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1 Glacial and gully erosion on Mars: A terrestrial perspective

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15 Abstract

16 The mid- to high latitudes of Mars host assemblages of landforms consistent with a receding glacial
17 landscape on Earth. These landforms are postulated to have formed >5 Ma under a different climate
18 regime when Mars' orbital obliquity was on average 10° higher than today. Here, we investigate the
19 spatiotemporal relationship between gullies and glacial landforms, both common in the mid-
20 latitudes. Gullies are kilometre-scale landforms with a source alcove, transportation channel, and
21 depositional apron. The glacial landforms comprise (1) extant viscous flow features (VFF) that extend
22 from the base of crater walls into the interior of crater floors and are widely interpreted as debris-
23 covered glaciers containing extant ice, and (2) landforms such as arcuate ridges at the base of crater
24 walls that have been interpreted as relicts of more recent, less extensive glacial advances focussed
25 on crater walls. We measure headwall retreat associated with glacial landforms and date their host-
26 craters to constrain minimum headwall retreat rates. We record headwall retreat rates up to $\sim 10^2$ m
27 My^{-1} for the youngest suite of glacial landforms, equivalent to erosion rates of wet-based glaciers on
28 Earth and to headwall retreat rates associated with martian bedrock gully systems. We find
29 extensive evidence for a single erosional episode dating 5-10 Ma, which postdates emplacement of
30 the majority of VFF but seems to predate formation of the gullies. We propose that the wet-based
31 glacial episode was associated with glaciation focussed on the crater walls rather than melting of the
32 glacial ice deposits on the crater floors (VFF). This is consistent with our observations of crater wall
33 morphologies, including the presence of arcuate ridges consistent with terrestrial glaciotectionic
34 features that require liquid water to form, textural alteration of the eroded bedrock surface
35 consistent with ice-segregation and frost-shattering, and the presence of downslope pasted-on
36 terrain, tentatively interpreted here as glacial till deposits sourced from glacial erosion of the crater
37 wall. The pasted-on terrain is usually interpreted as a thicker, latitude-dependant mantle located on
38 sloping terrain formed from airfall of ice nucleated on dust, but we suggest that it has been
39 reworked by glaciation and is predominantly glacial in origin. Although our results cannot

40 substantiate that gullies are produced by meltwater, the discovery of this wet glacial event does
41 provide evidence for widespread meltwater generation in Mars' recent history.

42

43 *Keywords:* Mars; martian gullies; viscous flow features; glacier-like forms; liquid water

44 1. Introduction

45 The martian mid-latitudes are host to a suite of landforms that indicate significant geologically
46 recent (10s – 100s Ma) surface-atmosphere exchanges of water ice. This study focuses on two of the
47 most common landforms: martian gullies and glacier-like forms. We examine the role they have
48 played in landscape evolution over the last ten to hundreds of millions of years by using statistical
49 analysis of topographic data. In this introduction, we first present a brief overview of the state of
50 knowledge concerning the present and past distribution of ice and related landforms on Mars, then
51 specify how martian gullies fit into this context and finally present the scope of the present study.

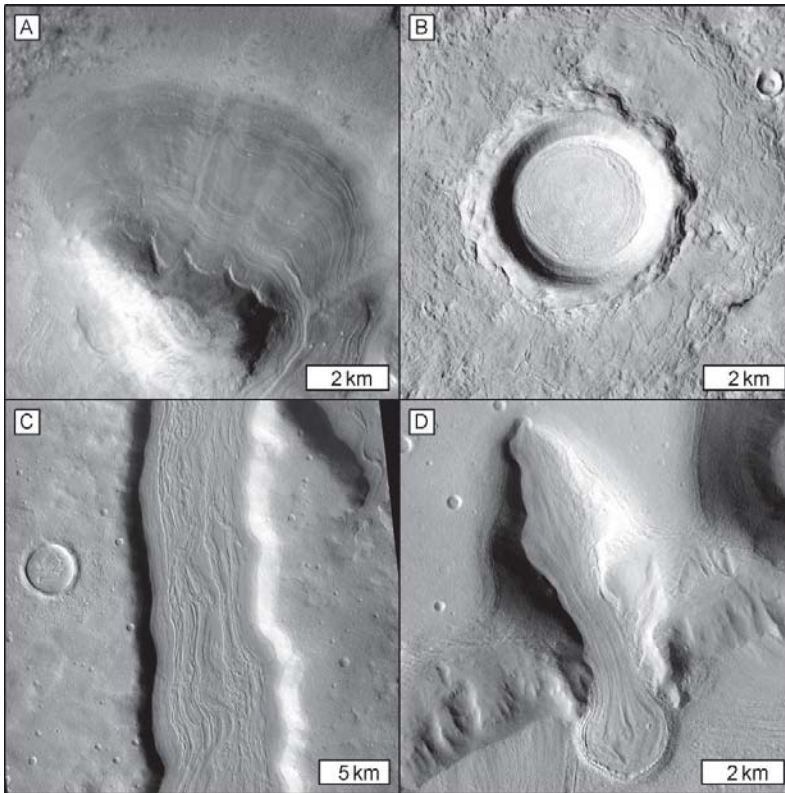
52 *1.1. The distribution of water ice on Mars*

53 Water ice is stable and exposed at the surface at the two polar caps of Mars, which each contain a
54 volume of ice similar to the Greenland ice sheet on Earth — $\sim 10^6 \text{ km}^3$ (Plaut et al., 2007; Putzig et al.,
55 2009; Bamber et al., 2013; Levy et al., 2014) . Water vapour is contributed to the atmosphere by
56 seasonal sublimation of the north polar cap, which has higher summer temperatures than the
57 southern cap because of its lower altitude and higher atmospheric pressure (e.g., Richardson and
58 Wilson, 2002). It also has a larger part of the water ice cap exposed at the surface compared to the
59 south, where the rest of the surface is partially hidden by a perennial, thin, CO_2 ice layer and by
60 surface debris (e.g., Thomas et al., 2000). Water vapour contributed to the atmosphere is
61 redistributed across the planet and can be deposited as surface frosts down to the mid-latitudes
62 (Svitek and Murray, 1990). Theoretical modelling predicts that ground ice on Mars should exist in
63 diffusive equilibrium with the atmospheric water vapour — it should be cold trapped into the pores
64 of the regolith (Mellon and Jakosky, 1993; Fisher, 2005). This idea is supported by observations from
65 the Neutron and Gamma Ray Spectrometers on Mars Global Surveyor that found abundant
66 hydrogen in the top metre of the regolith down to $\sim 50^\circ$ latitude in both hemispheres, which can be
67 explained by an ice content of 4% to $>64\%$ in the regolith (Boynton et al., 2002; Feldman et al., 2004;
68 Jakosky et al., 2005). This geophysical evidence is further supported by the observation of ubiquitous

69 polygonally patterned ground in the same latitudinal band, consistent with thermal contraction
70 cracks formed in ice-cemented soil over annual timescales (Mellon, 1997; Mangold, 2005; Levy et al.,
71 2009a; Schon et al., 2012). A trench dug by the Phoenix lander at 68°N found excess and pore ice
72 centimetres below the surface (as predicted by diffusive equilibrium models; Mellon et al., 2009),
73 newly formed impact craters have exposed water ice in their ejecta at latitudes down to 39° N
74 (Byrne et al., 2009; Dundas et al., 2014) and exposures of almost pure ice in eight ~800-m-high
75 scarps have been found at ~55° latitude (Dundas et al., 2018).

76 The latitude zone hosting ground ice is also an area that is smooth at scales of a kilometre
77 (Kreslavsky and Head, 2000), as measured from elevation data from the Mars Orbiter Laser Altimeter
78 (MOLA). This smoothing is partly attributed to the presence of the latitude-dependant mantle (LDM)
79 — a deposit thought to consist of the remnants of an airfall deposit of ice nucleated on dust, where
80 the dust forms a sublimation lag protecting the remaining deposits from sublimation. Mustard et al.
81 (2001) and Milliken et al. (2003) found that the LDM exhibited degraded or pitted textures at
82 latitudes between 30 and 50° N and S, this change in texture coinciding with the change in surface
83 roughness found from MOLA data (Kreslavsky and Head, 2000). The surface age of the LDM has been
84 estimated at ~0.1-5 Ma and decreases in age with increasing latitude (Kostama et al., 2006; Levy et
85 al., 2009b; Schon et al., 2012; Willmes et al., 2012). The LDM generally consists of alternating
86 relatively ice- and dust-rich layers, indicating multiple generations of deposits formed under varying
87 climatic conditions (Schon et al., 2009a). Various surface textures/draping deposits have been
88 grouped under the term LDM, but we will argue below (see also Soare et al., 2017) that not all of
89 these necessarily represent airfall dust-ice deposits. In total, LDM deposits cover 23% of the surface
90 of Mars (Kreslavsky and Head, 2002) and are thought to contain 10^3 - 10^4 km³ of ice (Mustard et al.,
91 2001; Kreslavsky and Head, 2002; Conway and Balme, 2014) .

92



93

94 **Fig. 1.** Viscous flow features on Mars in local sinusoidal projections, north is up in all panels. (A)
 95 Lobate debris apron in Deuteronilus Mensae, CTX image F22_044466_2278. (B) Concentric crater fill
 96 in Utopia Planitia, CTX image P15_007067_2124. (C) Lineated valley fill in the Arabia Terra region,
 97 CTX image F06_038017_2202. (D) Glacier-like form in Protonilus Mensae, CTX image
 98 G03_019358_2225.

99 Another significant reservoir of ice on Mars are the viscous flow features (VFF) found in the martian
 100 mid-latitudes (e.g., Squyres, 1979; Milliken et al., 2003; Levy et al., 2014). They have a total
 101 estimated volume of $4.2 \times 10^5 \text{ km}^3$, which is $\sim 20\%$ of the total volume of the polar caps (Levy et al.,
 102 2014). We will use VFF as an umbrella term for a range of landforms thought to be similar to debris-
 103 covered glaciers on Earth. They include the following (illustrated in Fig. 1):

- 104 • Lobate debris aprons (LDA) – these are smooth aprons that extend from and encompass
- 105 mesas. They can extend several to tens of kilometres from their mesa, and orbital radar
- 106 observations have confirmed that they can be up to 1 km thick and contain $>90\%$ ice (Holt et
- 107 al., 2008; Plaut et al., 2009). Their ages range from $\sim 40\text{-}500 \text{ Ma}$ (e.g., Mangold, 2003;

108 Morgan et al., 2009; Baker et al., 2010; Hartmann et al., 2014) to ~1 Ga (e.g., Levy et al.,
109 2007; Baker and Head, 2015; Berman et al., 2015). The LDAs are the largest reservoir of ice
110 among the VFFs (Levy et al., 2014).

- 111 • Crater interior ice deposits, including concentric crater fill (CCF) – these deposits fill and
112 smooth-over the floors of impact craters and are ubiquitous in the mid-latitudes (they
113 number >9000; Levy et al., 2014). Levy et al. (2014) estimated that ice thicknesses can range
114 up to 1.7 km. The surface of these fills often expresses lineations either concentric to the
115 crater walls (in this case they are called CCF) or can instead be oriented in one direction
116 (Dickson et al., 2012), and these orientations vary with latitude (pole-facing 30-45° and
117 concentric >45°). They have been dated to be as young as 10 Ma (Levy et al., 2009b) and as
118 old as 700 Ma (Fassett et al., 2014). Crater interior ice deposits can cover part or all of the
119 crater floor. Crater interior ice deposits are found in craters 2-72 km in diameter (Dickson et
120 al., 2012) and represent the second largest reservoir of ice of the VFFs.
- 121 • Lineated valley fill (LVF) – these infill existing valleys, being particularly abundant near the
122 martian dichotomy boundary. Their longitudinal ridges are thought to represent flow lines or
123 lateral compression from convergence of LDA on opposite valley walls (e.g., Squyres, 1978,
124 1979; Carr, 2001). Age estimates for LVF are similar to those for LDA: 100-500 Ma (e.g.,
125 Morgan et al., 2009). The LVF represents the third largest reservoir of ice among the VFFs.
- 126 • GLF, glacier-like-forms – these landforms are the most similar in scale and form to terrestrial
127 valley glaciers (Hubbard et al., 2011; Souness and Hubbard, 2012; Souness et al., 2012;
128 Brough et al., 2016a). They usually originate from large alcoves in escarpments that can be
129 on crater walls, valley walls, or mesas; are laterally confined to a valley/depression; and can
130 extend out onto plains. The GLF are on average ~4 km long and ~1 km wide, and the largest
131 examples are ~35 km long (Souness et al., 2012). We also include in this category lobate
132 forms, such as those described in Milliken et al. (2003), Berman et al. (2009), and Hartmann
133 et al. (2014), which are not topographically confined in their source area. Because of the lack

134 of superposed craters, these landforms are generally assumed to have an age of 10-100 Ma
135 and are thus younger than other VFF (Arfstrom and Hartmann, 2005; Berman et al., 2009;
136 Hartmann et al., 2014; Hepburn et al., 2018).

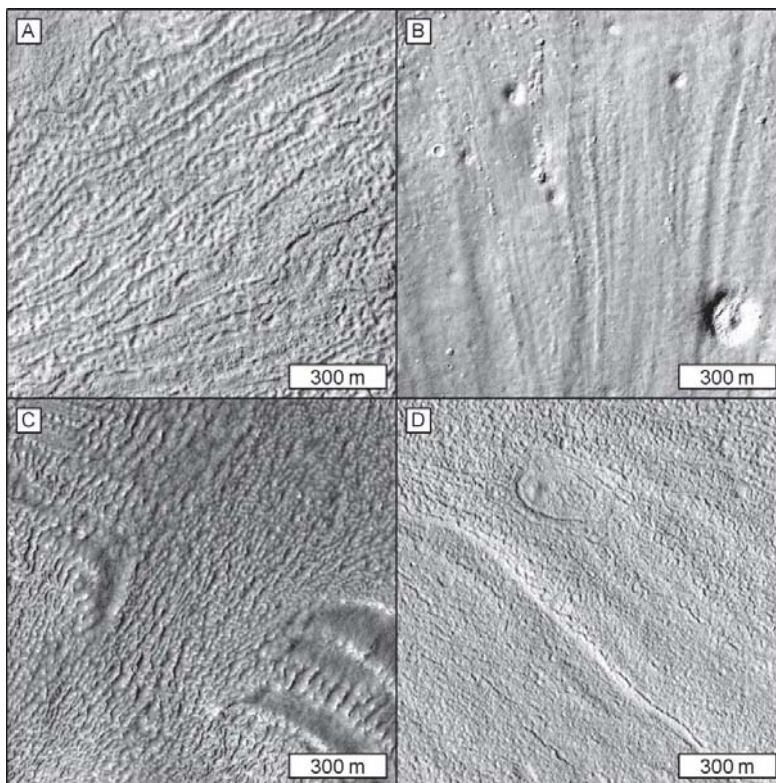
137 Apart from occasional radar evidence for water ice under the debris cover of VFFs (Holt et al., 2008;
138 Plaut et al., 2009), evidence for ice content is indirect and includes the presence of ablation textures
139 (fractures, pitted surface, inverted/deformed impact craters, including *ring mold* craters) (Kress and
140 Head, 2008), lineated surface topography thought to represent flow lines (e.g., Brough et al., 2016b),
141 and depositional landforms (e.g., terminal moraines) (Head et al., 2010; Brough et al., 2016b) (Fig.
142 2). The VFFs are believed to behave like cold-based glaciers on Earth, where the ice is frozen to the
143 bed and the flow is dominated by deformation of mass above by gravity-driven viscous creep
144 (Mangold and Allemand, 2001; Pierce and Crown, 2003; Li et al., 2005; Karlsson et al., 2015). Only in
145 two locations have eskers been linked to VFF, providing evidence of basal glacial melting (Gallagher
146 and Balme, 2015; Butcher et al., 2017). Rare supraglacial valleys have been attributed to transient
147 supraglacial melting encouraged by focussing of solar radiation onto VFF surfaces from steep
148 adjacent topography (Fassett et al., 2010).

149 The ice present in the ground (including LDM) and in VFF is believed to be in diffusive equilibrium
150 with the atmosphere, yet insufficient water vapour can be mobilised at the present day to explain
151 the presence of these large ice masses. The formation of VFF and LDM has been attributed to an
152 increase in average orbital obliquity, leading to transfer of large amounts of ice to the mid-latitudes
153 (Levrard et al., 2004, 2007; Madeleine et al., 2009). Mars' present axial obliquity is similar to that of
154 the Earth (25°), but in the past 5 Ma has oscillated between 15° and 30° (with a periodicity of
155 ~100 ka) and from 5 to 10 Ma between 28° and 45° (Laskar and Robutel, 1993; Laskar et al., 2004).
156 Seasonal temperature asymmetries increase with obliquity: at high obliquity, polar regions receive
157 increased insolation and the north polar cap is believed to have been completely destabilised around
158 4 Ma when the obliquity was on average higher (Levrard et al., 2007). The surface-atmosphere

159 exchange in water ice and CO₂ becomes more intense at higher obliquity, and surface ice and ground
160 ice are predicted to become stable even at equatorial latitudes.

161 *1.2. Gullies and their relation to ice on Mars*

162 The global distribution of martian gullies is spatially correlated with the distribution of water ice
163 landforms described above. Gullies occur across the same range of latitudes as VFF, but they are less
164 common where high concentrations of LDA and GLF occur and are only found in 12% of craters with
165 interior ice deposits (Conway et al., 2018a). They are strongly linked to the presence of LDM (Levy et
166 al., 2011; e.g., Dickson et al., 2015), although in the case of gullies the textures of this slope-side
167 surficial deposit, into which they incise, are somewhat different to the LDM found in the plains;
168 hence in this contribution we will refer to this unit using the term first coined by Christensen (2003)
169 in relation to martian gullies: *pasted-on terrain*.



