Ice-wedge polygon networks are current as well as paleo-markers of temperature variances.
92 The nascent conditions required for the ice-wedges to form are continuous permafrost, thermal-
93 contraction cracking, meltwater, hoarfrost (to a much lesser degree), and seasonal/annual (mean)
94 temperature fluctuations above and below 0° C (e.g. Lachenbruch, 1962). Where low-centred
95 (aggraded) and high-centred (degraded) polygons are closely set, this marks a system actively in
96 thermal disequilibrium and transition (Washburn, 1973; French 2007; also, Rampton and
97 Bouchard, 1975; Rampton, 1988; Schirrneister et al. 2002; Grosse et al. 2007; Morgenstern et al.
98 2013).

99 Mars-based *lcps* and *hcps* that are similar in shape and size to those observed on Earth and,
100 on occasion have been reported widely in the literature (e.g. Lefort et al. 2009; Levy et al. 2009a,
101 2010; Séjourné et al. 2011; Soare et al. 2014, 2018). Questions concerning whether ice or sand
102 underlies the margins of the Martian *lcps* and *hcps* have been and remain open-ended, for two
103 main reasons (e.g. Lefort et al. 2009; Levy et al. 2009a, 2010; Séjourné et al. 2011; Soare et al.
104 2014, 2018).

105 First, and based solely on the plan-based observation of *lcps* or *hcps* on Earth and on Mars,
106 the identification or discrimination of sub-marginal materials is not possible. This is due to the
107 similarity of form and scale between ice-and sand-wedge polygons when viewed planimetrically,
108 from above (Levy et al. 2009a, 2010; Hallet et al. 2011; Hauber et al. 2011; Soare et al. 2014,
109 2018).

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

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

122 northern plains of Mars. We assume that here in our study region, as elsewhere on Mars, the
123 stability of ground ice increases with its poleward latitude. This is because (mean) annual surface
124 and near-surface temperatures tend to remain below the atmospheric frost-point with poleward
125 proximity, mitigating ice loss by sublimation (e.g. Clifford, 1993; Mellon and Jakosky, 1993,
126 1995) Consequently, ice-aggraded polygons should show a higher density of distribution with
127 poleward proximity than ice-degraded polygons and *vice versa*.

128 2. Methods

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

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

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

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

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

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

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

168 N) was selected within the study region as it contained polygons throughout the image; counts
169 were conducted on the entire image covering 150 km² at 0.5 m pix⁻¹ resolution using the
170 *Cratertools* plug-in for the *ERSI ArcGIS* to measure crater diameters (Kneissl et al. 2011). The
171 crater-size frequency distribution (*CSFD*) was compared with modeled crater-retention age
172 isochrons from Hartmann (2005) using *Craterstats* (Michael and Neukum, 2010; Michael et al.
173 2016).

174 3. Observations

175 Small-sized ($\sim \leq 25$ m in diameter) and clastically non-sorted polygons are ubiquitous in
176 our study region. Typically, the polygons are hexagonal and enclosed by metre-scale margins. The
177 margins exhibit: **a**) slightly-elevated (metre-high) shoulders (low-centred polygons, *lcps*) (**Figs.**
178 **3a, c**); **b**) depressed troughs (high-centred polygons, *hcps*), relative to the polygon centres (**Figs.**
179 **3b-c**); or, **c**) variances of elevation that are small or non-existent (**Fig. 3c**), at the highest resolution
180 available (~ 25 cm/pixel, *HiRISE* image).

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

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

191 observed individually (**Fig. 4a**), clustered or coalesced (**Fig. 4c**), or in wave-like forms (**Fig. 4d**).
192 Orientation, defined by the loss of elevation relative to the surface datum, is longitudinal and
193 equatorward (e.g. Morgenstern et al. 2007; Lefort et al. 2009; Séjourné et al. 2011, 2012). Depths
194 range from metres to decametres (Morgenstern et al. 2007; Lefort et al. 2009; Séjourné et al. 2011,
195 2012). Inward-oriented and roughly equatorial tiers or terraces often demarcate the changes of
196 elevation and are commonplace amidst the scalloped depressions (**Fig. 4c**).

197 **4. Ice-complex constituents on Earth**

198 *4.1 Thermal-contraction cracking and ice wedge polygons*

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

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

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

215 networks (Washburn, 1973; Mackay, 1974; Mackay and Matthews, 1983). The latter may
216 comprise hundreds of square kilometres of continuous or discontinuous coverage and represent
217 hundreds of fill and crack-formation cycles (e.g. Black, 1954; Lachenbruch, 1962; Washburn,
218 1973; Mackay, 1974) (**Fig. 5b**).