170
171 **Fig. 2.** Surface textures of viscous flow features on Mars at scales of 1:15,000. North is up in all
172 panels and lighting is from the left (east). (A) Surface texture of crater interior ice deposits showing
173 surface lineations comprising alternating discontinuous ridges and chains of elongated pits in HiRISE

174 *image ESP_028651_1370. (B) A ring-mold crater (bottom right) on the surface of a crater interior ice-*
175 *deposit together with muted craters and normal craters, also showing surface lineations in HiRISE*
176 *image ESP_046308_1425. (C) A lobate debris apron in Deuteronilus Mensae showing typical*
177 *contorted raised and pitted surface patterns in HiRISE image ESP_018515_2225. (D) Lineated valley*
178 *fill in the Acheron Fossae region with pitted and cracked textures and a deformed impact crater (the*
179 *valley is oriented NW-SE) in HiRISE image ESP_016266_2165.*

180 Gullies are typically kilometre-scale landforms found on steep slopes in the mid- to high latitudes of
181 both hemispheres ($\sim 30^\circ$ to the polar caps; Harrison et al., 2015; Conway et al., 2017). They comprise
182 a tributary source alcove, transport channel, and depositional apron (Malin and Edgett, 2000). They
183 are found primarily on pole-facing slopes at latitudes between 30° and 40° and then mostly on
184 equator-facing slopes poleward of 40° (but they can also occur on pole-facing slopes in this latitude
185 interval). The relative lack of gullies in regions with no evidence for ground ice (between 30° N and
186 30° S), and their trends in orientation with latitude have led authors to conclude that changes in
187 orbital obliquity are required to explain their present-day distribution (e.g., Costard et al., 2002; de
188 Haas et al., 2015a). However, the process by which gullies form is under strong debate. Present-day
189 activity in gullies is thought to be brought about via slope instability caused by sublimation of carbon
190 dioxide (Diniaga et al., 2010; Dundas et al., 2010, 2012, 2015, 2017; Raack et al., 2015; Pasquon et
191 al., 2016), and the distribution of surface carbon dioxide ice is strongly controlled by the presence of
192 (water) ground ice (Vincendon et al., 2010). However, whether such a process alone can explain the
193 formation of the whole landform remains uncertain. In particular, equator-facing slopes at the
194 present-day and in the past are not favourable locations for deposition, and thus sublimation, of CO_2
195 ice (Conway et al., 2018a). Formation by liquid water is consistent with the morphology and
196 morphometry of gullies (e.g., Malin and Edgett, 2000; Conway and Balme, 2016), yet liquid water is a
197 poor candidate to explain the present-day activity and can only be invoked under certain specific
198 conditions in the past (Costard et al., 2002; Richardson and Mischna, 2005; Williams et al., 2008,

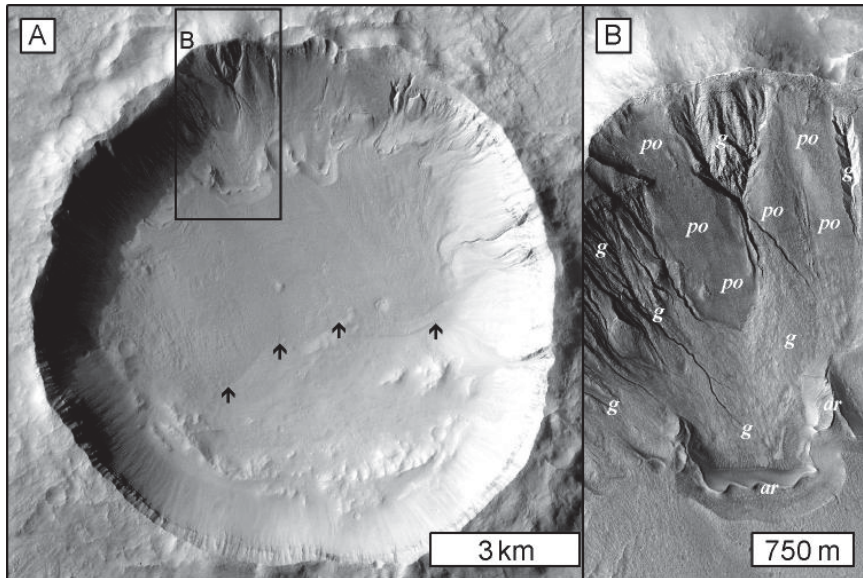
199 2009). Hence, understanding the interaction of gullies with other landforms in time and space has
200 the potential to break this impasse.

201 Our knowledge on the timing of the activity and the overall age of gully systems is relatively limited.
202 We know that gullies on martian dunes have such frequent and voluminous activity that their full-
203 extent can be explained on a timescale of hundreds of years (Pasquon et al., 2017). Dundas et al.
204 (2010, 2012) estimated that mid-latitude gullies (not formed in sand but in other materials, including
205 bedrock and mantling materials) could be formed within the last few millions of years given the
206 currently observed magnitude and frequency of their activity. Non-sand gullies also have
207 morphological evidence for multiple episodes of activity — in one case authors have ascertained the
208 age of part of a gully fan where other parts of the fan pre- and postdate the determined age of ~1.25
209 Ma (Schon et al., 2009b). Dickson et al. (2015) found common evidence of relict gullies, being
210 exhumed from and buried by *LDM* (pasted-on terrain), showing that gully activity must extend into
211 epochs when this deposit was being formed/removed (estimated to be ~0.1-5 Ma; Kostama et al.,
212 2006; Willmes et al., 2012) .

213 De Haas et al. (2015b) found that gully alcoves had a similar size independent of the age of the host
214 crater, and de Haas et al. (2018) proposed that gullies could have their erosion *reset* during glacial
215 epochs. In the youngest craters where gullies are found emanating from well-defined bedrock
216 alcoves cutback into the crater rim, de Haas et al. (2015b) estimated headwall retreat rates of 10^{-4} to
217 10^{-1} mm y^{-1} , which correspond to the backweathering rates experienced on rockwalls in Arctic,
218 Nordic, and Alpine environments on Earth. De Haas et al. (2018) classed a crater as *glaciated* if it
219 contained one or more of the following morphological features: arcuate ridges and/or spatulate
220 depressions at the foot of the slope (Fig. 3; Arfstrom and Hartmann, 2005; Berman et al., 2005; Head
221 et al., 2008; Hubbard et al., 2011; Jawin et al., 2018), floor-filling CCF indicated by concentric ridges
222 and/or pitted textures within identifiable in-crater deposits (e.g., Levy et al., 2010; Dickson et al.,
223 2012), or crater-scale downslope lobate features that diverge around obstacles on the crater floor

224 and have longitudinal or transverse ridges and/or pitted textures on their surfaces (Fig. 2; e.g., Baker
225 et al., 2010; Head et al., 2010). Several authors (e.g., Berman et al., 2005, 2009; Head et al., 2008;
226 Hauber et al., 2011a; Schon and Head, 2011, 2012; Dickson et al., 2015; de Haas et al., 2018; Jawin et
227 al., 2018) have suggested that these glacial landforms represent different stages of ice-deposition
228 and removal (glacial-interglacial cycles) during martian obliquity cycles. The VFFs were able to form
229 during glacial epochs when average obliquity was higher than today (>5 Ma). Since the last
230 formation of VFF, LDM (pasted-on terrain) was deposited during smaller ice ages, which occurred
231 during obliquity peaks in the present, lower-average obliquity regime (e.g., Mellon and Jakosky,
232 1995; Head et al., 2003; Madeleine et al., 2014). Many gullies that are present on Mars today have
233 formed into VFF or LDM remnants (e.g., Dickson et al., 2015; de Haas et al., 2018), and thus their
234 formation is strongly linked to VFF and LDM (pasted-on terrain).

235 Berman et al. (2005) and Head et al. (2008) noted that the rims of craters with arcuate ridges and
236 gullies on their pole-facing walls had a lower elevation than equator-facing walls of the same craters.
237 The model proposed by de Haas et al. (2018) hypothesise that evidence of gully erosion (alcoves and
238 channels) is removed from crater walls by glacial erosion and/or buried by glacial deposits after
239 sublimation/melting. Further, Levy et al. (2016) estimated that VFF can be responsible for 10^{-5} - 10^{-4}
240 mm γ^{-1} of erosion. However, no study to date has attempted to directly assess the amount of
241 erosion associated with the gullies and glacial landforms found on crater walls.



242

243 **Fig. 3.** Niquero Crater on Mars, an example with a crater interior ice deposit, arcuate ridges, pasted-
 244 on terrain and gullies. North is up in all panels. (A) Overview of crater in CTX image
 245 P03_002383_1417. Arrows indicate the southernmost extent of the crater interior ice-deposit. (B)
 246 Detail of gullies (labelled as 'g') and arcuate ridges (labelled as 'ar') within a spatulate depression
 247 with the presence of pasted-on terrain labelled as 'po'. HiRISE image ESP_030021_1410.

248 1.3 Objectives and structure

249 Here we use high resolution topographic data and dating of host craters to assess the amount and
 250 rate of erosion that these small-scale crater wall glaciers may have been responsible for and use
 251 analogy with glacial systems on Earth to assess by what mechanisms this erosion could have
 252 occurred. In our approach we use the consistency and coherence of the landforms and their spatial
 253 and topographic properties and relationships to go beyond analogy using the similarity of
 254 appearance of single landforms (Baker, 2014, 2017). We also compare to erosion rates in gullied
 255 craters without pasted-on terrain to assess whether glaciation is capable of erasing gullies on Mars.

256 This paper is structured as follows. First, we explore the link between the pasted-on terrain (into
 257 which gullies often incise) and glaciation by undertaking a global survey and highlighting key
 258 relationships and observations. Second, we undertake a detailed topographic analysis of a small

259 subset of these craters and of gullied craters to ascertain the amount and rate of erosion of their
260 walls and its link to previous glaciation. Third, we consider the morphological and topographical
261 evidence gathered on Mars in light of the present knowledge of glacial erosion on Earth in order to
262 build a coherent picture of the degradation of crater walls on Mars. Finally, we bring together
263 knowledge from the martian literature to assess the consistency of our new hypothesis within the
264 context of previous findings.

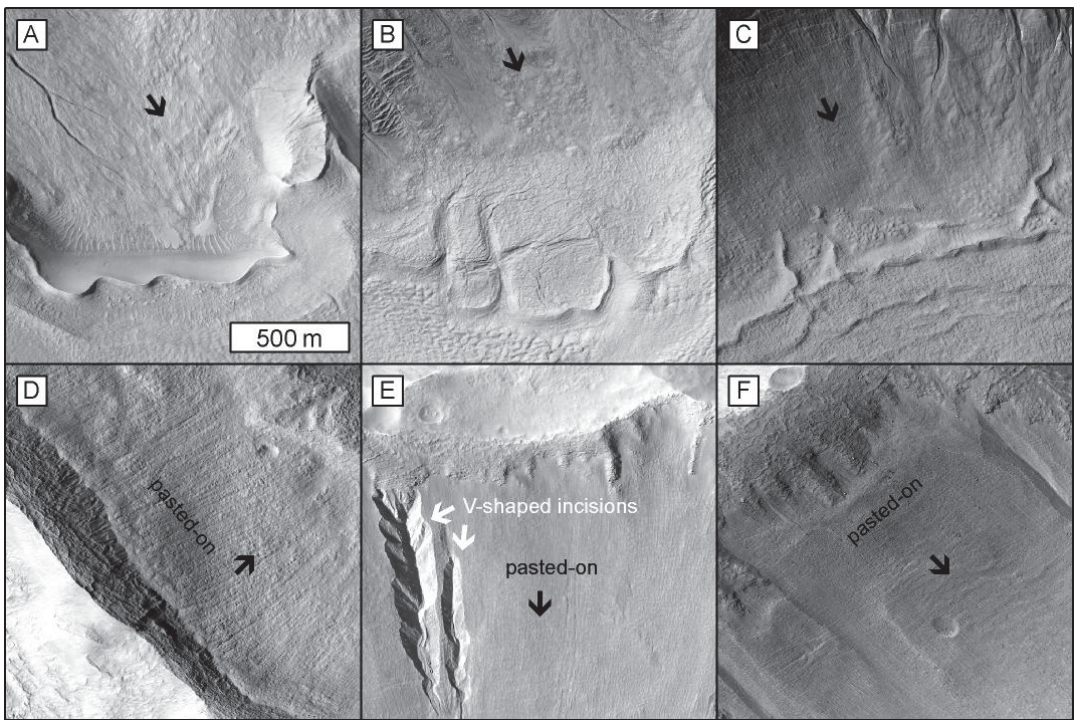
265 2. Global distribution and thickness of pasted-on terrain

266 2.1. Approach

267 We performed the majority of our morphological observations on NASA Mars Reconnaissance
268 Orbiter High Resolution Imaging Science Experiment (MRO HiRISE) images at 25-50 cm/pixel taken
269 up to April 2017. We used the Harrison et al. (2015) gully-distribution database as the basis for our
270 initial survey, down-selecting to HiRISE images that were in the vicinity of gullies. Within this
271 database we noted the occurrence of arcuate ridges associated with gully incisions and made
272 observations on the nature of the pasted-on deposits associated with the gullies. We also searched
273 for examples of craters without gullies but with pasted-on terrain and/or arcuate ridges by
274 examining craters <10 km from the Robbins and Hynes (2012) database with (projection corrected)
275 slopes of >20° derived from the MOLA-gridded data that had overlapping HiRISE images. We
276 performed a random sample of this data set and many craters that do not have gullies but have
277 pasted-on terrain and/or arcuate ridges will have been missed. Hence, this data set was not used to
278 derive any global distribution statistics but to highlight pertinent relationships.

279 We recognised arcuate ridges in HiRISE images using the following criteria: a sinuous to highly
280 arcuate ridge located at the foot of the crater wall and somewhat parallel to the crater rim (Figs. 4A-
281 C). We identified pasted-on terrain as being a draping unit that smoothed the appearance of the
282 crater wall at scales of ~1:25,000 with an upper lobate boundary (Figs. 4D-F). Gullies incised into this
283 unit have a distinctive V-shaped cross section in their mid-section, and these gullies are often

284 completely sourced within this unit, having a simple single elongate incision (Fig. 4E; Aston et al.,
 285 2011; Conway and Balme, 2014). It is worth noting that not all gullies have such V-shaped incisions.
 286 We previously performed measurements on HiRISE digital terrain models (DTM) and determined
 287 that the slopes of these incisions were on average 20° and ranged between 15° and 30° (Conway and
 288 Balme, 2014). Assuming a 20° slope, these incisions can be used as an estimate of the minimum
 289 depth of the pasted-on terrain simply by measuring the wall-length of the incision (Conway et al.,
 290 2015). We undertook a systematic survey of the incision wall-lengths in our HiRISE database in order
 291 to assess any trends in the thickness of pasted-on deposits with latitude. For craters hosting multiple
 292 gully systems, we measured the deepest incision because the depth of this incision should be closest
 293 to the total thickness of the pasted-on terrain.



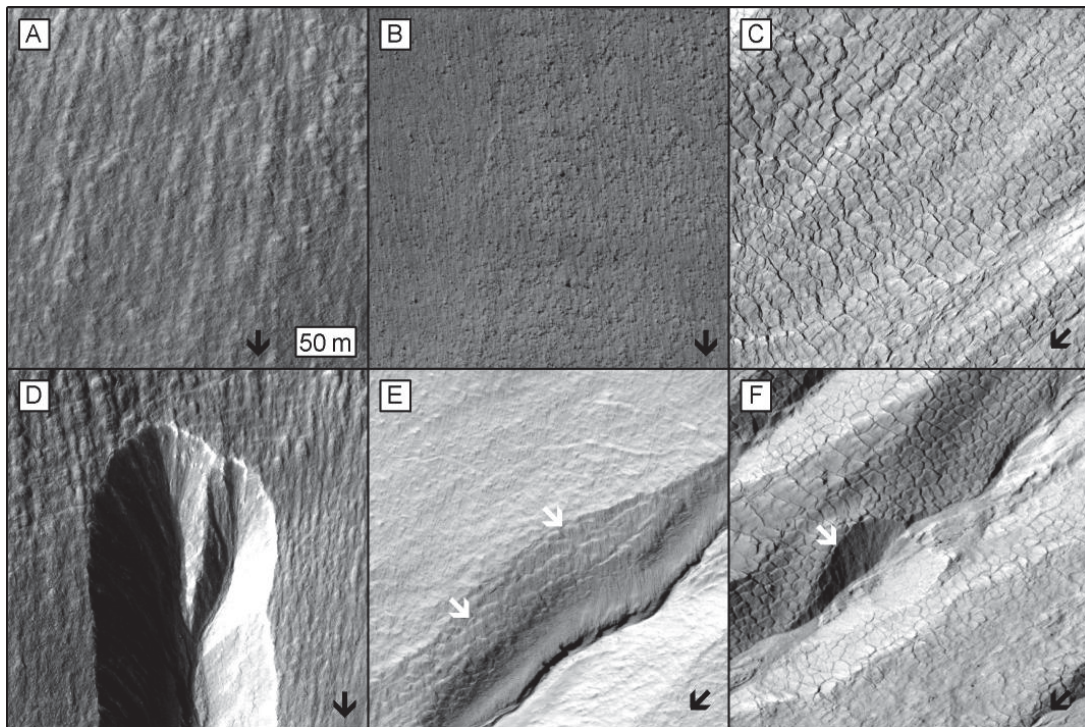
294
 295 **Fig. 4.** Recognition of arcuate ridges and pasted-on terrain in HiRISE images at 1:25,000, north is up
 296 in all panels, the downslope-direction is indicated by the black arrows and the scale for all panels is
 297 as indicted in panel (A). (A-C) Arcuate ridges in HiRISE images: ESP_038236_1410, ESP_023809_1415
 298 and ESP_020051_1420. (D-F) Pasted-on terrain in HiRISE images: ESP_027477_2170,
 299 ESP_038236_1410, and ESP_013858_1405. Note in panel (E) the sharply incised gullies with V-

300 *shaped incisions to the left of the panel, where the right-hand one of the pair is completely within the*
301 *pasted-on terrain.*

302 *2.2. Pasted-on terrain*

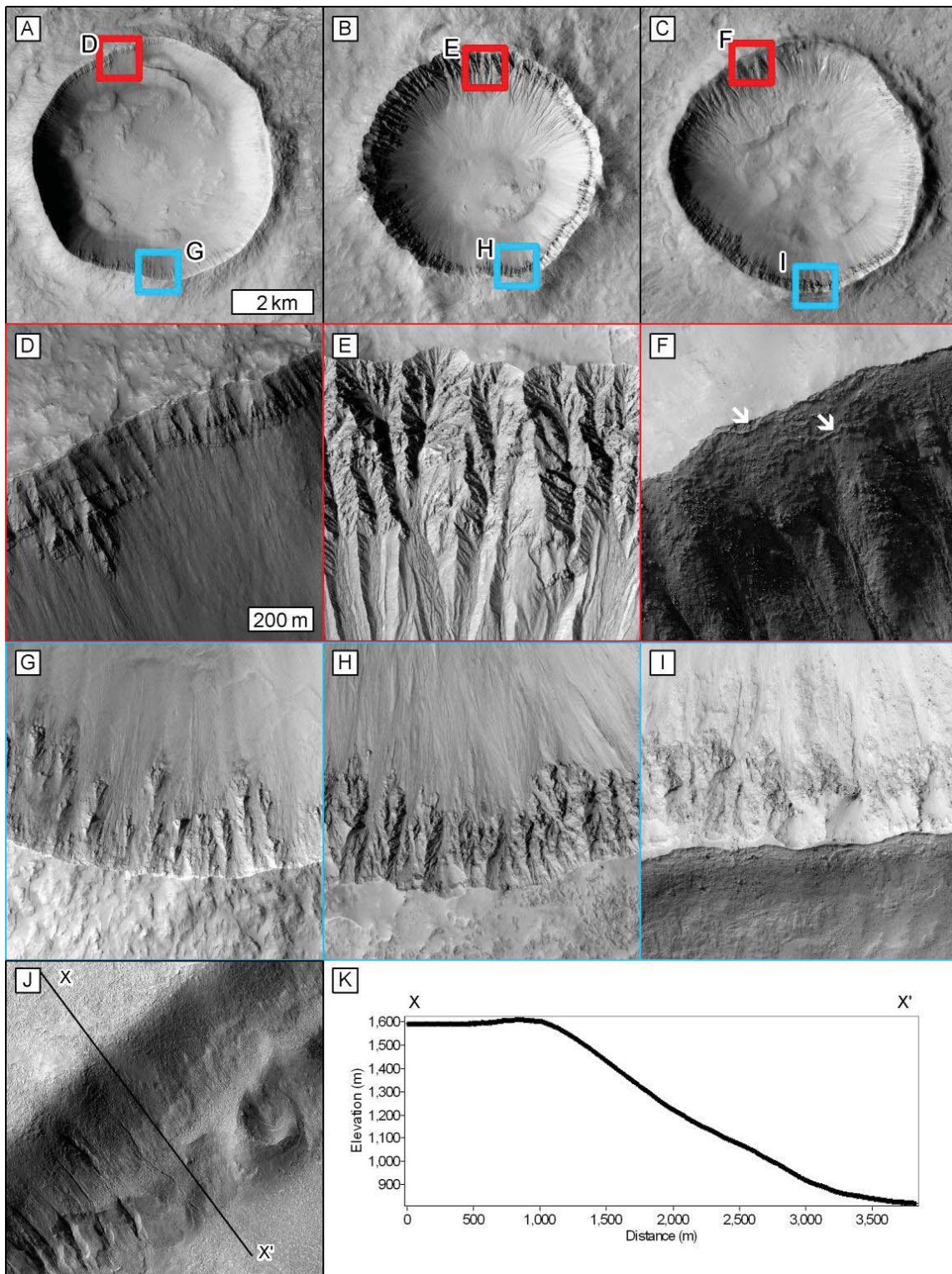
303 The surface texture of the pasted-on terrain often appears smooth and uniform at scales of
304 $\sim 1:25,000$, yet at scales of $\sim 1:5000$ or less we were able to identify three principal textures:
305 downslope lineated (Fig. 5A), smooth (Fig. 5B) and polygonised (Fig. 5C). Polygonal textures were
306 more often most clearly expressed within the gully incisions, as previously noted in the literature
307 (Levy et al., 2009c; Soare et al., 2014) (Fig. 5D). As noted by Conway and Balme (2014), the lineations
308 are cut by the gully incisions (Fig. 5E) and fresh incision walls have no polygonal textures and sharp
309 breaks in slope, whereas incisions with polygonal textures are shallower and have a more subdued
310 break in slope (Fig. 5F). We found that the pasted-on terrain is systematically associated with a
311 change in the appearance of the crater rim compared to the opposite crater rim (Fig. 6F vs. 6I) and
312 compared to crater rims of fresh equatorial craters (Watters et al., 2015; Tornabene et al., 2018)
313 (Fig. 6F vs. 6D, G). Bedrock is often exposed in the top few hundred metres of crater walls and is
314 particularly evident in young impact craters (Figs. 6D, G). It usually appears as a series of sub parallel
315 relatively massive beds and, particularly in fresh equatorial impact craters, displays a *spur-and-gully*
316 morphology with a series of regularly spaced shallow alcoves. The rim crest of these craters remains
317 relatively linear despite the presence of these alcoves (Figs. 6D, G). In comparison, the crater wall
318 immediately above the upslope boundary of pasted-on terrain does not have a spur-and-gully
319 morphology but is rather planar with a distinctive *mottled* texture (we refer to this as *texturally*
320 *altered bedrock* in the rest of this paper Fig. 6F). The rock exposure is rather patchy with no clear
321 massive bedrock outcrop, and the surface is often dominated by loose-appearing metre- to
322 decametre-scale boulders. Discontinuous rim-parallel lineations (Fig. 6F) are sometimes apparent
323 with no clear relationship to the underlying bedrock structure. Where texturally altered bedrock is
324 present, the crater rim has very little deviation from circular in planform (Fig. 6C), and sometimes
325 the crest is sharp while it can also be rounded (Figs. 6J, K), particularly in older craters. This texture

326 has been previously associated with erosion of crater walls (diameter <35 km) by lobate debris
 327 aprons in Alba Mons by Sinha and Vijayan (2017). This contrasts with the appearance of the rims of
 328 craters with gullies and no pasted-on terrain, which maintain the massive bedrock appearance, and
 329 have rim-traces that are crenulated with greatly accentuated spur-and-gully morphology compared
 330 to unmodified crater walls as noted in de Haas (de Haas et al., 2015b, 2018) (Figs. 6B,E).