219 Thermal-contraction polygons exhibit low (**Fig. 5c**) or high centres (**Figs. 5d-e**) relative to
220 their margins or centres, or margins that are relatively-equal in elevation (**Fig. 5b**), depending upon
221 their developmental stage. Ice-wedge aggradation is informed by the (seasonally) iterative infilling
222 and freezing of meltwater at polygon margins, as well as by the active-layer expansion that moves
223 sediment radially to the polygon margins from the centres. This uplifts the sedimentary
224 overburdens, forms raised shoulders (Mackay, 1981; French and Guglielmin, 2000; French, 2007)
225 and gives the polygons a distinctly low-centred appearance (Czudek and Dumek, 1970; Rampton,
226 1988; Hallet et al. 2011). Ice-wedge degradation follows from a sustained rise of thaw-tolerant
227 mean temperatures. This deflates polygon-margin shoulders, perhaps sinking them below polygon-
228 centre elevation, morphing low-centred into high-centred polygons (Czudek and Dumek, 1970;
229 Rampton, 1988; Hallet et al. 2011). Regions or sub-regions where *lcps* and *hcps* are observed
230 concurrently are in thermal flux (Czudek and Dumek, 1970; Rampton, 1988; Hallet et al., 2011).

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

234 *4.2 Sand-wedge, composite and sublimation-enhanced polygons*

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

238 they are absent from the landscape (e.g. Péwé, 1959; Berg and Black, 1966; French and
239 Guglielmin, 2000; Marchant et al. 2002; Bockheim et al. 2002; Sletten et al. 2003; Hallett et al.
240 2011). In both cases the sand-wedge polygons are paleo-markers of climatic and hydrological
241 regimes that are drier and/or colder than those associated with the origin and development of ice-
242 wedge polygons (Washburn, 1973; Mackay and Matthews, 1983; Murton et al., 2000; Wolfe et al.
243 2018). For example, contrast the annual mean-temperatures and -precipitation at the Tuktoyaktuk
244 coastlands (discussed above) with the mean annual air-temperatures and precipitation at Beacon
245 Valley, one of the *MDVs*. At Beacon Valley, annual mean-temperatures are $\sim -34^{\circ}$ C and soil
246 temperatures possibly do not exceed $\sim 0^{\circ}$ C (Marchant et al., 2002). Precipitation is meagre
247 (< 50 mm of water equivalent precipitation per annum) and is comprised solely of snowfall
248 (Bockheim et al. 2002; Levy 2015). Boundary conditions consistent with the iterative formation
249 of ice-wedge polygons are rare, perhaps not exceeding a few days per annum.

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

259 *4.3 Thermokarst and ice-rich landscapes*

260 Thermokarst refers to landforms developed by way of freeze-thaw cycling in permafrost
261 composed of excess ice (e.g. Taber, 1930; Penner, 1959; Hussey, 1966; Hughes, 1974; Rampton,
262 1973, 1988; Harris et al. 1988; Osterkamp et al. 2009). Excess ice = "...the volume of ice in the
263 ground that exceeds the total pore volume that the ground would have under unfrozen conditions"
264 (e.g. Harris et al. 1988; also, see Taber, 1930; Penner, 1959; Rampton and Mackay, 1971;
265 Washburn, 1973; Murton, 1996; Rampton, 1988; French, 2007).

266 Excess ice comprises ice lenses, veins, wedges or larger masses of consolidated ice that are
267 formed by ice segregation. Fine to medium-grained soils such as silts or silty clays are particularly
268 adept at hosting segregation ice because they have relatively-small interstices (Taber, 1930; Black,
269 1954; Penner, 1959; Rampton and Mackay, 1971; Rampton, 1988; Sone et al. 2006; French, 2007;
270 Abramov et al., 2008). Small interstices facilitate the migration of pore water to and the formation
271 of discrete layers or lenses of segregation ice at the freezing front (Taber, 1930; Penner, 1959;
272 Rampton and Mackay, 1971; Rampton, 1988; French, 2007). Were the grain-size too fine, pore-
273 water migration would be overly constricted and ice lenses would not form; were the grain-size
274 too coarse, pore-water migration would be overly expeditious and flush out of the system before
275 segregation ice could form.

276 Where excess ice aggrades, frost heave follows (Taber, 1930; Penner, 1959; Rampton,
277 1988). Where excess ice thaws and degrades, the terrain subsides and settles under its own weight,
278 until it reaches a consolidated state (Harris et al., 1988) (**Figs. 6a-b**). If thaw-generated meltwater
279 pools then a thermokarst pond or lake forms; were the pooled meltwater to be lost by evaporation
280 or drainage, an emptied basin or alas would form (Washburn, 1973; Rampton, 1988; French,
281 2007). Were the loss of pooled water partial and episodic, inwardly-oriented benches or terraces

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

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

291 **5. Results**

292 *5.1 LCPs/HCPs and poleward latitude (Table 1)*

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

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

305 statistically-significant linear correlation appears between the *lcp*-to-*hcp* ratio and latitude ($R^2 >$
306 0.87 , p -value <0.05 (**Fig. 8**). Note that **Fig. 8** and **Table S1** data include three latitudinal ranges
307 without *lcps* (40° - $<41^\circ$, 42° - $<43^\circ$, 43° - $<44^\circ$; $lcp/hcp = 0$), and one range without polygons (41° -
308 $<42^\circ$; $lcp/hcp =$ no data).

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

318 Sixty-eight percent of the squares with *lcps* also have *hcps*. This increases to 83% if **Type**
319 **2** grid squares are excluded from the count and decreases to 37% when **Type 2** grid-squares alone
320 are counted.

321 *5.2 Surface-age estimates*

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

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

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

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

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

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

352 *6.1 The origin and development of ice-rich terrain*

353 **Stage 1a:** Atmospheric water-vapour (the Martian equivalent of winter hoar-frost on Earth)
354 diffuses and charges the pore space of near-surface regolith (top few metres) with ice at all latitudes
355 during one or more periods of high obliquity during the Late Amazonian Epoch (e.g. Clifford,
356 1993; Mellon and Jakosky, 1993, 1995; Douglas and Mellon, 2019). In concert (or not) with this
357 diffusion-driven ice-charging of near-surface regolith, high obliquities engender enhanced periods
358 of H₂O-ice precipitation. The latter accumulates at the surface and forms an icy mantle (e.g. Head
359 et al. 2003; Forget et al. 2006; Madeleine et al. 2009, 2014).

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

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

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

375 **Stage 2:** As high obliquity/ies wanes/wane, near-surface ground ice at the lower mid-
376 latitudes of our study region loses stability and sublimates (Mellon and Jakosky, 1993, 1995).
377 Thermokarst or alas-like basins form where the volatile loss occurs (Morgenstern et al. 2007;
378 Lefort et al. 2009; Ulrich et al. 2010; Séjourné et al., 2011, 2012; Mangold, 2011; Soare et al.
379 2014; Dundas et al. 2015; Dundas, 2017). The episodic or intermittent return to higher obliquities,
380 much like glacial inter-stades on Earth (e.g. Rampton and Bouchard, 1975; Rampton, 1988),
381 could/would have replenished and recharged the volatile-depleted terrain with ice. Against this
382 backdrop, the rimless and sometimes scalloped depressions in our study region, and throughout
383 the mid-latitudes of *UP* might be indicative of the net loss of ice-richness through the Late
384 Amazonian Epoch to the present day. Their shape, size, depth, and location possibly reflect random
385 differences in the distribution and depth of ground ice as well as in the reach of thermal
386 disequilibrium (Soare et al. 2018). In this regard, the inwardly-oriented steps or tiers often
387 observed within the depressions could be markers of an episodic loss of ice, much in the same way
388 as tiered thermokarst basins in the Tuktoyaktuk region point to the episodic loss of water (Soare
389 et al. 2007, 2008; 2017; also, Séjourné et al. 2011, 2012).

390 *6.2 The origin and development of ice-wedge polygons*

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

395 fills the polygonised cracks episodically. As the ice aggrades the regolith margins rise above the
396 elevation datum of the polygon centres. This forms low-centred polygons (Séjourné et al. 2011,
397 2012; Soare et al. 2014; 2018). As earlier, recharge by the atmospheric diffusion of water vapour
398 could be of assistance in maintaining the volume and structural integrity of the ice wedges
399 (Clifford, 1993; Mellon and Jakosky, 1993, 1995).

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

406 **Stage 4a:** In conjunction with more recent high-obliquities a metres-deep icy and lightly-
407 toned dusty mantle(s) accumulates in our study region (e.g. Mustard et al. 2001; Head et al. 2003;
408 Forget et al. 2006; Madeleine et al. 2009). The mantle buries or mutes the thermokarstic
409 (polygonised) depressions and the polygonised terrain incised by the depressions (Soare et al.
410 2017).

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

416 **7. Discussion & Conclusion**

417 *7.1 What if...*

418 Questions concerning the possible existence of ice-rich terrain on Mars date back to the
419 grainy Mariner-mission photographs of the 1960s and 1970s (e.g. Belcher et al. 1971; Sharp et al.
420 1971). In the aftermath, data-based validation of this hypothesis has been fleeting. “Looks-like”
421 therefore “must-be” inferences of thermokarst-like landforms or of possible ice-wedge polygons
422 are suggestive, even beguiling, but inadequate.