331
 332 **Fig. 5.** Examples of pasted-on terrain textures, each image has the same scale as indicated in panel
 333 (A) at 1:5000. Black arrows indicate the downslope direction. (A) Lineated texture of the pasted-on
 334 terrain on the south-facing wall of Bunnik Crater, HiRISE image PSP_002514_1420. (B) Smooth
 335 texture of the pasted-on terrain, with rocks visible at the surface in HiRISE image ESP_033173_1400.
 336 (C) Polygonised texture of the pasted-on terrain in HiRISE image ESP_011672_1395. (D) Lineated
 337 texture of the pasted-on terrain cut by gully incision in Bunnik Crater, where polygons are weakly
 338 visible in the sun-facing wall of the incision, HiRISE image PSP_002514_1420. (E) Polygons visible
 339 inside the gully incision (white arrows) but not on the surrounding pasted-on terrain in HiRISE image
 340 ESP_033173_1400. (F) Sharp gully incision whose walls are not polygonised cutting through the

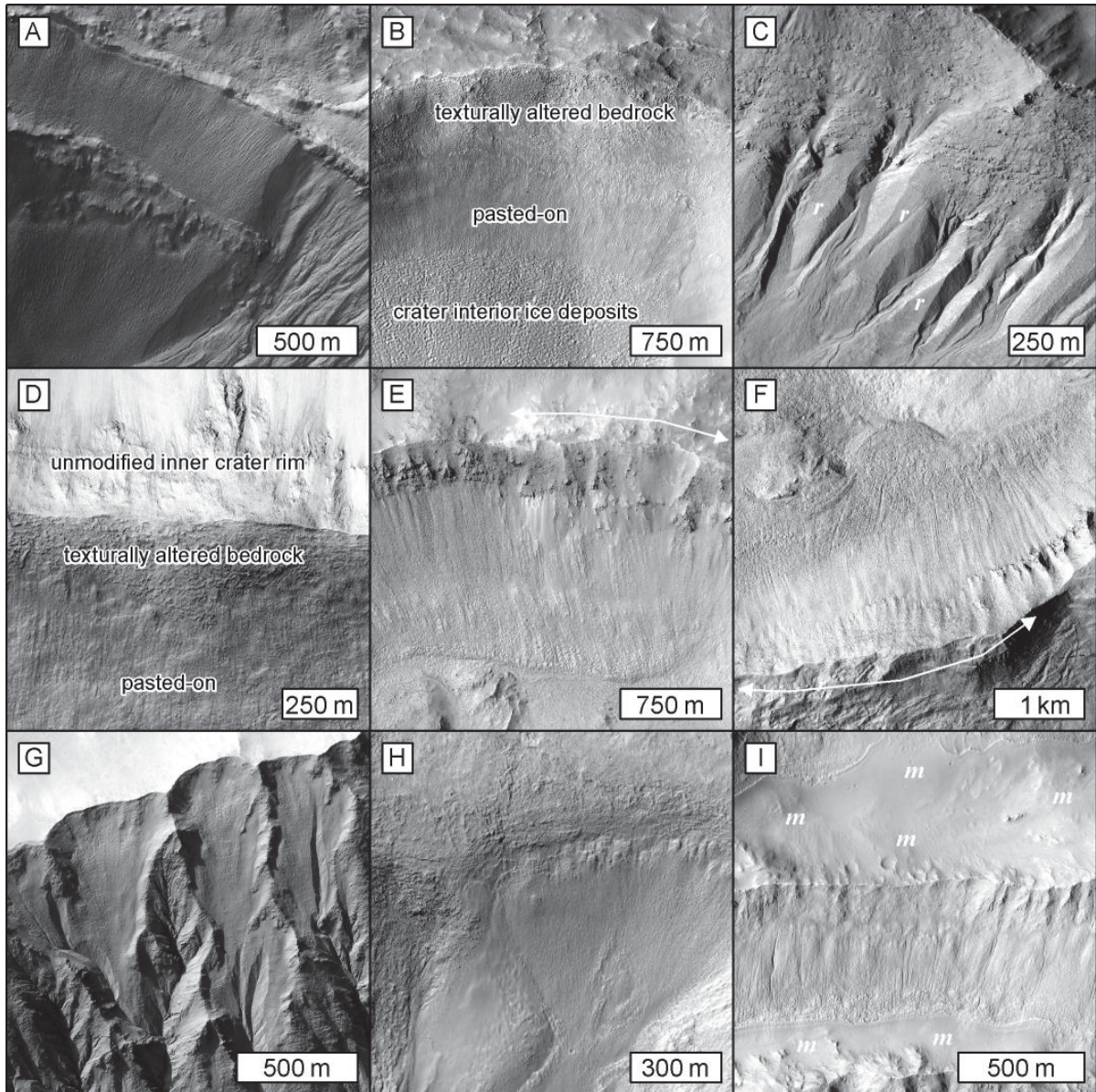
341 polygonised pasted-on terrain (white arrow). Polygonal patterns are again present on more gently
 342 sloping incisions. HiRISE image ESP_011672_1395.



343

344 **Fig. 6.** Rock exposures in fresh equatorial impact craters (left panels), gullied impact craters (middle
 345 panels), and craters with pasted-on terrain (right panels). The scale in panels (A-C) is given in panel

346 (A) and is 1:120,000 and the scale in panels (D-I) is given in panel (D) and is 1:15,000. North is up in
347 all panels. All craters are in the southern hemisphere. Red panels show the pole-facing slopes and
348 blue panels show the equator-facing slope. (A,D,G) Kenge Crater, a fresh equatorial crater from the
349 database of Tornabene et al. (2018) in (A) CTX image B07_012315_1635 and (D,G) HiRISE image
350 ESP_011893_1635. (B,E,H) Galap Crater with gullies eroded into the bedrock in (B) CTX image
351 B09_012971_1421 and (E,H) HiRISE image PSP_003939_1420. (C,F,I) Nybyen Crater with pasted-on
352 terrain and gullies in (C) CTX image G09_021563_1427 and (F,I) HiRISE image PSP_006663_1425.
353 Discontinuous ridges are highlighted by white arrows. (J,K) A crater in the Newton Basin with a
354 planed-off rim in (J) HiRISE image PSP_002620_1410 and (K) elevation profile taken from the
355 publically available DTM DTEEC_002620_1410_002686_1410_A01.



356

357 **Fig. 7.** Configuration of pasted-on terrain within craters on Mars. North is up in all panels. (A) Pasted-on
 358 on terrain extending to within ~20-50 m of a crater rim in HiRISE image ESP_034363_1380. (B)
 359 Pasted-on terrain covering only the lower 1/2-1/3 of the inner crater wall in HiRISE image
 360 ESP_038157_1415. (C) Pasted-on terrain as remnants between gully systems (labelled by white 'r') in
 361 HiRISE image ESP_011672_1395. (D) Pasted-on terrain on the exterior wall of an impact crater in
 362 HiRISE image ESP_023809_1415. (E,F) Pasted-on terrain where only some of the bedrock has been
 363 texturally altered above it (extent indicated by arrowed line) and the rest still has alcoves. HiRISE
 364 images ESP_014400_1525 and PSP_009164_2140, respectively. (G,H) Pasted-on terrain extending
 365 upward into wide alcoves on crater walls and leaving spurs of bedrock with an unmodified texture in

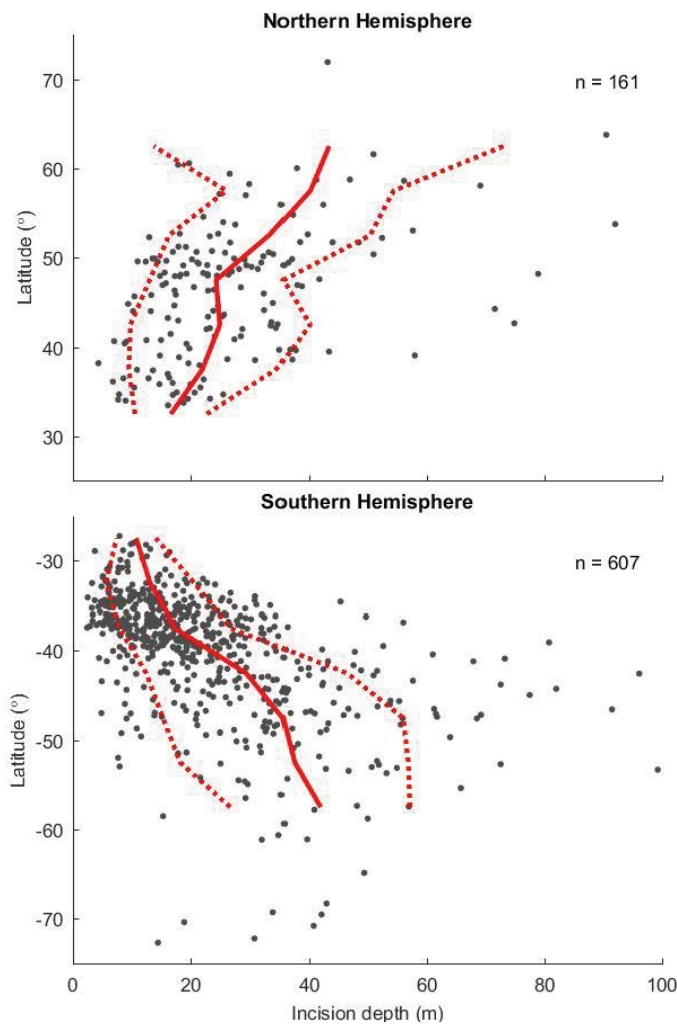
366 *panel (G). HiRISE images PSP_006629_1425 and ESP_033398_1420 respectively. (I) Pasted-on terrain*
367 *that appears to emerge from beneath mantle material (marked by 'm') with degradation textures as*
368 *described by Mustard et al. (2001) and Schon et al. (2009a), HiRISE image ESP_011839_1460.*

369 The pasted-on terrain can be found extending nearly all the way up to the top of the crater rim (Fig.
370 7A), or only on the lower half of the crater wall (Fig. 7B), or as remnants between extensive gully
371 systems (Fig. 7C). Pasted-on terrain can be found on the interior (Figs. 7A-C) and exterior crater walls
372 (Fig. 7D) and is almost exclusively associated with an altered texture of the upslope bedrock. In two
373 rare circumstances, texturally altered bedrock is found only on part of the wall with pasted-on
374 terrain (illustrated in Figs. 7F-I). The first of these circumstances involves very young craters (as
375 attested by few superposing impact craters <1 Ma, Table 1), where pasted-on terrain is found in the
376 absence of any other evidence of glaciation (e.g., crater interior ice deposits, arcuate ridges, large
377 lobate tongues, glacial textures) and also occurs without gullies. In the two cases that we found (one
378 in the northern and one in the southern hemisphere; Figs. 7E, F) alcoves in texturally unaltered
379 bedrock are present above some of the pasted-on terrain; but where the pasted-on terrain extends
380 beyond the foot of the crater wall, the alcoves are reduced and textural alteration has occurred. In
381 the second circumstance, pasted-on terrain can extend upward into wide (hundreds of metres)
382 alcoves leaving texturally unaltered bedrock on the spurs separating the alcoves, yet texturally
383 altered bedrock upslope (Fig. 7G). In other places where pasted-on terrain extends into alcoves, the
384 bedrock on the alcove spurs and on the rim-crest are texturally altered as shown in Fig. 7H. In one
385 example we found pasted-on terrain coexisting with terrain hosting textures normally associated
386 with degrading latitude-dependent mantle, namely exposed layers and extensive pitting (e.g.,
387 Mustard et al., 2001; Schon et al., 2009a). As shown in Fig. 7I, the pasted-on terrain seems to be
388 emerging from beneath this mantle.

389 Our global incision survey shows that the thickness of the pasted-on terrain tends to increase with
390 increasing latitude in both hemispheres, from a mean of ~10 m at 28-30° to 40 m at around 60°, with

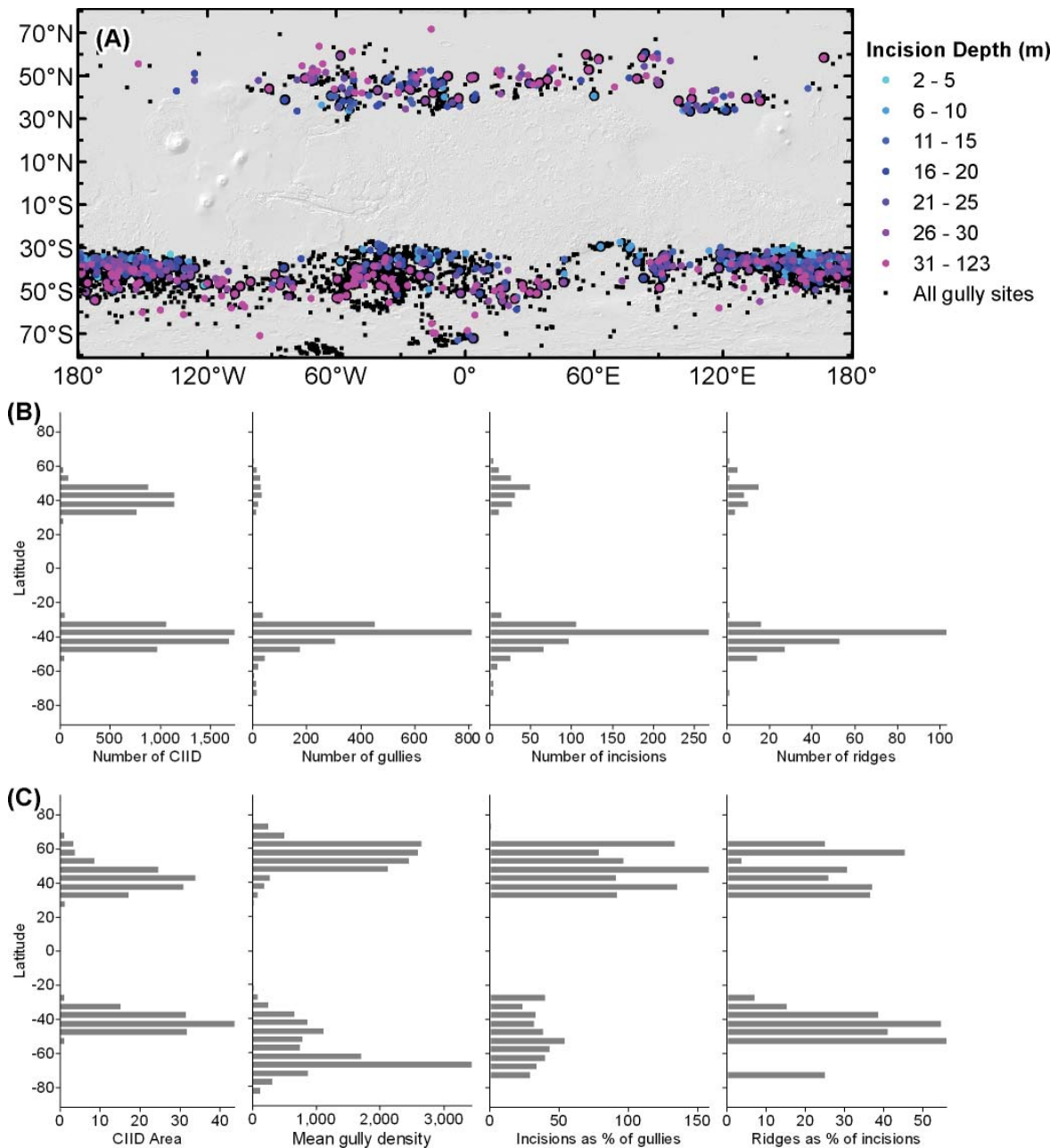
391 increasing scatter at higher latitudes (Fig. 8). The only difference between the hemispheres in Fig. 8
392 is that incisions are noted at lower latitudes in the southern hemisphere compared to the northern
393 hemisphere (28°S vs. 32°N). Incision depths are greater in the Argyre Basin than elsewhere (Fig. 9A).
394 Gullies with incisions follow the same spatial distribution as the gully population as a whole (Fig. 9)
395 and form a larger percentage of the overall gully population in the northern hemisphere compared
396 to the southern hemisphere (~100% compared to ~50% respectively; Fig. 9).

397 The pasted-on terrain, where present with other crater interior ice deposits, gradually transitions in
398 texture and topography into these other bodies, not presenting a single clean delimitation between
399 these units (Figs. 10G, 10L, 4C, 7B).



400

401 **Fig. 8.** Estimated gully incision depth against latitude, top: northern hemisphere and bottom:
 402 southern hemisphere. The solid lines represent mean values for each 5° of latitude. Dashed lines
 403 represent standard deviations for each 5° of latitude. The variable *n* is the number of data points
 404 within the graph.