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

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

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

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

443 7.2 Poleward latitude, ice-wedging and the freeze-thaw cycling of water

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

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

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

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

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

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

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

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790 **Figures**

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

799 **Fig. 2. a)** *HiRISE* image ESP_055038_2250 gridded by 500x500 metres squares (background:
 800 *MOLA* elevation data overlapping a *THEMIS* daytime-*IR* mosaic). **b)** Zoom on four
 801 500x500 metres-squares that illustrate the mapping method. Red and blue dots respectively
 802 locate *lcps* and *hcps*. North is up on all tiles. *HiRISE* image credit: *NASA/JPL/University*

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

805 **Fig. 3.** Examples of the three polygon-types, categorised by the relative elevation-differences
 806 between polygon centres and margins, and disparate locations, in our study region. **a)**
 807 Scalloped depressions overprinted completely by *lcps* (*HiRISE* image PSP_006908_2215,
 808 41.178° N, 124.130° E, 25cm/pixel). **b)** Ubiquitous distribution of *hcps* throughout
 809 scalloped terrain (*HiRISE* image ESP_046467_2280, 47.594° N, 119.679° E, 25cm/pixel).
 810 **c)** Proximal *lcps* (black arrows) and *hcps* (white arrows) (*HiRISE* image
 811 ESP_055038_2250, 44.757° N, 113.888° E, 50cm/pixel). **d)** Polygons with no observable
 812 elevation-difference between centres and margins (*HiRISE* image ESP_026450_2270,
 813 46.823° N, 110.880° E, 50cm/pixel). **e)** Larger-scale view of panel **d)** within the inter-crater
 814 plains. **f)** *Lcps* (black arrows) and *hcps* (white arrows) in close proximity (*HiRISE* image
 815 ESP_036366_2235, 42.939° N, 115.428° E, 50cm/pixel). **g)** Panel **f)** at a larger scale. Note:
 816 these polygons are in **Type 2** terrain (background: *THEMIS* day-*IR* controlled mosaic).
 817 North is up in all images. *HiRISE* image credits: *NASA/JPL/University of Arizona*.

818 **Fig 4. a)** Circular and individual rimless depressions (*HiRISE* image ESP_053205_2235,
 819 43.350° N, 117.223° E, 25cm/pixel). **b)** Elongated rimless-depression (*HiRISE* image
 820 ESP_026094_2250, 44.657° N, 111.415° E, 25cm/pixel). **c)** Coalesced rimless
 821 depressions; note the scalloping of margins (*HiRISE* image ESP_026094_2250, 44.657°
 822 N, 111.415° E, 25cm/pixel). Note the inward-facing tiers (black dashed lines). **d)** Wave-
 823 like depressions (*HiRISE* image ESP_035667_2295, 49.377° N, 115.622° E, 25cm/pixel).
 824 North is up on all tiles. *HiRISE* image credits: *NASA/JPL/University of Arizona*.

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

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

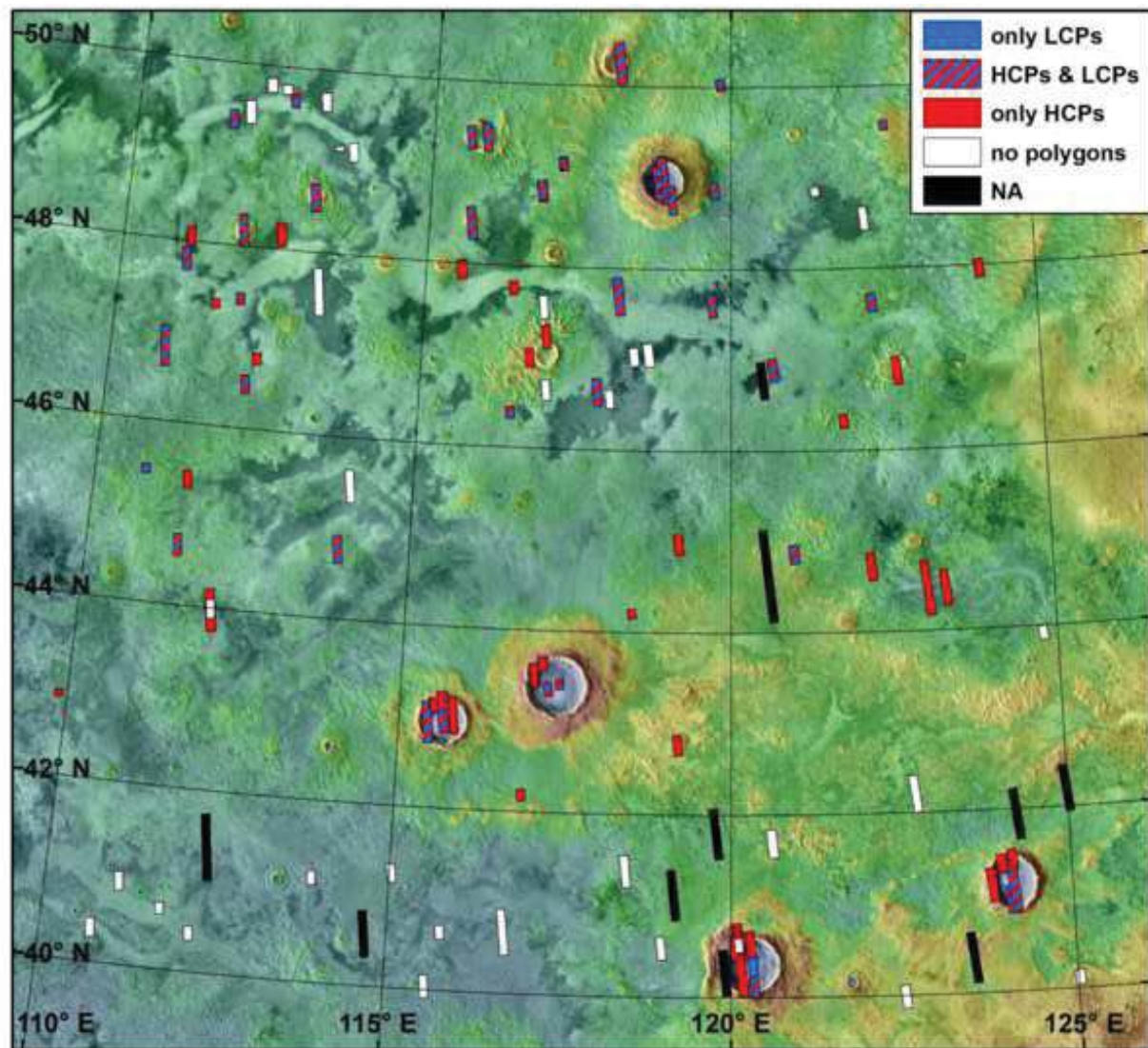
838 **Fig. 6. a)** Hummocky thermokarst, featuring thermokarst lakes, polygonised alases and beaded
 839 streams linking thaw pools of degraded ice-wedge polygons (Husky Lakes, Northwest
 840 Territories). **b)** Panoramic and cross-sectional view of an ice complex comprised of ice
 841 wedges (on the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