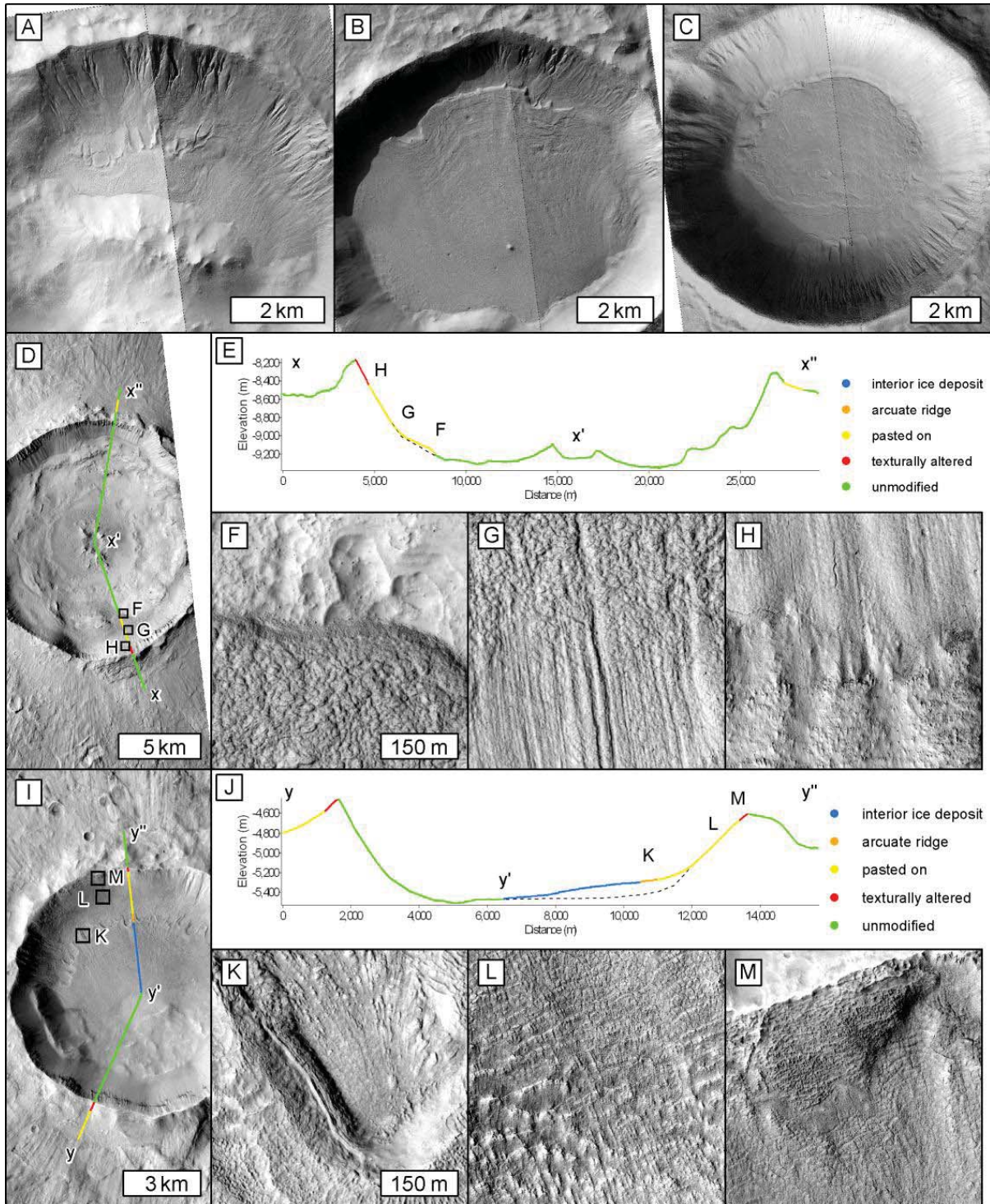
405
 406 **Fig. 9.** Global and latitudinal distribution of gullies, gully incisions, crater interior ice deposits, and
 407 arcuate ridges. (A) Map of gullies (Harrison et al., 2015) and depth of incisions measured in this
 408 survey overlain on a relief-shaded rendering of the Mars Orbiter Laser Altimeter elevation data. (B)

409 *Bar charts giving the counts, from left to right, of crater interior ice deposits (CIID), gullies, incisions,*
410 *and arcuate ridges per latitude band. Bar graphs were made from supplementary data of Levy et al.*
411 *(2014) and Harrison et al. (2015). (B) From left to right: summed crater interior ice-deposit (CIID) area*
412 *per 5° latitude bin in 10³ km², mean gully density per latitude from Conway et al. (2018a) that*
413 *normalises gully density based on the frequency of steep slopes found at that location, number of*
414 *incisions as a percentage of the number of gullies in the same latitude bin, and number of arcuate*
415 *ridges as a percent of the number of incisions per latitude bin. Note that the percent of incisions*
416 *exceeds 100% because of the nature of the two surveys: Harrison et al. (2015) counted sites (in which*
417 *many gullies could be present), and our incision survey counted individual gullies; hence there are*
418 *latitude bins where the number of incisions exceeds the number of sites.*

419 *2.3. Arcuate ridges and crater interior ice deposits*

420 Fig. 9C shows the latitudinal distribution of the 148 arcuate ridges coexisting with incised gullies.
421 Where ridges are associated with incised gullies, they tend to occur at higher latitudes — a trend
422 more obvious in the southern hemisphere. We find no difference between with thickness of pasted-
423 on terrain in the systems with arcuate ridges and those without, even when considered by latitude
424 (Fig. 9).

425 The arcuate ridges are often found in association with crater interior ice deposits (57% found within
426 5km of the boundary of crater interior ice deposits mapped by Levy et al., 2014), which can form
427 lobes of material that extend from the pole-facing crater wall beyond the ridges onto the crater floor
428 (Figs. 3A, 10A-C, I). In both cases the arcuate ridges are commonly outlined by spatulate depressions
429 that appear to push into the VFF. We found no evidence of arcuate ridges occurring without any kind
430 of existing crater interior ice deposit. We find that pasted-on terrain always occurs upslope of the
431 arcuate ridges and, by association, that arcuate ridges always coexist with texturally altered bedrock
432 upslope (Fig. 10).



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438

Fig. 10. Relationship between pasted-on terrain, arcuate ridges, and lobate or crater interior ice deposits. North is up in all panels. (A) Arcuate ridges within a lobe of crater interior ice deposits where the lobate margin appears to curve around obstacles. HiRISE images ESP_023809_1415 and ESP_024943_1415. (B) Arcuate ridges into crater interior ice deposits that extend across the whole crater floor with flowlines extending toward the south in the southern hemisphere. HiRISE images

439 *ESP_016227_1405 and ESP_016438_1405. (C) Arcuate ridges into crater interior ice deposits that*
440 *extend across the whole crater floor with no flowlines, yet pitted surface texture similar to other VFF,*
441 *in the northern hemisphere. HiRISE images ESP_013277_2155 and ESP_022059_2150. (D) Crater E*
442 *overview in CTX image P19_008307_2138 showing the position of the cross-profile in (E). (F-H) Panels*
443 *using HiRISE image PSP_009164_2140 at 1:10,000 showing (F) the boundary between the pasted-on*
444 *terrain and the crater floor, (G) the transition from the lineated pasted-on terrain texture to glacial*
445 *texture midway down the crater wall, and (H) the boundary between the lineated pasted-on terrain*
446 *and the texturally altered bedrock. (I) Taltal Crater overview in CTX image G09_021712_1402,*
447 *showing the location of the cross-profile in panel (J). (K-M) Panels using HiRISE image*
448 *ESP_21712_1400 at 1:10,000 showing (K) the boundary between the pasted-on terrain and the*
449 *arcuate ridges, (L) the transition between the polygonised pasted-on texture and VFF surface textures*
450 *midway down the crater wall, and (M) the boundary between the polygonised pasted-on terrain and*
451 *the texturally altered bedrock.*

452 Although we did not survey for arcuate ridges that do not coexist with gullies (this is ongoing),
453 Berman et al. (2005) systematically surveyed all the gullies and arcuate ridges in Newton Basin using
454 Mars Orbiter Camera (MOC) narrow angle images at 1.5 m/pixel and THEMIS (Thermal Emission
455 Imaging System) visible images. They found 188 craters with gullies, 118 craters with arcuate ridges,
456 and 104 craters with both, consistent with our finding that up to ~55% of incised gullies also have
457 associated arcuate ridges (Fig. 9). Berman et al. (2009) found that for craters with diameters >20 km
458 in Arabia terra, Newton Basin, and Eastern Hellas, arcuate ridges were almost exclusively associated
459 with gullies. The arcuate ridges had an even stronger preference for pole-facing crater walls than
460 gullies. Berman et al. (2005) and de Haas et al. (2018) noted a spatial correlation between gully
461 alcoves and the position of the most arcuate sections of the ridges: inflections between successive
462 arcuate segments have similar widths to the alcoves located topographically above them; our
463 observations agree. Moreover, the arcuate ridges below the largest alcoves also protrude farthest
464 onto the crater floor. The arcuate ridges mapped by Berman et al. (2005) systematically occurred

465 with *patterned floors* interpreted as ice-rich crater fill (here termed *crater interior ice deposits*) and
466 mantling materials. These crater interior ice deposits sloped away from the margins of the fill
467 proximal to the arcuate ridges, the slope direction aligning with the lineations expressed within the
468 surface of these fills, also noted by Head et al. (2008).

469 3. Erosion of crater walls

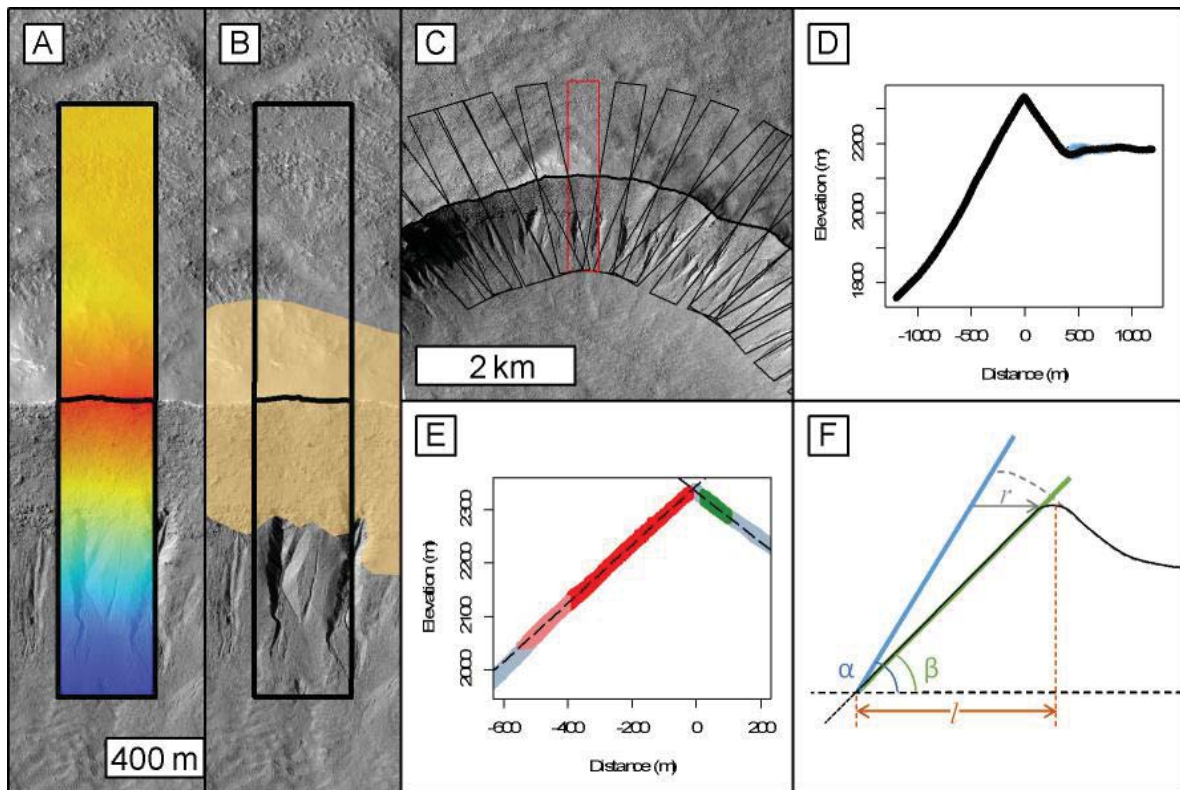
470 3.1. Approach

471 In order to assess the magnitude and rate of erosion associated with crater glaciation, we made a
472 series of topographic measurements and dated host craters via measurement of crater size-
473 frequency distributions.

474 The first criteria in our site selection was the existence of either a preexisting digital terrain model
475 (DTM) or a suitable stereo pair of images for creating a DTM. From these we selected three classes
476 of craters:

- 477 • Pristine impact craters in the databases of Watters et al. (2015) and Tornabene et al. (2018),
478 preferably located in the equatorial latitudes, but otherwise without evidence for any of the
479 following features: pasted-on terrain, other mantling deposits, arcuate ridges, or gullies.
- 480 • Craters with gullies, but no evidence of pasted-on terrain, other mantling deposits, or
481 arcuate ridges; preferably possessing HiRISE elevation data spanning north- and south-facing
482 walls.
- 483 • Craters with pasted-on terrain that can have arcuate ridges and/or gullies. Preferably having
484 HiRISE elevation data spanning north- and south-facing walls.

485 In all cases we favoured craters with a simple rim morphology, lacking wall terraces and wall collapse
486 features. This criteria and use of HiRISE images restricted our crater diameters to <20 km. We
487 avoided craters located on antecedent relief (faults, other crater rims, valleys).



488

489 **Fig. 11.** Overview of approach for measuring recession rates of crater walls. (A) Talu Crater with
 490 elevation swath overlain on HiRISE image ESP_011672_1395 (blue colours are high elevations, red
 491 colours low). (B) The swath profile from panel (A) with the mapped bedrock shown in orange. (C)
 492 Overview of the swath profiles measured on the northern rim of Talu Crater with the one in panel (A)
 493 highlighted in red. (D) Elevation data against distance from the rim-crest extracted from the swath
 494 shown in panel (A) with blue shades representing the raw data and black points showing the mean
 495 elevation every 1 m. (E) The same profile as in panel (B) but showing only the crater rim. Grey points
 496 are all the profile datapoints, superposed by red points that are used to perform a linear fit to obtain
 497 the bedrock slope on the inner wall (points 20-400 m from the rim with bedrock outcrop), pink points
 498 are bedrock outcrop lying farther than 400 m from the rim, and green points are those used to
 499 perform a linear fit to obtain the slope of the bedrock on the exterior wall (points 20-100 m from the
 500 rim with bedrock outcrop). Linear fits are shown as black dashed lines. (F) Diagram illustrating the
 501 method by which we calculated the headwall retreat. The black line represents the crater wall for
 502 which we want to estimate the retreat with the green line representing the linear fit to obtain the

503 *slope β . The dotted grey line is either a pristine crater wall or an unmodified wall in the same crater*
504 *with the blue line representing the linear fit to obtain its slope α . The orange distance l is the exposed*
505 *length of bedrock as observed from the orthorectified images. The recession r is calculated as $l - l$*
506 *($\tan \beta / \tan \alpha$). For interpretation of the references to colour in this figure legend, the reader is*
507 *referred to the web version of this article.*

508 Where existing HiRISE elevation data did not exist we used either the SocetSet ISIS3 workflow (Kirk
509 et al., 2008) or the freely available Ames Stereo Pipeline (ASP) (Moratto et al., 2010; Shean et al.,
510 2016) to produce additional elevation data. For both, we followed standard procedures in ISIS3 to
511 produce single unprojected seamless images from the 9-10 individual HiRISE CCD images. These
512 mosaicked images were map projected in ISIS3 to improve the point matching in ASP. We used
513 standard settings to run ASP, with the Bayes EM subpixel mode with a 15 pixel subpixel kernel to
514 improve results on steep terrain. We used ESA Mars Express High Resolution Stereo Camera
515 elevation data to coregister and correct any tilting in the resulting point cloud with the routine
516 'pc_align'. For the SocetSet workflow we followed the standard USGS procedure, but sometimes
517 used the *low contrast* strategy at the DTM creation step, which can produce less noisy results.

518 In order to estimate the headwall retreat of the crater bedrock we measured the slope of the
519 bedrock materials found just below the crest of the crater rim. We automatically extracted these
520 data by taking 300-m-wide swath profiles perpendicular to the rim trace. We used swath profiles in
521 order to generate data robust to noise in the elevation data and whose slopes would be
522 independent of the presence or absence of alcoves. We mapped where the bedrock cropped out
523 within each swath. For each swath profile we created an average profile, by binning the data by
524 distance every metre from the crater rim and taking the mean of the points within each bin. We
525 performed a linear fit of the elevation data within bedrock outcrops between 20 and 400 m from the
526 crater rim as an estimation of the upper bedrock slope of the inner crater wall and between 20 and

527 100 m on the exterior crater wall. This procedure is illustrated in Fig. 11. We did not measure any
528 cases where pasted-on terrain extended up to the crater rim because no bedrock is exposed.

529 We classified the swath profiles into four types depending on their association with identified
530 landforms within each crater: i) unmodified, ii) gullies present, but no evidence of glacial
531 modification, iii) gullies present with evidence of glacial modification; and iv) evidence of glacial
532 modification, but no gullies.

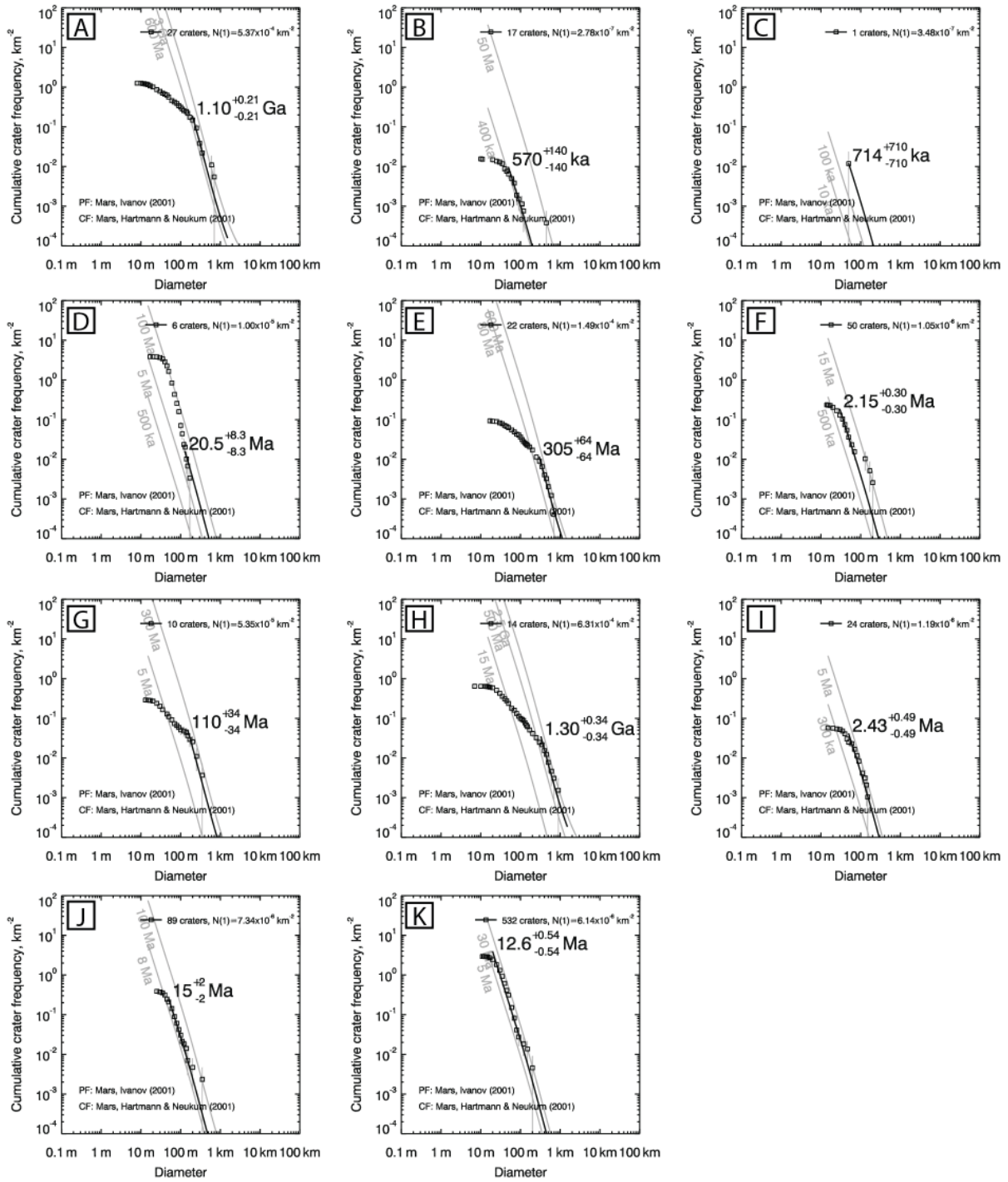
533 In order to convert the reduction in slope to a headwall retreat rate, we first need to estimate the
534 amount (length l , Fig. 11) of crater wall that has experienced this reduction in slope and then
535 combine this recession length with the crater age to obtain a minimum headwall retreat rate for our
536 studied craters. We measured the mean distance from the crater rim over which bedrock was
537 exposed to provide a minimum estimate of the horizontal length (l in Fig.11) of the crater wall that
538 had undergone slope reduction. We calculated two slope-reduction quantities:

539 (i) the slope reduction with respect to the opposing (unmodified) wall in the same crater (if
540 available) to give an estimate of the acceleration in erosion over the background rate; and

541 (ii) the slope reduction with respect to the maximum average wall slope value of pristine
542 equatorial craters to give the cumulative erosion over the lifetime of the crater.