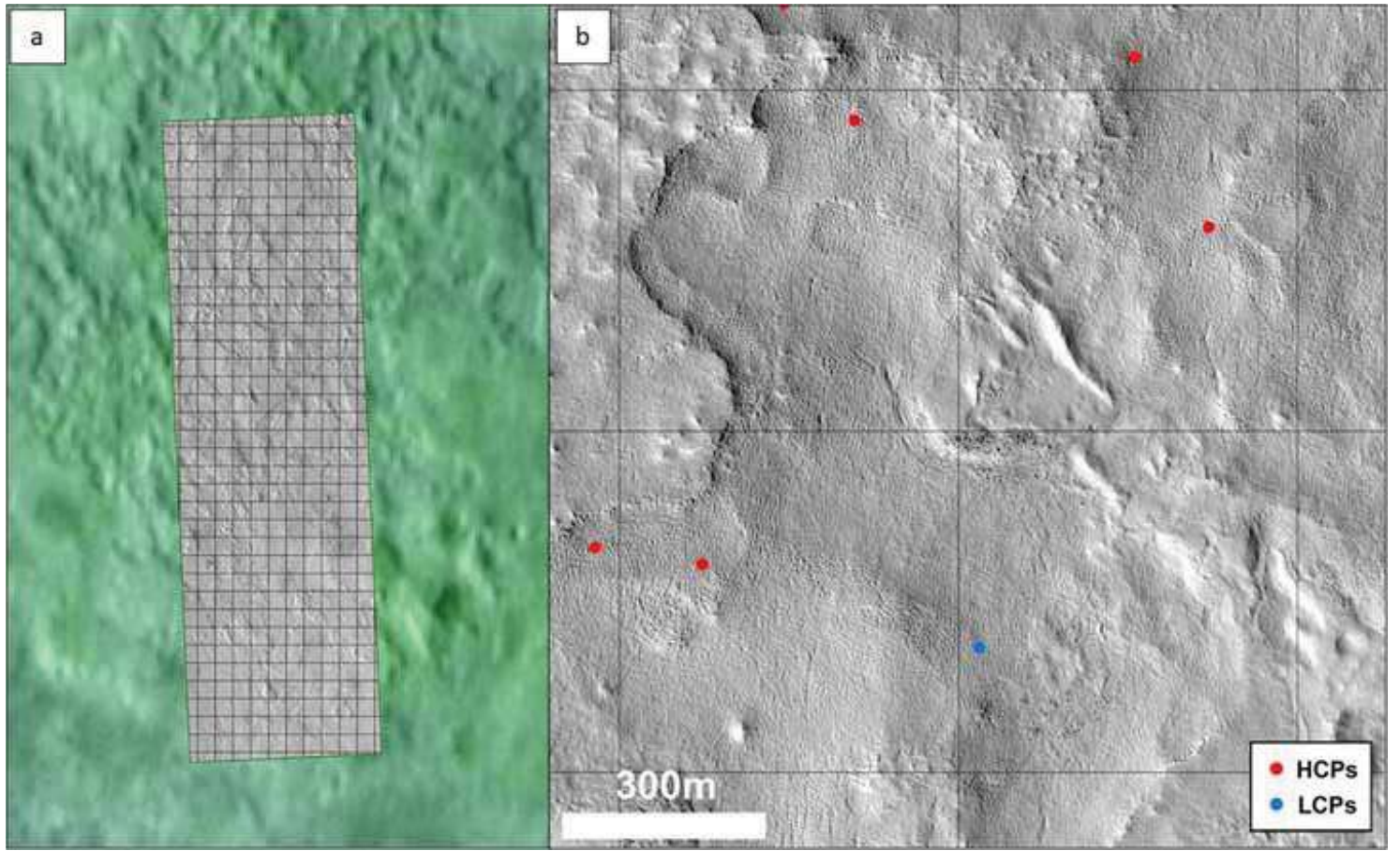
849 hand side of the terrace floor points to the recentness of thaw-based detachment as there is
 850 a gap in the overhang of the terrain immediately above the block location (Peninsula Point,
 851 Northwest Territories). Image credits: R. Soare.