543 To estimate the headwall retreat rate under the more likely scenario that periods of enhanced
544 erosion punctuated the background headwall retreat rate, we followed Levy et al. (2016) and
545 calculated the retreat rate over a time period of 0.5 Ma based on the slope-reduction given by (i).
546 Levy et al. (2016) used a time interval of 0.5 Ma based on previous estimates for the duration of the
547 last glacial epoch by Fassett et al. (2014) and the amount of time required, according to glacial flow
548 modelling, for a typical LDA to form (Fastook et al., 2014). For these relatively small systems this
549 duration is probably too long, but it provides a conservative estimate of erosion rate.

550 We used Crater Tools and CraterStats extensions for ArcGIS (Michael and Neukum, 2010; Kneissl et
551 al., 2011) to estimate the ages of those ejecta blankets of our study craters, which did not already
552 have published ages. In brief: we outlined the continuous impact ejecta on CTX images, which was
553 then used as our crater count area. We counted all impact craters that had raised rims and were not
554 covered/infilled by the ejecta. We fitted the Hartmann and Neukum (2001) isochrons to the resulting
555 cumulative size-frequency distributions to obtain an estimated age and error (Fig. 12). These ages
556 are taken as maximum values for the age of the features found within the craters because the
557 features formed after the impact. We were conservative in determining the potential uncertainties
558 on the crater counting: we manually placed isochrons that bracketed our fitted ages and touched
559 the error bars and/or points at the oldest and youngest end of the data. In one case no identifiable
560 primary craters were visible on the ejecta surface at CTX resolution so we calculated a maximum age
561 by placing one *false* impact crater in the count area of 55 m diameter — the smallest diameter that
562 we judged to be identifiable in the CTX image.



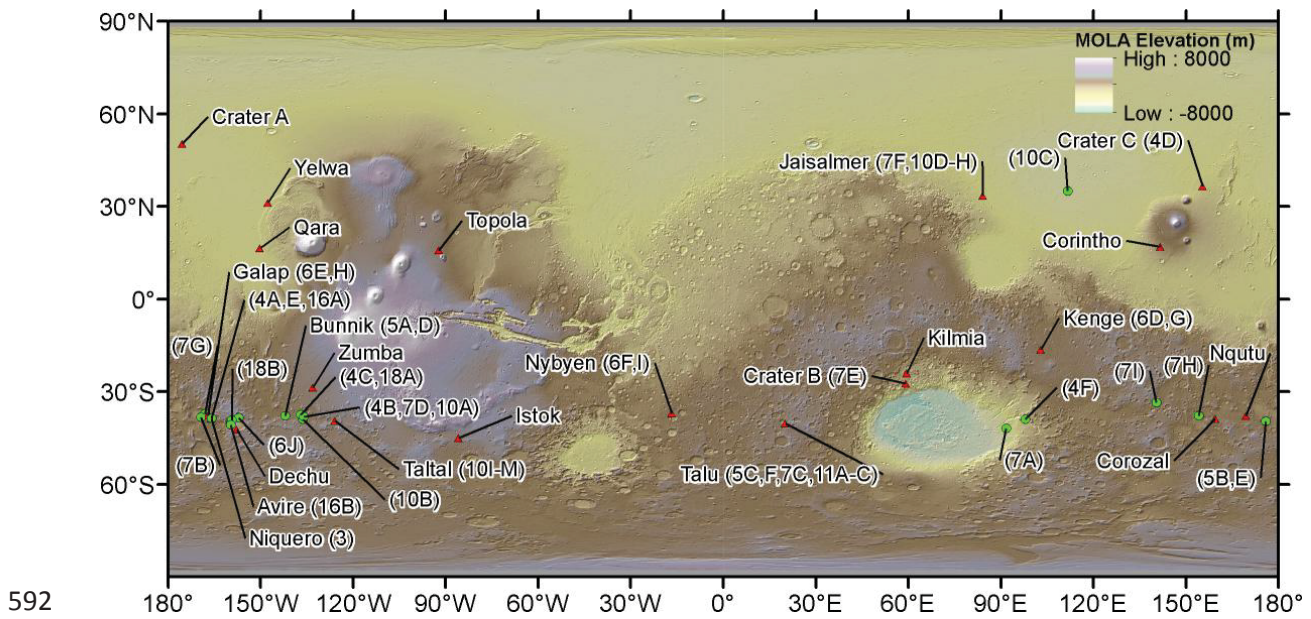
563

564 **Fig. 12.** Crater size-frequency distribution for the eight previously undated craters in our study (Table
 565 1). Binning is performed via the pseudo-log method of Neukum (1983) and Hartman and Neukum
 566 (2001). (A) Kilimia Crater where CTX image G22_026861_1557 was used to count the craters. Fitting
 567 was performed between diameters of 200 m and 1.5 km. (B) Jaisalmer Crater where CTX images
 568 P21_009164_2137, P15_006949_2150, P19_008307_2138, and P21_009098_2138 were used to

569 count the craters. Fitting was performed between diameters of 50 m and 1 km. (C) Unnamed crater
570 'Crater A' where CTX image G19_025498_2305 was used to count the craters. For this crater no
571 primary craters could be found, so one crater measuring 55 m in diameter was placed in the count
572 area to obtain a maximum age. (D) Yelwa Crater where CTX image G06_020539_2114 was used to
573 count the craters. Fitting was performed between diameters of 130 m and 1 km. (E) Dechu Crater
574 where CTX images P12_005798_1396, P07_003675_1391, G14_023612_1375, and
575 D10_031102_1378 were used to count the craters. Fitting was performed between diameters of
576 300 m and 1.5 km. (F) Unnamed crater 'Crater B' where CTX image G06_020703_1524 was used to
577 count the craters. Fitting was performed between diameters of 30 m and 1 km. (G) Nybyen Crater
578 where CTX image B05_011436_1427 was used to count the craters. Fitting was performed between
579 diameters of 150 m and 1.5 km. (H) Niquero Crater where CTX images F01_036258_1410,
580 P03_002383_1417, and P11_005284_1419 were used to count the craters. Fitting was performed
581 between diameters of 330 m and 1.5 km. (I) Unnamed crater 'Crater C' where CTX images
582 B02_010559_2168 and G22_026910_2168 were used to count the craters. Fitting was performed
583 between diameters of 50 m and 1 km. (J) Topola crater where CTX image G14_023662_1960 was
584 used to count the craters. Fitting was performed between diameters of 50 m and 1 km. (K) Kenge
585 Crater where CTX image F23_044780_1635 was used to count the craters. Fitting was performed
586 between diameters of 20 m and 1 km.

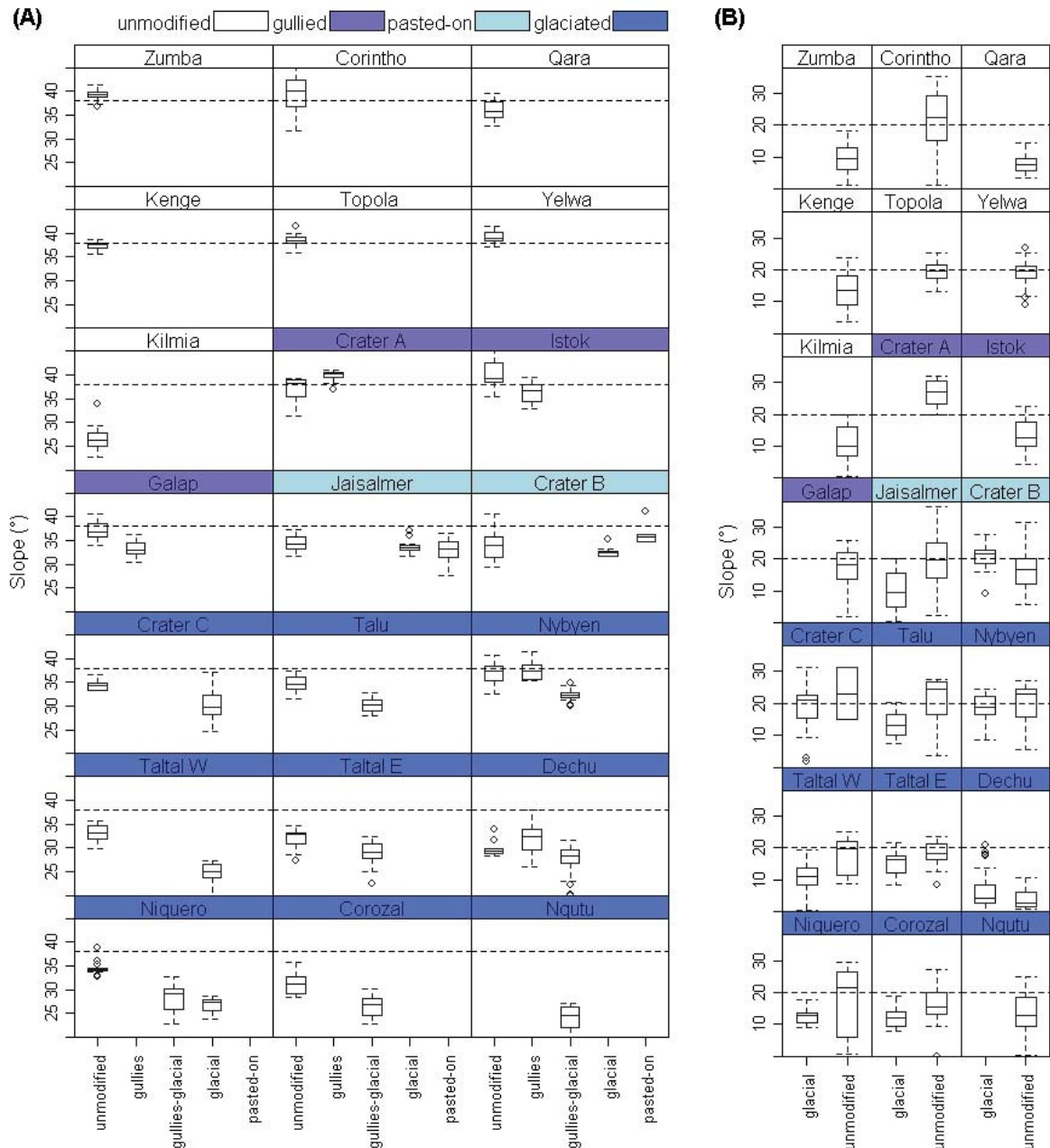
587 *3.2. Headwall retreat rate of glaciers and gullies*

588 We used a data set comprising three gullied craters, nine craters with arcuate ridges, two craters
589 with pasted-on terrain yet no arcuate ridges, six fresh craters, and one degraded-looking equatorial
590 crater with no gullies, pasted-on terrain, or arcuate ridges (Table 1, Fig. 13). These craters have
591 diameters ranging between 1.7 and 20 km.



592 **Fig. 13.** Location map of the DTMs used in this study (triangles) and locations of the additional sites
 593 used in the figures (circles) on a MOLA elevation shaded relief base map. The numbers/letters in
 594 brackets identify the relevant figure(s).
 595

596 We found that the upper wall slopes are $\sim 39^\circ$ for the two youngest equatorial craters (≤ 5 Ma,
 597 Zumba and Corinto), and in our oldest equatorial example (Kilmia, 1.1 Ga) the upper wall slope has
 598 reduced to 27° (Table 2 and Figs. 14, 15). In the terrestrial literature we found that a peak in slope
 599 exists at a similar value of $\sim 39^\circ$ in mountain ranges, which is thought to represent the strength of the
 600 bedrock (e.g., Burbank et al., 1996; Katsube and Oguchi, 1999; Korup, 2008; Lin et al., 2009).
 601 Similarly for the youngest craters in our modified sample (0.5-6.5 Ma craters A, B, C, Istok, Jaisalmer,
 602 Galap) the slopes of the walls unmodified by either gullies or glacial processes have high slope values
 603 with a range of $34-41^\circ$ (Figs. 14, 15). The unmodified walls of older craters (1.3-1.4 Ga, Niquero,
 604 Corozal,) have lower slopes in the range of $32-34^\circ$.



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Fig. 14. Box plots displaying results from our crater wall slope-analysis. Black boxes indicate the interquartile range, whiskers the maximum and minimum values, horizontal black line the median value, and black dots outliers (classed as those values farther than 1.5 interquartile-ranges from the median). (A) Box plots of slope data gathered for the inner crater wall of our studied craters. Dotted line is arbitrarily located at 38° for reference. (B) Box plots of slope data gathered for the exterior crater wall of our studied craters. Dotted line is arbitrarily located at 20° for reference. Note: *pasted-on* refers to parts of the crater wall with pasted-on terrain yet little textural alteration of the bedrock,

613 *whereas glacial refers to areas with pasted-on terrain and textural alteration of the bedrock (Fig.*
614 *7F). in the items in the legend refer to different classes of crater: unmodified = unmodified equatorial*
615 *craters, gullied = gullied and unglaciated craters, pasted-on = craters with pasted-on terrain but no*
616 *arcuate ridges or interior ice deposit, and glaciated = glaciated craters.*

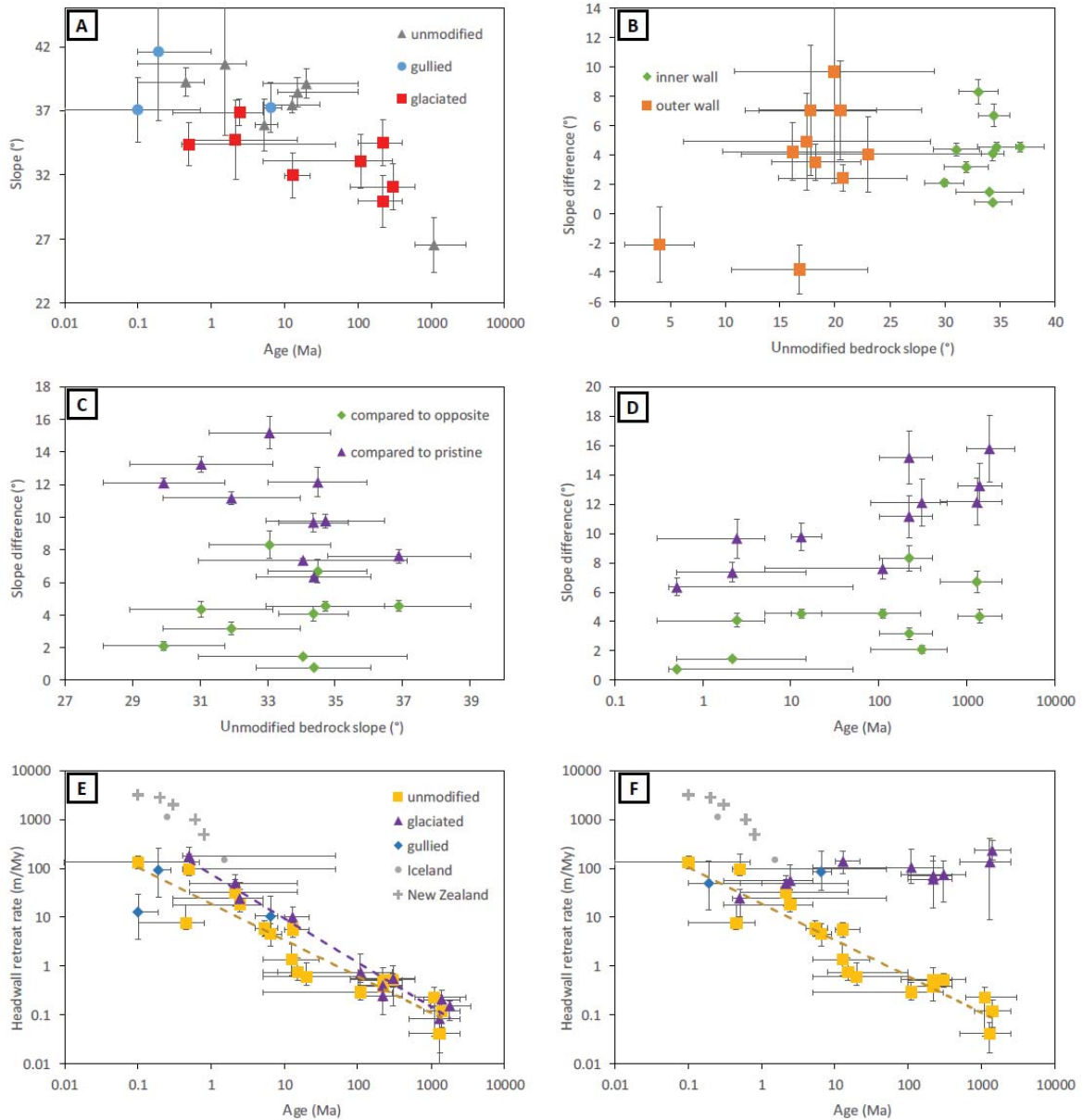
617 The reduction in interior crater wall bedrock slope in the three gullied yet unglaciated craters is 4-5°
618 compared to their unmodified crater walls and 6-9° compared to slopes in fresh equatorial craters
619 (Figs. 14, 15; Table 2). The interior walls of craters that have evidence of glaciation are further
620 reduced: up to 8° (on average 4°) lower compared to conjugate unmodified walls and up to 15° (on
621 average 11°) compared to fresh equatorial craters. We also found that when pasted-on terrain was
622 present on exterior crater walls, these bedrock slopes were reduced compared to unmodified
623 exterior crater walls (Figs. 14, 15; Table 2), although the signal is not as strong as for the inner walls.
624 This is probably because of a larger initial variation in the slope of exterior crater walls, as shown by
625 their variability in fresh equatorial craters (a range of 8° to 21° with standard deviations of 3-10°,
626 Table 2). Finally, for Jaisalmer Crater and Crater B (the youngest glaciated craters) we noted that wall
627 sectors with pasted-on terrain and texturally altered bedrock had lower slopes than rim portions
628 with only pasted-on terrain and both were lower than the unmodified portions of the crater wall
629 (Fig. 14C).

630 For the gullied craters the bedrock extends on average 120-520 m down the crater wall (Table 2);
631 therefore, the crater wall has receded by 1-70 m at the top. For Galap this results in an estimated
632 headwall retreat rate of 92 m My⁻¹ and for Istok 10 m My⁻¹, compared to 4.5 and 75 m My⁻¹
633 backweathering rates estimated by de Haas et al. (2015b) using an independent method.

634 The slopes within glaciated craters are partially covered by pasted-on terrain, hence as a
635 conservative estimate we can assume that only the bedrock that is exposed has been affected by the
636 slope reduction. The pasted-on terrain is found between 121 and 714 m and on average 368 m
637 distance from the rim in our studied craters (Table 2). We find that headwall retreat rates range

638 between 0.04 and 132 m My⁻¹ for unmodified walls of glaciated and gullied craters and overlap with
639 the values of 0.22 to 7.6 m My⁻¹ for completely unmodified craters. We further find that the
640 headwall retreat rates of glaciated walls fall between 0.08 and 181 m My⁻¹. When these headwall
641 retreat rates are plotted against crater age (Fig. 15E), a decreasing trend in headwall retreat rate
642 against age is observed. Similarly for the glaciated crater walls, the calculated headwall retreat rate
643 has a similar decrease with crater age but transposed to higher headwall retreat rates than the
644 unmodified walls. Calculating the headwall retreat rate for the glaciated crater walls over a fixed
645 time interval of 0.5 Ma (as detailed in section 3.1) and using the slope reduction compared to the
646 unmodified wall of the same crater (rather than compared to a fresh equatorial crater wall), we find
647 that headwall retreat rate no longer has a substantial trend with age (Fig. 15F).

648 We briefly note here that our data do not show that glaciated inner crater walls have significantly
649 and/or systematically lower elevation than unmodified inner walls in the same crater as observed
650 previously by Berman et al. (2005) and Head et al. (2008). We posit that this could be because of the
651 fact that we find erosion on the inner and outer crater walls, hence north- and south-facing walls are
652 being lowered by the action of glaciation.



653

654 **Fig. 15.** Wall slopes, reduction in wall slope, and headwall retreat rate for our studied craters. (A) The
 655 slope of exposed bedrock in the unmodified walls of craters in this study, including craters with
 656 neither glacial nor gully features (unmodified), craters with gullies but no glacial features (gullied),
 657 and craters with glacial features (glaciated). (B) The slope difference between glaciated walls and
 658 unmodified walls from the same crater versus the unmodified wall slopes for outer and inner crater
 659 walls. (C-D) The slope difference against unmodified wall slope and age respectively. The legend for
 660 both plots is in panel (C). (E) Estimated headwall retreat rate against age for our craters separated
 661 into unmodified crater walls and those affected by glacial erosion and gully erosion. Plotted for

662 *comparison are erosion rates from modelled glacial landscapes in New Zealand (Egholm et al., 2012)*
663 *and measured glacial erosion from Iceland (Geirsdóttir et al., 2007). Trendlines are added only to*
664 *guide the eye. (F) The same plot as in panel (5E), but the headwall retreat rate is calculated over a*
665 *fixed 500 ky interval and uses the slope difference compared to the unmodified wall in the same*
666 *crater (rather than comparison to an equatorial crater wall slope). NB: The error bars for slope*
667 *represent the standard deviation of the slope measurements and age error bars the potential age*
668 *range for the host craters (Table 1). The uncertainty in slope difference is calculated by propagation*
669 *of errors and the uncertainty in retreat rate is dominated by the length of exposed bedrock, hence*
670 *uncertainties are calculated by recalculating the retreat rate using the minimum and maximum*
671 *length of the exposed bedrock for that crater.*

672 4. Glacial erosion

673 4.1. Mechanisms of glacier erosion on Earth

674 Mechanical erosion by glaciers is dominated by abrasion, quarrying, plucking, and glaciotectionism
675 (see Benn and Evans, 2010, for an extensive review). Abrasion is the process by which the glacier bed
676 and entrained clasts are scoured either by debris entrained within basal glacier ice or, less
677 commonly, by the basal ice itself (e.g., Hallet, 1979). Quarrying is the process by which bed-clast
678 contact, or overriding of bedrock cavities, generates foci of pressure at the ice-bed interface and
679 liberates fragments from the bed (e.g., Iverson, 1991; Hallet, 1996; Cohen et al., 2006). Plucking
680 involves freezing of meltwater within, or deformation of basal ice into, bedrock fractures and
681 prizing-off of fragments under subsequent glacier motion (e.g., Boulton, 1979; Röthlisberger and
682 Iken, 1981). Finally, glaciotectionism is the process by which subglacial, submarginal, and/or
683 proglacial materials deform under stresses induced by glacial ice (e.g., Hart and Boulton, 1991).
684 Glacier thermal regime exerts a fundamental control upon the efficacy of glacial erosion on Earth.
685 This arises from its influence on the generation of meltwater, entrainment of erosional *tools* (i.e.
686 debris), and dynamics of ice-bed interactions (e.g., Hallet et al., 1996; Hambrey and Glasser, 2012).

687 The thermal regimes of glaciers are categorised according to their temperature relative to the
688 pressure melting point of ice and are controlled by complex interactions between climatic,
689 environmental, and glaciological parameters (Benn and Evans, 2010). Temperate (warm-based)
690 glaciers (e.g., Haut Glacier d’Arolla, Swiss Alps; Goodsell et al., 2005) are at the pressure melting
691 point of ice throughout, whereas cold-based glaciers (e.g., Meserve Glacier, Antarctic Dry Valleys;
692 Holdsworth and Bull, 1970) are entirely below the pressure melting point. Polythermal glaciers (e.g.,
693 Midre Lovénbreen, Svalbard; Björnsson et al., 1996), where *warm* ice at the pressure melting point
694 coexists with *cold* ice below the pressure melting point, represent an intermediate condition
695 between temperate and cold-based glaciers.

696 The efficiency of mechanical erosion by glaciers is greatly enhanced in the presence of meltwater.
697 Under a wet-based thermal regime, significant basal sliding can occur at an unfrozen ice-bed
698 interface, promoting efficient abrasion of the bed (e.g., Hallet, 1979) and liberation of rock
699 fragments via plucking. Sliding over obstacles at the bed also opens lee-side cavities, promoting
700 quarrying (e.g., Iverson, 1991; Hallet, 1996). Quarrying is particularly effective under wet-based
701 regimes because meltwater, which is dynamic on shorter timescales than glacial ice, can greatly
702 enhance pressure fluctuations at the bed (Iverson, 1991; Cohen et al., 2006). Liquid water is thought
703 to play an important, possibly essential, role in promoting glaciotectionic deformation. Pore water
704 reduces the yield stress of glacial sediments, making them more susceptible to deformation under
705 stresses induced by glacial ice (e.g., Bennett, 2001). Despite generally lower meltwater volumes
706 associated with glaciers in cold-climate or permafrost regions, ground ice and/or frozen glacier
707 margins may encourage glaciotectionic processes by preventing efficient drainage of meltwater from
708 aquifers that they confine (e.g., Moran et al., 1979; Fitzsimons, 1996; Boulton et al., 1999; Benn and
709 Evans, 2010).

710 The frozen ice-bed interface and absence of meltwater in cold-based glacial systems, it is leads to
711 the assumption that cold-based glaciers flow entirely by internal deformation, do not erode their

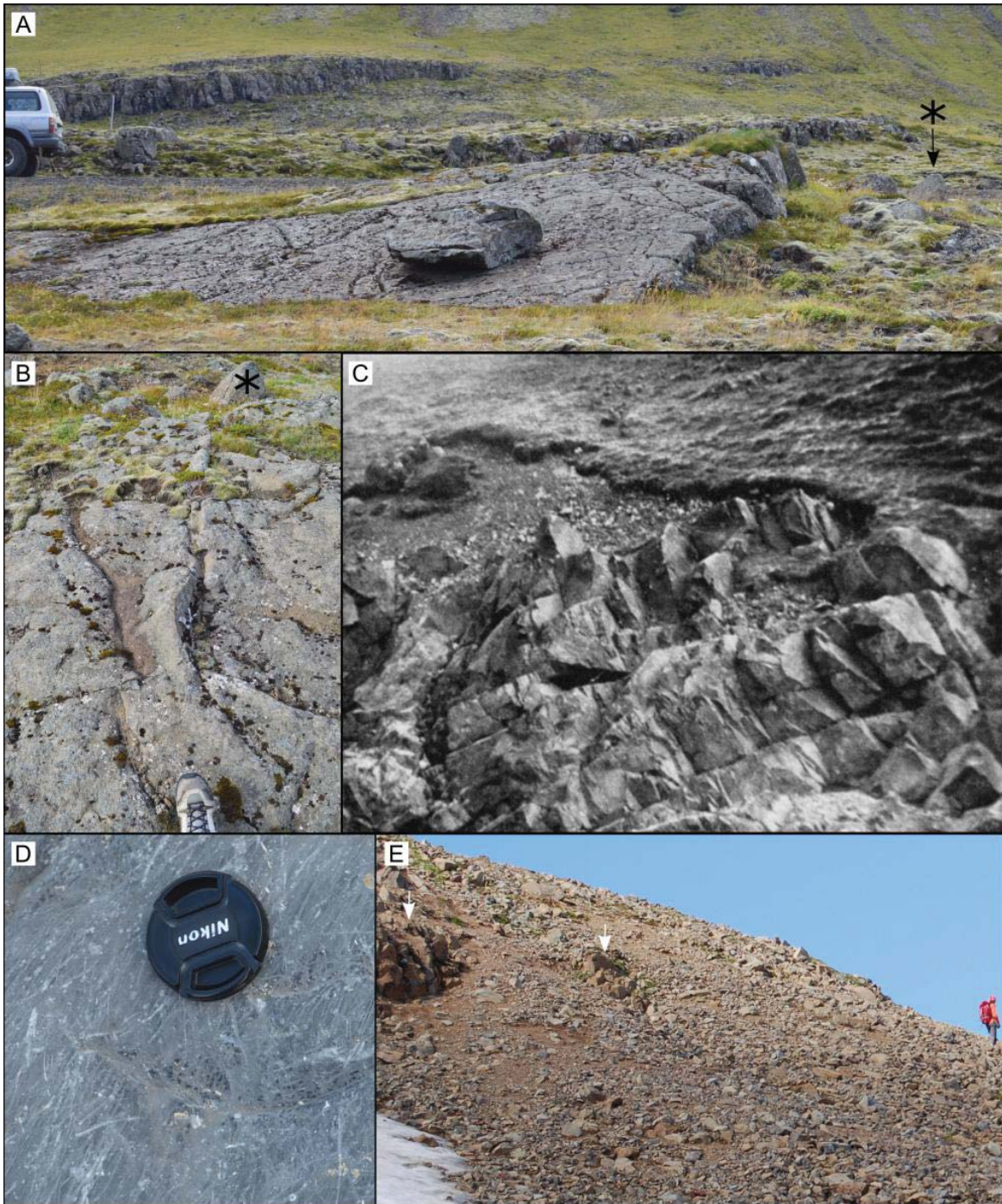
712 beds, and exert little or no detectable geomorphic influence upon the underlying landscape. Indeed,
713 coverage by cold-based ice is frequently invoked as a protective mechanism to explain the
714 preservation of features generated by previous wet-based glaciations (e.g., Dyke, 1993; Smith et al.,
715 2009). However, theoretical (Shreve, 1984) and field-based (Echelmeyer and Wang, 1987) studies
716 have noted that sliding rates of cold-based glaciers are, in fact, non-zero when integrated over long
717 timescales (Cuffey et al., 2000; Waller, 2001). Field observations have also identified evidence for
718 reworking of cold glacier beds (Cuffey et al., 2000; Atkins et al., 2002; Lloyd Davies et al., 2009), such
719 that it is becoming increasingly evident that cold-based glaciers can exert significant geomorphic
720 influence over timescales of glacial advance, although still substantially less than warm-based
721 glaciers.

722 In examining isotopic composition of dirty basal ice layers derived from marginal apron overriding at
723 Meserve glacier in the Antarctic Dry Valleys, Cuffey et al. (2000) suggested that interstitial water
724 films between ice and immersed solids within dirty basal ice layers of cold-based glaciers may permit
725 sliding and abrasion down to temperatures of -30°C , an effect that could be particularly important in
726 the presence of a highly saline substrate. Additionally, Lloyd Davies et al. (2009) proposed a
727 mechanical model by which abrasion and quarrying can operate beneath cold-based glaciers in the
728 absence of liquid water, based on field observations in the Allan Hills region of Antarctica. Ice-
729 marginal aprons comprising collapsed ice blocks commonly accumulate at the foot of steep terminal
730 ice cliffs of cold-based glaciers. These aprons can be incorporated into basal ice as they are
731 overridden during glacier advance (Cuffey et al., 2000; Atkins et al., 2002; Lloyd Davies et al., 2009).
732 Lloyd Davies et al. (2009) suggested that this incorporates a weak, low-density layer into the
733 submarginal basal ice that focusses stress onto up-glacier bedrock contacts and promotes fracturing
734 and quarrying of bedrock protuberances. Entrained rocks are then available as *tools* for abrasion
735 during subsequent glacier motion (Lloyd Davies et al., 2009).

736 4.2. Landscapes of glacial erosion and application to Mars

737 We now consider the potential contributions of these erosional processes to the following
738 geomorphic features that we observe on Mars: texturally altered bedrock of upper walls of impact
739 craters, lowering of slopes on texturally altered crater walls compared to fresh crater walls,
740 instances of sharp crater rim crests despite lowered wall slopes, and associated slope-side pasted-on
741 terrain and arcuate ridges that are commonly but not ubiquitously associated with spatulate
742 depressions within crater interior ice deposits.

743 Many geomorphic features that are indicative of past glacial erosion, including striae, gouges,
744 fractures, scrapes, and chattermarks (Atkins et al., 2002; Lloyd Davies et al., 2009; Benn and Evans,
745 2010), are undetectable at the decametre- to metre-scale resolutions of existing orbital remote
746 sensing data sets for either Earth or Mars. However, the integrated effect of these processes
747 operating over large areas typically manifests as smoothed, rounded, planed-off, or '*really-scoured*
748 surfaces (Benn and Evans, 2010). Such landscapes bear similarities to the smoothed morphologies of
749 the walls of the glaciated craters in our survey, within which the upper bedrock protuberances (*spur*
750 *and gully* forms) have been planed-off and possibly old gully alcoves strongly modified (de Haas et
751 al., 2018). If the textural disruption of these slopes is a primary morphology inherited directly from
752 glacial erosion, it seems more consistent with erosion via quarrying and plucking, which can
753 generate rough, *craggy* surfaces (e.g., Roberts and Long, 2005), than by abrasion, which tends to
754 smooth and polish bedrock surfaces (Fig. 16). Highly brecciated and fractured bedrock associated
755 with impact crater walls (de Haas et al., 2015a) may make these bedrock surfaces particularly
756 susceptible to erosion by quarrying and plucking, as in tectonically active glaciated regions on Earth
757 (e.g., Hallet et al., 1996). Further, the amount of slope lowering we observe is independent of the
758 estimated initial slope. This is consistent with glacial erosion, which is more strongly influenced by
759 the ice-surface slope (which controls basal shear stress) than the bed-slope (Oerlemans, 1984).



760

761 **Fig. 16.** Examples of differences in macroscale morphology between surfaces influenced by glacier
 762 abrasion or plucking on Earth and a comparison to slopes affected by periglacial erosion. (A) A small
 763 *rôche moutonnée* (image credit Frances Butcher) at the eastern end of Hvalfjörður (west Iceland),
 764 looking toward the southern side of Botnsdalur showing the difference in macroscale surface texture
 765 generated by abrasion on the stoss slope and plucking on the lee slope. Ice flow was from left to
 766 right. The star indicates the same boulder that is visible in panel (B), looking along the same landform

767 *close to its crest, showing the transition from smoothed to craggy surface form (image credit Frances*
768 *Butcher). (C) A craggy surface in Glenridding Beck (The Lake District, UK) generated by plucking by*
769 *glacial ice. Reproduced from Hay (1934). (D) Small-scale forms generated by abrasion and quarrying*
770 *on the glacially smoothed surface of a transported boulder in Long Hill esker pit, County Westmeath,*
771 *Ireland. Striae formed by abrasion are oriented in the former direction of ice flow (either bottom left*
772 *to top right, or vice versa). A crescentic or lunate gouge (dependent on ice flow direction) formed by*
773 *glacial quarrying of the boulder surface crosses the centre of the image (image credit Frances*
774 *Butcher). (E) Periglacially modified slopes on the eastern flank of Lambahnjúkur in northern Iceland*
775 *with two bedrock outcrops (white arrows), illustrating the similarity between macro-scale*
776 *morphology of surfaces affected by plucking (e.g., panel C) and periglacial processes (image credit*
777 *Costanza Morino).*

778 Another possible contributor to the textural alteration of the bedrock is enhanced weathering under
779 the ice or snowpack at the top of the slope (Fig. 16E). In a terrestrial setting, rock weathering in cold
780 regions is controlled by moisture availability (e.g., Sass, 2005) and by the existence of the thermal
781 conditions required for ice segregation. These conditions are enhanced by the presence of
782 snowpatches or snowfields (e.g., Thorn and Hall, 1980), as ice lens growth during ice segregation is
783 favoured by slow freezing rates and sustained below-freezing temperatures (Matsuoka and Murton,
784 2008). Ice lens formation is the primary cause of physical rock breakdown or frost-shattering
785 (Murton et al., 2006; Hales and Roering, 2007; Matsuoka and Murton, 2008). These physical
786 weathering mechanisms require the presence of thin films of liquid water, yet ice segregation on
787 Mars has been hypothesised to occur without need for the liquid phase (e.g., Lacelle et al., 2013;
788 Sizemore et al., 2015).

789 The observed lowering of crater wall slopes presents a more complex challenge for the glacial
790 hypothesis. Glaciers tend to steepen upper valley slopes, in longitudinal (e.g., MacGregor et al.,
791 2000) and cross-sectional (Hirano and Aniya, 1988; Harbor, 1992) profiles, giving rise to classic U-

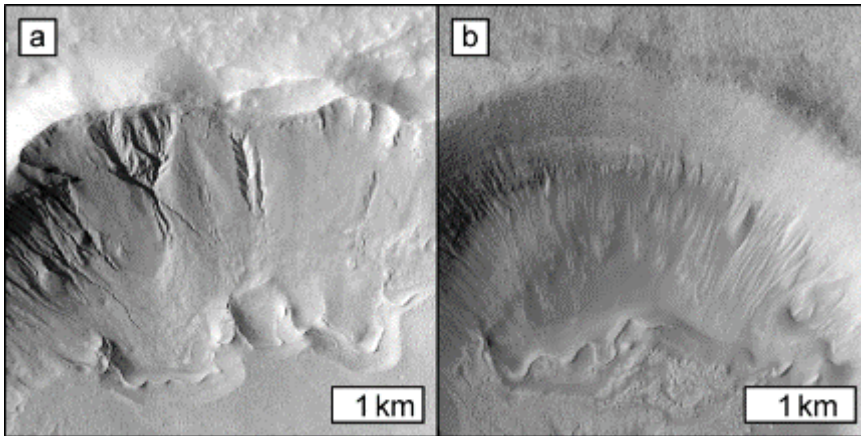
792 shaped valley and cirque landscapes. Conversely, our observations show shallower crater wall
793 slopes, and an absence of pronounced U-shaped undulations or bedrock-confined overdeepenings.
794 However, Hirano and Ania (1988) and Harbor (1992) found topographic steepening to be an
795 outcome of enhancement of vertical erosion by topographic confinement of glaciers and its effect on
796 ice flow. Glaciers that are not topographically confined can preferentially widen rather than deepen
797 their host valleys, resulting in broad-scale lowering of slopes (Hirano and Aniya, 1988). We therefore
798 suggest that valley and cirque-style glacial configurations are inconsistent with our observations and
799 that unconfined, piedmont-style (LDA-like) glaciers may provide the optimum explanation for the
800 observed morphologies. The location of pasted-on terrain in wide alcoves (which could be
801 interpreted to resemble cirques) does not necessarily imply that the ice was limited to those alcoves.
802 Weak vertical erosion could explain the presence of distinct rim and interalcove crests in some
803 instances, despite clear evidence that these same rim segments have undergone significant lowering
804 of slope.

805 However, the present understanding of glacial slope modification on Earth is largely based on
806 assumptions that glaciers inherit preexisting fluvial valleys with gentle slopes and have a ready
807 supply of meltwater to their beds (e.g., Hirano and Aniya, 1988; Harbor, 1992; MacGregor et al.,
808 2000). These assumptions conflict significantly with the present understanding of recent glacier
809 thermal regimes on Mars (largely cold-based) and the impact crater wall landscapes explored here,
810 which have steep initial slopes and lack mature fluvial valleys. Therefore, we emphasise that
811 constraining the configuration of glacial ice at small spatial scales would be highly speculative given
812 existing data and the distinctly atypical initial topography compared to terrestrial case studies upon
813 which glacial theory is based.