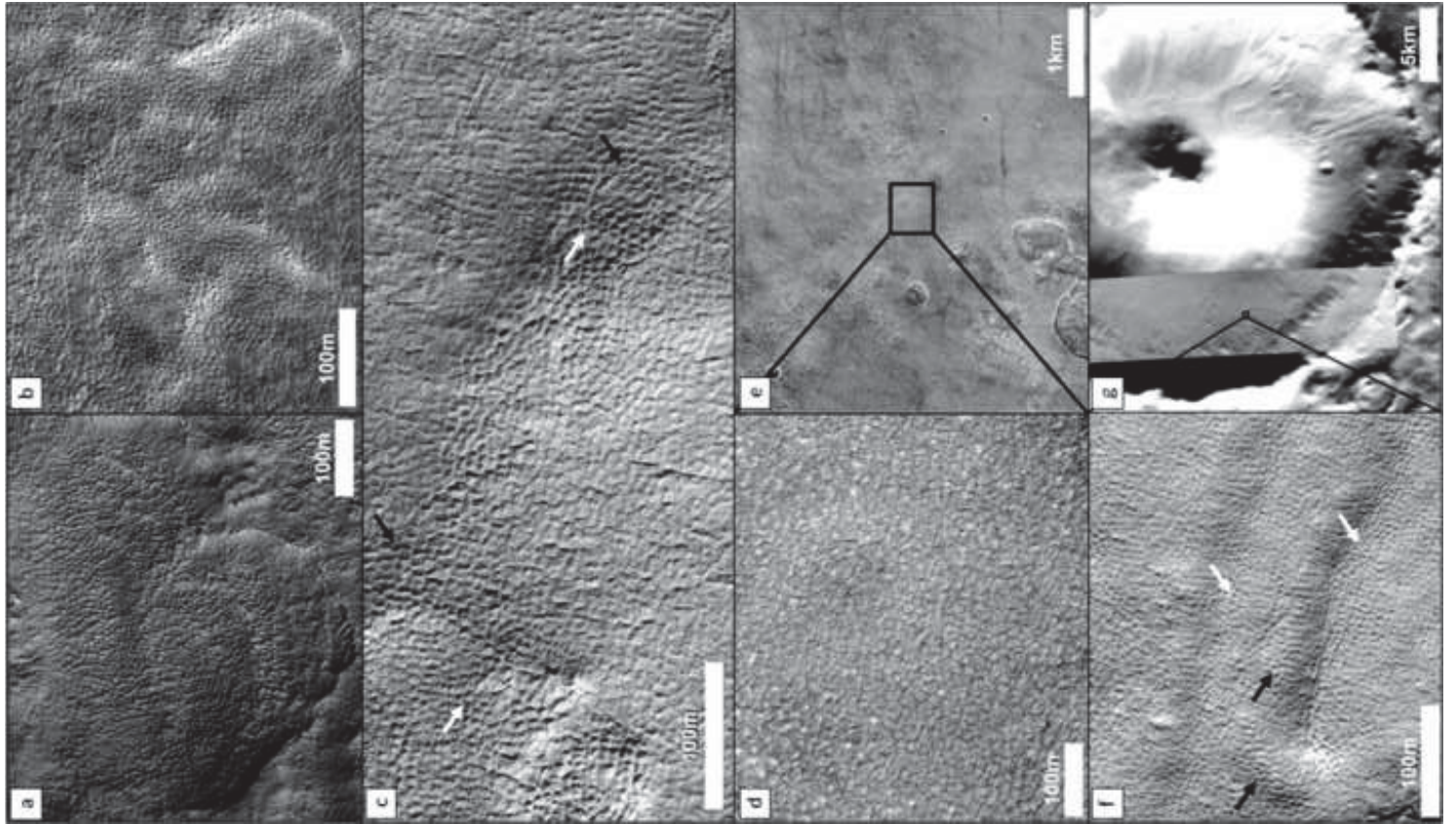
852 **Fig. 7.** Low-centred vs high-centred polygon latitudinal ratios for all terrain types. Data for all
 853 images have been binned for each degree of latitude excluding the upper bound. The black
 854 dashed line represents the linear correlation between lcp/hcp and latitude, its R^2 value
 855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