814 The magnitude of lowering of crater wall slopes, demonstrates that significant volumes of material
815 have been mobilized by crater wall erosion. Within the surveyed craters, pasted-on terrain is
816 invariably located downslope of texturally-altered bedrock. Accordingly, we propose that upper

817 crater walls and rims may have provided a viable source of debris for this pasted-on terrain and that
818 glaciers could have acted as important erosional and depositional agents in its formation. In this
819 context, the origin for the downslope lineations within pasted-on terrain could be streamlined glacial
820 bedforms such as flutes or megascale glacial lineations. However, the pasted-on terrain is
821 significantly more extensive and voluminous (Christensen, 2003; Conway and Balme, 2014; section
822 2.2) than rare, patchy, and thin (<2 m thick) till deposits identified in association with cold-based
823 glaciers on Earth (Atkins et al., 2002; Lloyd Davies et al., 2009). If the pasted-on terrain on formerly
824 glaciated crater walls is indeed glacial till, some degree of meltwater activity seems necessary,
825 contrary to the idea that Amazonian glaciation was almost entirely cold-based (with a few localised
826 and transient exceptions, e.g., Fassett et al., 2010; Hobley et al., 2014; Scanlon et al., 2014, 2015;
827 Gallagher and Balme, 2015; Butcher et al., 2017). We therefore emphasise that the origin of the
828 pasted-on terrain should be investigated further.

829 Given that aeolian processes are widespread and active geomorphic agents on Mars in the present
830 day and have been shown to modify gullied slopes (de Haas et al., 2013, 2015c), an aeolian origin for
831 the downslope lineations within pasted-on terrain might be just as likely as an origin as streamlined
832 glacial bedforms such as flutes or megascale glacial lineations. Therefore, convergence of form
833 between morphologies generated by glacial and aeolian streamlining prevents conclusive distinction
834 between these formation mechanisms given the resolution of HiRISE images. Thus, their potential
835 origin as streamlined glacial bedforms should not be excluded considering the apparent close
836 association between pasted-on terrain and glaciated slopes.



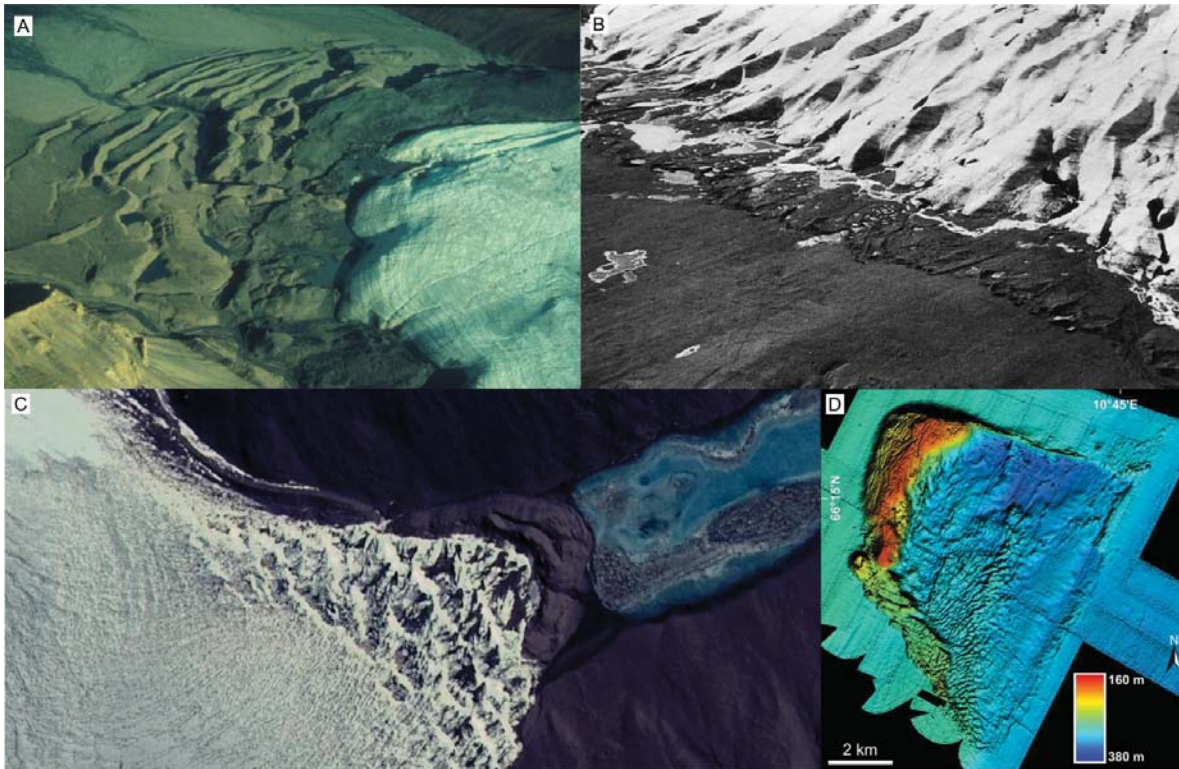
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838 **Fig. 17.** Examples of sinuous arcuate ridges with and without associated upslope alcoves. (A) HiRISE
 839 image ESP_038236_1410 where loops in the arcuate ridges appear to be associated with upslope
 840 alcoves of a similar width. (B) Avire Crater in HiRISE image ESP_029467_1390 with a similarly sinuous
 841 arcuate ridge to panel (A) but without any obvious upslope alcoves.

842 Prominent arcuate ridges provide particularly convincing evidence for past glaciation of crater walls.
 843 Although hypotheses that such arcuate ridges represent proglacial ramparts have been proposed
 844 (Whalley and Azizi, 2003), they are widely interpreted as end moraines based on their similarity to
 845 moraine ridges on Earth and to arcuate ridges bounding other ice-rich glacier-like forms on Mars
 846 (e.g., Milliken et al., 2003; Arfstrom and Hartmann, 2005; Hubbard et al., 2014; Brough et al., 2016a).
 847 Typically, multiple arcuate ridges coexist, forming laterally extensive moraine complexes. Although
 848 some arcuate ridge complexes can clearly be associated with bedrock alcoves in the crater walls
 849 above (Fig. 17A) (Arfstrom and Hartmann, 2005), many arcuate ridges lack distinct association to
 850 discrete undulations in the crater walls (Fig. 17B). Thus, we consider the ridges to be more consistent
 851 with unconfined piedmont-style glaciation, as proposed above, than with the formation of each
 852 arcuate section at the terminus of a discrete glacier tongue. The highly curvilinear morphologies of
 853 the ridge complexes are not inconsistent with such an origin, as differential flow within broad ice
 854 lobes on Earth commonly forms highly curvilinear moraine complexes (Fig. 18B) (Bennett, 2001;
 855 Arfstrom and Hartmann, 2005).

856 The common association between arcuate ridges and spatulate depressions within crater interior ice
857 deposits (Head et al., 2008) indicates that ice-contact bulldozing contributes to the formation (e.g.,
858 Arfstrom and Hartmann, 2005). However, several features lead us to support that glaciotectionic
859 processes also provided a significant contribution to their formation, as was considered by Arfstrom
860 and Hartmann (2005). The distinction between bulldozing and glaciotectionic processes is in the
861 nature of ice contact; while bulldozing relates to movement of material in contact with glacial ice,
862 glaciotectionism can influence materials tens to hundreds of metres below glacier beds and into the
863 proglacial zone (Benn and Evans, 2010). Compressive stresses induced by convergence of crater wall
864 glaciers with preexisting crater floor ice bodies, such as crater interior ice deposits (e.g., Fig. 10),
865 would enhance the likelihood that glaciotectionism supplemented bulldozing in moraine construction
866 where these preexisting ice bodies are present. Furthermore, bulldozing of crater floor materials
867 alone cannot necessarily provide adequate explanation for apparent disruption within crater interior
868 ice deposit surfaces beyond the arcuate ridges (Fig. 17). Thus, complex stress regimes influencing
869 submarginal and proglacial materials may also be required to explain the full suite of arcuate ridge
870 and modification features in crater interior ice deposits that we observe.

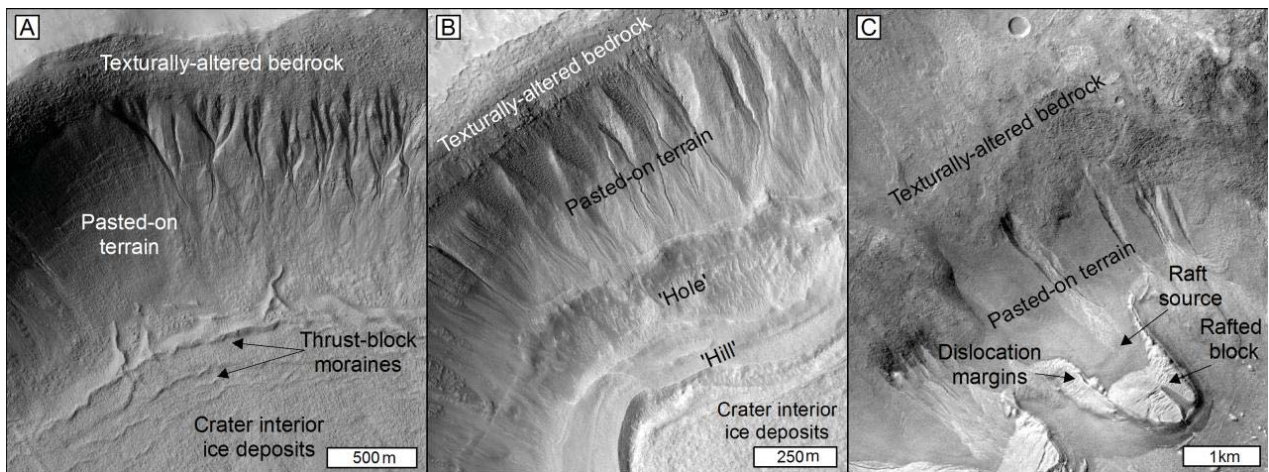
871 Landforms associated with glaciotectionic deformation of material at or below glacier margins on
872 Earth include thrust block moraine ridges (Figs. 18A and C; e.g., Bennett, 2001), hill-hole pairs (Fig.
873 18D; e.g., Rise et al., 2016), and rafted megablocks (e.g., Bennett et al., 1999). We have identified in
874 our study possible examples of all three landform types, shown in Fig. 19. Longitudinal ridges within
875 crater interior ice-deposit surfaces (Fig. 19), proximal to the spatulate depressions, could represent
876 faulting of proglacial materials under stresses induced by crater wall glaciation (Figs. 18A, 19A).



877

878 **Fig. 18.** Examples of arcuate glaciotectonic landforms on Earth. (A) Oblique view of active thrusts
 879 within proglacial sediments of a piedmont glacier, Canada (taken from Evans, 2007). No scale was
 880 provided in the original publication, but we estimate the valley floor to be 1.5 km wide. (B) Oblique
 881 view of small seasonal glaciotectonic moraines at Briedamerkurjökull, Iceland illustrating the highly
 882 arcuate nature of some moraine complexes in front of a broad glacier terminus (taken from Bennett,
 883 2001). No scale was provided in the original publication, but such seasonal ridges are typically metres
 884 in width. (C) A thrust-block moraine comprising lacustrine sediments (Fitzsimons, 1996) at the
 885 terminus of Wright Lower Glacier, Antarctic Dry Valleys. The ridge is ~80 m wide at its widest point
 886 and north is toward the top (Digital Earth image from Google Earth). (D) Marine bathymetry showing
 887 an arcuate hill-hole pair on the Norwegian continental shelf (taken from Rise et al., 2016). For
 888 interpretation of the references to colour in this figure legend, the reader is referred to the web
 889 version of this article.

890



891

892 **Fig. 19.** Possible glaciotectonic landforms on Mars. (A) Possible examples of glaciotectonic landforms
893 beyond the foot of the glaciated crater wall, possibly analogous to the proglacial thrust block
894 moraines in Fig. 18A. HiRISE image ESP_020051_1420. (B) Example of lower crater wall hole which
895 truncates gully fans upslope of arcuate ridge (hill) at CCF margins. This may be analogous to hill-hole
896 pairs on Earth (e.g., Fig. 18D). HiRISE image PSP_003253_1405. (C) Possible example of a rafted
897 megablock at the foot of a slope within the rim of the Argyre impact basin. Note that this feature is
898 not within an impact crater included in our slope measurement survey. However, it provides the best
899 type example of similar features observed within impact craters and illustrates the same
900 relationships between texturally altered bedrock, pasted-on terrain, and arcuate ridges. The bright
901 lobate plateau appears to have been mobilized downslope along shear planes that are visible within
902 upslope lateral extensions of the block. Its plateau morphology, gently sloping upslope margin, steep
903 downslope margin, and clear source region leads us to propose that it represents a rafted megablock
904 mobilised by stresses induced below the margin of a glacier. It does not appear to have been
905 distorted or reworked. Hence we favour a raft origin over a hill-hole origin. An alternate
906 interpretation is that it represents an extant lobe of glacier ice. CTX image G11_022685_1402.

907 Cold-based glaciers on Earth are rarely associated with prominent moraine ridges at their margins.

908 However, despite subfreezing temperatures and deep, continuous permafrost in the Antarctic Dry

909 Valleys, Fitzsimons (1996) described several arcuate, sharp-crested, ice-cored moraine ridges at the
910 entry points of glaciers into saline proglacial lakes (Fig. 18C). Salinity of pore water exerts a critical
911 control upon the deformability of sediments and glacitectonic moraine formation at the cold
912 margins of glaciers in permafrost environments on Earth (Etzel Müller et al., 1996). Thus, while we do
913 not invoke saline proglacial lakes such as those in the Antarctic Dry Valleys (Fitzsimons, 1996) to
914 explain the origin of the arcuate ridges on Mars, we do suggest that salinity of the martian substrate
915 may play an essential role in generating deformable glacier beds despite subfreezing temperatures.
916 Salts, such as perchlorates and sulphates capable of freezing-point depression, are known to be
917 abundant on the martian surface from in situ investigations (e.g., Hecht et al., 2009) and orbital
918 spectral observations (e.g., Langevin et al., 2005; Massé et al., 2010) as well as from theoretical
919 considerations of surface geochemistry (e.g., Chevrier and Altheide, 2008; Chevrier et al., 2009;
920 Toner et al., 2014). Generation of small interstitial meltwater films at the interfaces between
921 impurities and ice crystals within the basal ice of cold-based glaciers in the Antarctic Dry Valleys has
922 been posited as a mechanism for permitting basal sliding at rates that, while negligible on small
923 timescales, can be significant over very long timescales (see section 4.1 above; Cuffey et al., 2000).
924 Abundant salts detected on Mars' surface could therefore provide a ready supply of impurities to
925 glacier beds and permit generation of interstitial meltwater films and basal sliding at subfreezing
926 temperatures. It has been suggested that this mechanism could be effective at temperatures down
927 to -30°C on Earth (Cuffey et al., 2000).

928 Our results only provide evidence for meltwater generation in the latest phase of glaciation (~5-10
929 Ma) and only for small, thin glaciers rather than the more ancient VFF. Although recent studies have
930 identified compelling evidence for rare, localised occurrences of past basal melting of glaciers on
931 Mars, in relation to existing VFF (e.g., Hubbard et al., 2011; Gallagher and Balme, 2015; Butcher et
932 al., 2017) and to past late Amazonian-epoch glaciation (e.g., Scanlon et al., 2014, 2015), a majority of
933 Amazonian-aged glacial assemblages on Mars provide little apparent evidence for the role of
934 meltwater (e.g., Head and Marchant, 2003; Marchant and Head, 2007). Fassett et al. (2010)

935 attributed rare valleys on VFF surfaces to localised melting from the reflection of solar insolation by
936 adjacent topography, rather than melting induced by bulk VFF thermal regime or climate. Hence,
937 Amazonian glaciation on Mars was likely predominantly cold-based and the closest terrestrial
938 analogues to existing VFF are thought to be cold-based debris-covered glaciers in the Antarctic Dry
939 Valleys (e.g., Head and Marchant, 2003; Hambrey and Fitzsimons, 2010).

940 In summary, the landscape assemblage described here provides the first evidence that liquid water
941 could indeed have played a role in late Amazonian glaciation. Although the quantities were likely
942 relatively limited, the geographical distribution was widespread: pasted-on terrain and texturally
943 altered bedrock are found on almost every sloping surface in the mid-latitudes. We do not support
944 the production of large quantities of meltwater as this would have obliterated the arcuate ridges
945 and even the pasted-on terrain and would have produced a suite of landforms akin to wet-based
946 glaciers on Earth (e.g., eskers, meltwater channels, hummocky moraines).

947 *4.3. Glaciation and erosion rates*

948 Comparisons of sediment fluxes from glaciated and nonglaciated basins by Hallet et al. (1996) found
949 that, on Earth, glaciation is *unsurpassed* in its ability to erode and mobilize sediment compared to
950 other erosional mechanisms operating over similar timescales. They attribute comparable sediment
951 yields from lowland rivers draining marine clays in nonglaciated basins to high susceptibility of these
952 clays to mass wasting and, in particular, gullyng (Hallet et al., 1996). The discussion above
953 demonstrates that even cold-based glaciers can erode their beds but at rates that are orders of
954 magnitude lower than erosion by wet-based glaciation. Cold-based polar glaciers typically erode
955 their beds at rates of $\sim 10^1$ m My⁻¹ (Hallet et al., 1996; Cuffey et al., 2000), compared to rates of
956 $\sim 10^3$ m My⁻¹ for small temperate valley glaciers in the Swiss Alps and 10^4 - 10^5 m My⁻¹ beneath large,
957 fast-flowing temperate glaciers in southeast Alaska (Hallet et al., 1996). However, when integrated
958 over long timescales, the cumulative geomorphic influence of cold-based glaciation may be
959 detectable; Cuffey et al. (2000) estimated that 10 to 30 m of erosion may have occurred in the U-

960 shaped valley occupied by Meserve glacier on Earth, over a period of ~ 10 My of cold-based
961 glaciation. Rates of glacial erosion are greatly enhanced in the presence of weak (e.g., sedimentary
962 or highly fractured) bedrock, particularly in tectonically active regions such as southeast Alaska and
963 the Himalaya (Hallet et al., 1996).

964 The fastest retreat rates found in this study of $\sim 10^2$ m My⁻¹ (Table 2; Fig. 15) exceed the expected
965 erosion rate on Earth for cold-based glaciers and are instead equivalent to erosion rates for
966 temperate valley glaciers where melt is abundant. Similar erosion rates for icecaps and glaciers are
967 predicted over 0.1 Ma timescales on Earth for newly constructed volcanic provinces (e.g., Iceland,
968 Geirsdóttir et al., 2007) or actively uplifting areas (e.g., New Zealand, Egholm et al., 2012). Our
969 highest erosion rates are above those estimated by Levy et al. (2016) for VFF: namely 10^2 - 10^1 m My⁻¹
970 of erosion. In our study headwall retreat rates $< 10^1$ m My⁻¹ are only found for craters with ages
971 > 5 Ma, and these craters have similar headwall retreat rates to those estimated by Levy et al. (2016).

972 For comparison, crater erosion rates by aeolian processes in equatorial regions of Mars are
973 estimated to be on the order of 10^0 - 10^1 m My⁻¹ over 0.5-2 Ma timescales (Golombek et al., 2014a),
974 decreasing with increasing temporal baseline. The decrease in erosion rate over time reflects the
975 diffusive nature of the dry denudation processes, and the diffusivity is 10^2 - 10^3 times lower than
976 terrestrial values and similar to that on the Moon (Golombek et al., 2006; Fassett and Thomson,
977 2014; Golombek et al., 2014a).