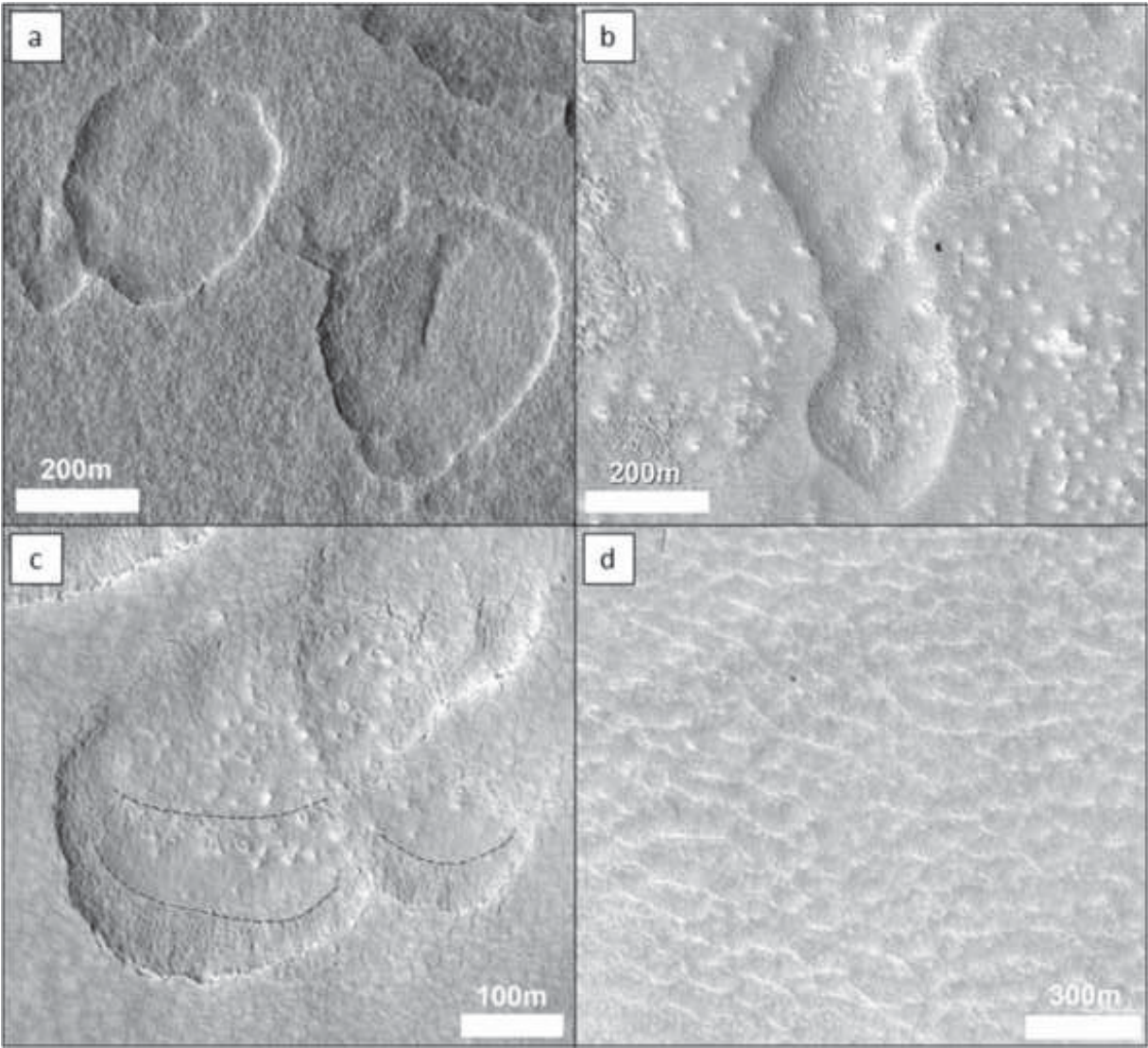
858 **Fig. 8.** Low-centred vs high-centred polygon latitudinal ratios (plains, excluding craters). Data for
 859 all images have been binned for each degree of latitude excluding the upper bound. The
 860 black dashed line represents the linear correlation between $lcps/hcps$ and latitude. The R^2
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

864 **Fig. 9. a)** Portion of *HiRISE* image ESP_026450_2270 within the study region (centered on
 865 46.823° N, 110.880° E) showing impact craters on the polygonized terrain. **b)** Cumulative
 866 unbinned and **c)** log-incremental root-2 binned crater size-frequency distribution for the
 867 entire *HiRISE* image (Area: 150 km^2) with absolute model age isochrons for 1 Ma, 10 Ma,
 868 and 100 Ma surfaces shown from Hartmann (2005). *HiRISE* image credit:
 869 NASA/JPL/University of Arizona.









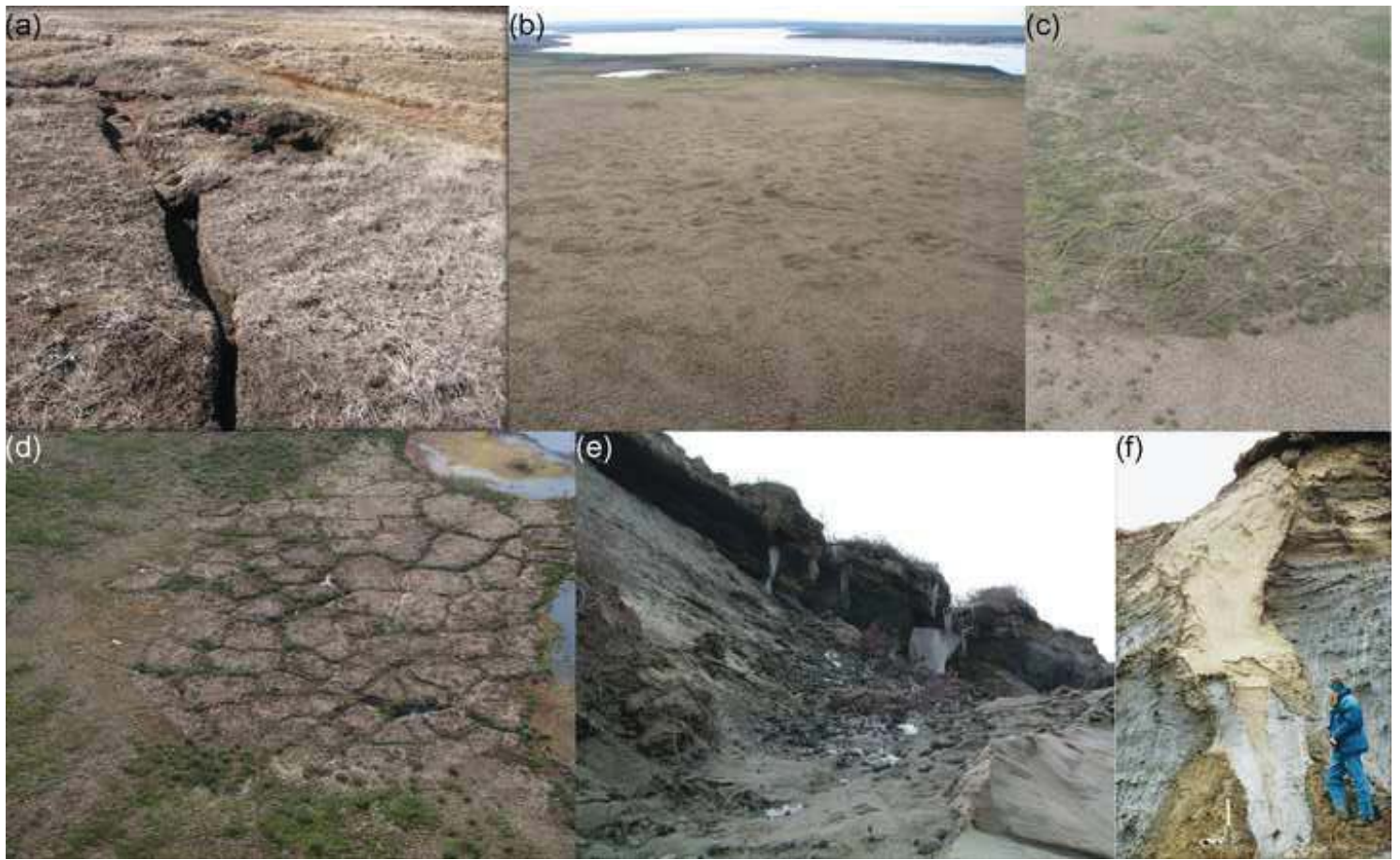




Fig. 7