978 A single (recent) event of intense glacial erosion related to the presence of the pasted-on terrain, is
979 one way in which to explain the parallel trends on Fig. 15E. This does not preclude the possibility
980 that previous erosion events have occurred and are either, too ancient to be distinguishable from
981 the diffusive decline in crater bedrock outcrop slopes or occurred earlier than our studied timeframe
982 and been overprinted by this later event — a notion supported by the slightly elevated erosion rates
983 expressed by craters older than 1 Ga in Fig. 15F. As shown in Fig. 15B, the erosion intensity seems
984 independent of slope — which is not the case for diffusion-like processes that tend to diminish in

985 intensity with decreasing slope — hence that the glaciated craters would follow this trend seems
986 improbable unless it was imposed by the ongoing background crater degradation. If punctuated
987 accelerated erosion by glaciers was happening throughout a crater’s history, we would expect the
988 glaciated crater walls and unmodified glacier walls to diverge with crater age on Fig. 15E and the
989 glaciated crater walls to have increasing erosion rates with time in Fig. 15F, which is not the case. If
990 we assume a single 500-ky event of glaciation we see that the glacial retreat rates become
991 independent of age with values of $\sim 10^2$ m My⁻¹ (Fig. 15F). This single event is supported by the
992 morphology of the arcuate ridges — they are never found as multiple superposed sets. Because our
993 youngest crater with pasted-on terrain dates to 0.5 (0.4-50) Ma, this one event is likely to be shorter;
994 hence, the retreat rates are probably an order of magnitude higher. Further that the bedrock slope
995 lowering extends underneath the pasted-on terrain that we see today is likely, another factor
996 leading to a potential underestimation of the headwall retreat rate. Given the large uncertainty on
997 the age of this crater and that our other craters with pasted-on terrain date to >10 Ma (consistent
998 also with dating presented in de Haas et al., 2018), these dates point to the “glacial event” having
999 occurred between 5 and 10 Ma. It seems logical to link it to the shift in mean obliquity that
1000 happened ~ 5 Ma, but further dating on young craters with pasted-on terrain and texturally altered
1001 bedrock would be needed to substantiate this claim. Previous work has linked this decrease in mean
1002 obliquity as the transition point between *glacial* and *interglacial* Mars with recent high obliquity
1003 excursions representing mini-ice-ages (Head et al., 2003; Madeleine et al., 2014). This shift marks a
1004 transition from a period with lower average surface temperatures to higher average surface
1005 temperatures, particularly in the mid-latitudes (e.g., Kreslavsky et al., 2008).

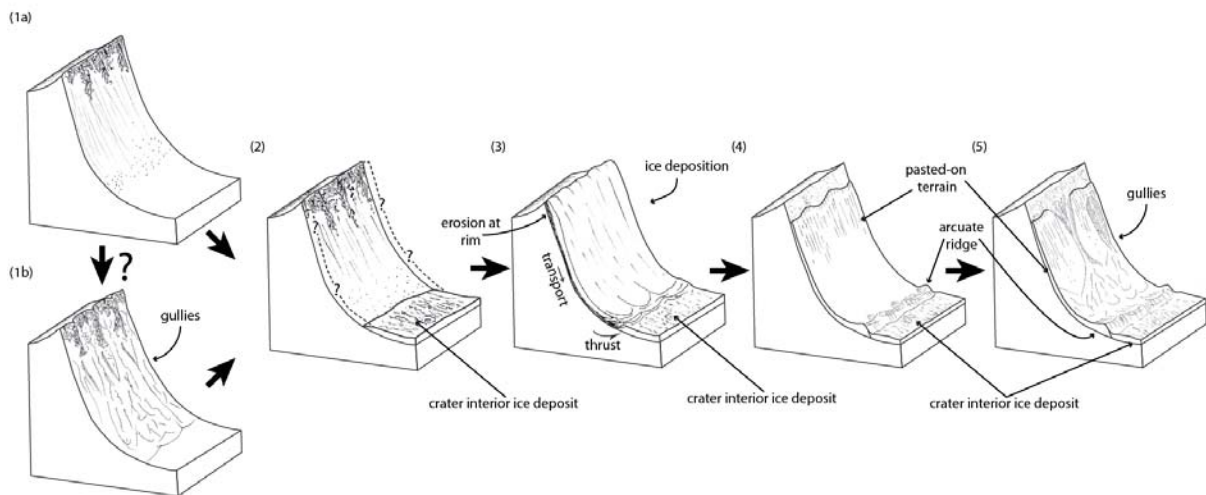
1006 This general shift in climate conditions is supported by observational evidence as follows. Berman et
1007 al. (2009) reported that GLFs (termed lobate forms by these authors) tend to be found in smaller
1008 craters (<70 km in diameter) with no particular trend observed for other crater interior ice deposits.
1009 This crater diameter dependency was further explored in Fassett et al. (2014) who found that
1010 synglacial craters tend to be of smaller diameter than preglacial craters. They concluded that this

1011 observation is a function of the age and duration of the *glacial* epoch superposed on the crater
1012 production function. In agreement, recent work by Hepburn et al. (2018) reported that smaller
1013 glacial forms are significantly younger than LDA, LVF, or crater interior ice deposits (including CCF),
1014 as detailed in section 1.1. That the present epoch is one of glacial retreat is broadly acknowledged
1015 (Brough et al., 2016a). Our results agree with the broad overall picture from this previous work; that
1016 a long glacial epoch that comprised several phases of glacial growth was followed by glacial retreat,
1017 dominating present conditions. Our results also show that that crater interior ice deposits predate
1018 the pasted-on terrain (as noted by Levy et al., 2009b) and that the presence or absence of the crater
1019 interior ice deposits do not seem to have any relation to the amount of erosion recorded by these
1020 craters (Fig. 20).

1021 However, the question still remains as to how the terminal stages of an ice age under cold and dry
1022 Amazonian conditions could generate the conditions for production of meltwater, even if only in
1023 small amounts. Other authors examining glacial melt on Mars (e.g., Fassett et al., 2010; Hobbey et al.,
1024 2014; Scanlon et al., 2015; Butcher et al., 2017) have reviewed and proposed the following
1025 scenarios: (i) orbital-driven increases in surface temperature with current atmospheric
1026 composition/pressure, (ii) increased surface snow/ice thickness reducing heat loss to the
1027 atmosphere, (iii) geothermal heat increase, (iv) atmospheric change inducing heating caused by
1028 impact or volcanic events (greenhouse effect), and (v) direct heating caused by impact events.

1029 We can immediately rule out direct heating caused by impact events and geothermal heat increase
1030 because both of these would be expected to be expressed locally rather than at a global scale as our
1031 results indicate. Insulation by thick snow and ice are unlikely because the ice thicknesses in our
1032 proposed glaciers is almost certainly less than the >1-km thickness estimated for this effect to
1033 become important (e.g., Carr and Head, 2003) — because their moraines are <100 m in height.
1034 Atmospheric changes caused by impacts require a large enough impact to substantially change the
1035 atmosphere and such impacts are estimated to produce craters hundreds of kilometres in size

1036 (Segura et al., 2008). According to the Hartmann and Neukum (2001) production function, this
1037 should happen on Mars every ~ 1 Ga and Lyot Crater at 215 km in diameter is generally thought to be
1038 the youngest example, whose age is generally agreed to be older than 1 Ga (Harrison et al., 2010).
1039 This leaves the two remaining scenarios of atmospheric heating through secular changes in orbital
1040 parameters and/or injection of greenhouse gases from volcanic events. Alone, changes in orbital
1041 parameters are not anticipated to cause substantial surface warming with the current atmospheric
1042 density and composition (e.g., Kreslavsky et al., 2008; Mansfield et al., 2018). Volcanism is thought
1043 to have continued into the Amazonian (Neukum et al., 2004; Werner, 2009) with estimates as recent
1044 as the last few Ma for Elysium Mons (Vaucher et al., 2009) and low shields in Tharsis (Hauber et al.,
1045 2011b), which could tie with the timing of the glacial erosion event. However, further work would be
1046 needed on the dating and duration of the glacial erosion event to assess if recent volcanic activity
1047 would be of sufficient intensity and duration to cause the required temperature increase. As noted
1048 in section 4.3, another factor likely to be playing a role is the abundant presence of salts on Mars,
1049 which would diminish the perturbation required by either climate or volcanic-induced atmospheric
1050 change to generate subglacial meltwater. We suggest these possibilities as fruitful avenues to direct
1051 future studies.



1052
 1053 **Fig. 20.** Schematic diagram illustrating the development of pasted-on terrain, arcuate ridges, and
 1054 gullies on crater walls on Mars. Step 1a is a pristine crater wall with exposed bedrock, and step 1b
 1055 shows the development of gullies. Steps 1a and 1b could lead to step 2 onward, but evidence of the
 1056 starting configuration is erased at step 3. At step 2 crater floor ice deposits are laid down by some
 1057 process that does not significantly erode the crater wall. At step 3 glacial erosion occurs, scouring rock
 1058 from the crater rim and transporting the debris downslope, depositing it as till (pasted-on terrain). On
 1059 interaction with the preexisting ice deposit, the moist till is thrust-up forming moraines (arcuate
 1060 ridges). After the glacier has sublimated away at step 4 onward, the crater wall hosts pasted-on
 1061 terrain (till) and arcuate ridges (moraines), and later (step 5) gullies can be formed into these easily
 1062 erodible deposits.

1063 5. Gully-glacier interactions on Mars

1064 Our headwall retreat rate calculations reveal that, unlike on Earth, glacial and gully headwall retreat
 1065 rates appear to be broadly equivalent on Mars. This means that even if gullies were present on the
 1066 crater walls prior to the glacial erosion event, then that much evidence remains is unlikely (Fig. 20).
 1067 Hence, we hypothesise that all gullies that are visible today were created after this glacial erosion
 1068 event (~5 Ma), consistent with other dating studies. Craters that postdate the glacial erosion event
 1069 possess gullies that solely erode bedrock, and craters that predate the glacial erosion event possess
 1070 gullies that primarily erode into pasted-on terrain (Fig. 20). Gullies in both contexts have similar-

1071 sized alcoves (de Haas et al., 2018). Our estimates of headwall retreat by gullies are similar to those
1072 calculated by de Haas et al. (2015b) using a different method. These estimates of martian gully
1073 retreat align with estimates of terrestrial rockwall retreat rates in Arctic, Nordic, and Alpine
1074 environments of 10^1 - 10^4 m My⁻¹ (see summary in Fig. 12 of de Haas et al., 2015b), strengthening the
1075 case made by these authors for water as a catalyst for backweathering in these craters.

1076 We suggest that pasted-on terrain on formerly glaciated crater walls could partly represent
1077 subglacial deposits, which would explain the frequent observation of downslope lineations present
1078 on the surface of this unit. We also note that the pasted-on terrain also often expresses a
1079 polygonised surface texture and that it blends gradually into more glacier-like textures lower on the
1080 crater wall (topographically and morphologically), suggesting that it could also represent part of a
1081 glacier-like body in and of itself. This is supported by measurements made by Conway and Balme
1082 (2014) who found that the pasted-on terrain contains between 46% and 95% ice by volume. Also
1083 Dickson et al. (2015) found that the pasted-on terrain often expressed contouring fractures, which
1084 cross-cut and are superposed by gully-fans (located mid- to lower-crater wall) hinting that this unit
1085 could still be in motion. The topographic relaxation of gully incisions as they become polygonised
1086 (Conway and Balme, 2014) supports this notion. The fact that our work shows that gully incisions
1087 tend to deepen toward higher latitudes in both hemispheres could therefore be interpreted as
1088 evidence that the pasted-on terrain thickens and has higher ice content at higher latitudes (Fig. 9), a
1089 concept supported by the recent finding of near-pure surface ice deposits by Dundas et al. (2018) at
1090 latitudes of 55°. Our work reveals that glacial erosion must have liberated a substantial quantity of
1091 rock debris, and logically this should be found downslope within the pasted-on terrain. Further, the
1092 surface textures expressed by the pasted-on terrain do not support debris covered ice, as suggested
1093 for the martian VFF (lack of pitting, expanded craters, holds cracks, and polygons) (Soare et al.,
1094 2017). Hence, we suggest that these are ice-rich sediments rather than sediments covering ice. De
1095 Haas et al. (2015c) found that incisions into gully-fans downslope of pasted-on terrain were much
1096 less likely to expose metre-scale boulders than incisions whose catchments included exposed

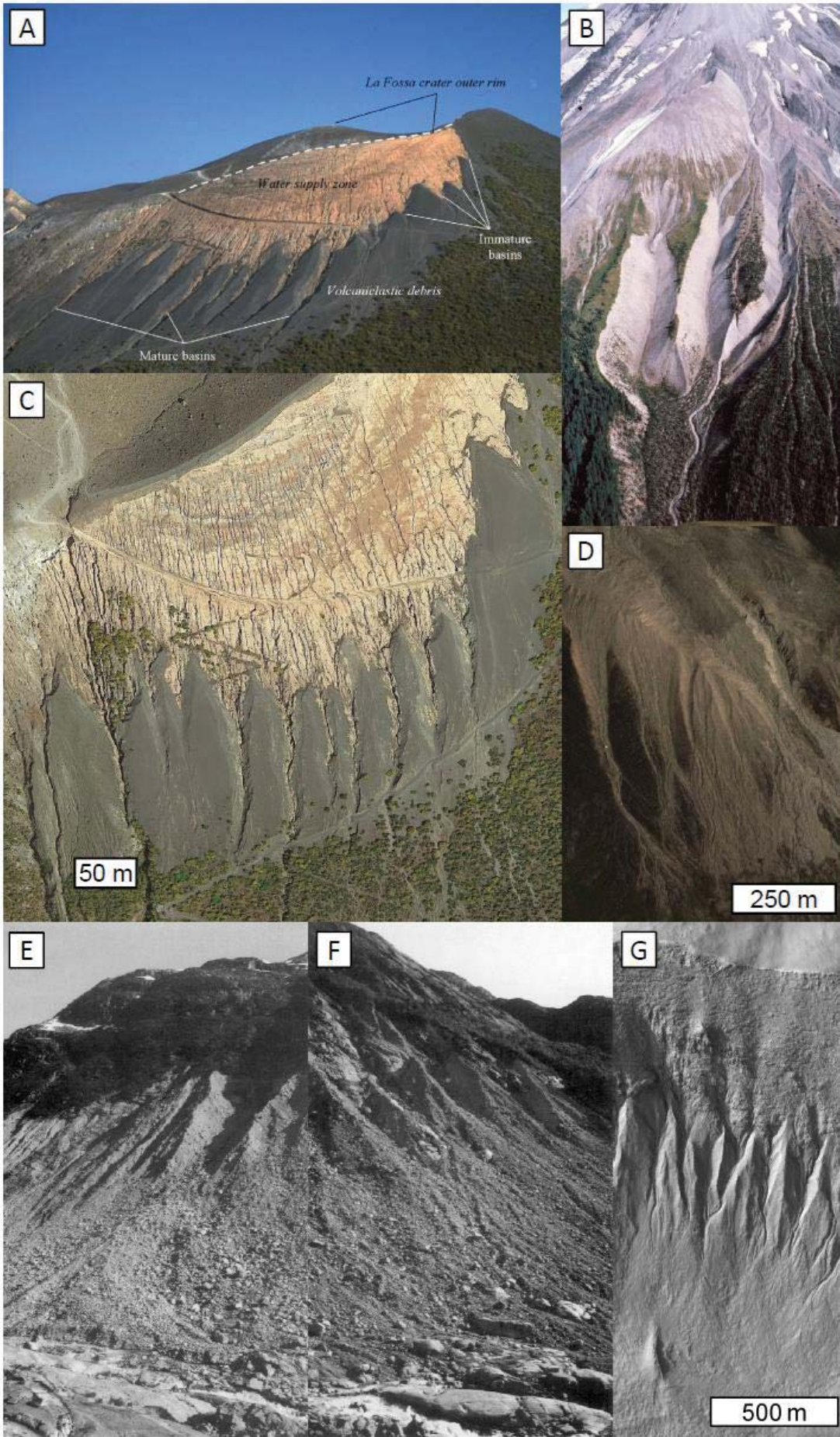
1097 bedrock, implying that the sediments within the pasted-on terrain tend to be finer grained than
1098 those liberated directly from rockwalls.

1099 Our hypothesis for a single *glacial event* at around 5-10 Ma fits with the interpretation made by
1100 Dickson et al. (2015) of one episode of pasted-on terrain emplacement based on the stratigraphic
1101 relationships between gullies and pasted-on terrain (they call it 'LDM'). Conway and Balme (2014)
1102 also found no evidence for layering within the incisions into the pasted-on terrain. Dickson et al.
1103 (2015) noted that inverted gully channels and that fans mainly occur on equator-facing slopes
1104 between 40°S and 50°S, where our study shows that the pasted-on terrain is at its thickest. We
1105 hypothesise that these inverted topographies are produced by the loss of ice: gullies could have
1106 deposited debris on top of glacial ice and when the surrounding ice has been lost, the gully deposits
1107 (channels and fans) preserve some of the ice and protrude from the surrounding surface. These
1108 inverted gullies appear beheaded, not because their upper half has been buried (as suggested by
1109 Dickson et al., 2015), but because it has been eroded away. If too much ice is removed then the
1110 inverted gullies disappear (explaining their paucity where the pasted-on terrain is thin).

1111 The observation by Dickson et al. (2015) that some gullies appear to be buried by the pasted-on
1112 terrain seems to contradict our hypothesis that this unit represents a glacial till under which a
1113 significant erosive event has occurred — such deposits should have been removed. However,
1114 Dickson et al. (2015) only presented two examples of gully deposits being revealed from under the
1115 pasted-on terrain, and both of these could be interpreted as being degraded gully deposits atop
1116 pasted-on terrain. Another possibility to explain these observations is that another mantling deposit
1117 superposes the pasted-on terrain, a possibility supported by occasional observations of pasted-on
1118 terrain emerging from under a light-toned mantle (e.g., Fig. 71): a phenomenon also observed by
1119 Soare et al. (2017). Many authors have noticed that alcoves appear from under the pasted-on terrain
1120 (e.g., Christensen, 2003; Dickson et al., 2015; de Haas et al., 2018), yet the link between these
1121 alcoves and preexisting gullies is uncertain and only circumstantial. Hence, we believe that the

1122 population of gullies we see today postdates the glacial event that laid down the pasted-on terrain
1123 (Fig. 20).

1124 Our hypothesis that the pasted-on terrain is predominantly an ice-rich glacial till is supported by the
1125 way in which gullies erode into it. Gullies always have a V-shaped incision, or chute, whenever they
1126 encounter the pasted-on terrain. Such incisions are also common in similar-scaled systems on Earth
1127 when the sediment they incise into is granular and loose (Fig. 21). The examples we present here are
1128 from glacial and volcanoclastic terrains where the loose surface deposits are quickly stripped by
1129 flowing liquid water, often involving debris flows, producing sharp V-shaped incisions whose floors
1130 are located upon more resistant materials (Ballantyne and Benn, 1994; Ferrucci et al., 2005). The V-
1131 shaped incisions into glacial moraine on Earth occur during the paraglacial phase of glacial retreat
1132 and can penetrate through ice-rich moraine deposits (e.g., Bennett et al., 2000; Ewertowski and
1133 Tomczyk, 2015). Where stripping of the pasted-on terrain is advanced, rills can be seen converging
1134 over the bedrock to meet at the incisions on Mars (Figs. 6F, 7C, 21G), and a similar phenomenon
1135 happens on Earth (Fig. 21C). Isolated gullies within the pasted-on terrain could originate by flow of
1136 water at the lower boundary of the pasted-on terrain, initiating collapse and upslope propagation of
1137 the incision (see Fig. 21B for a terrestrial example and Figs. 5D and 7G for martian ones). Our work
1138 shows that gullies can substantially erode the rims of impact craters without pasted-on terrain.
1139 However, substantial incision into the bedrock by gullies is rare in the presence of pasted-on terrain
1140 (e.g., Dickson et al., 2015). One of the reasons could be that gully-forming processes are enhanced
1141 on steep slopes, hence the lowering of the crater wall slope associated with the pasted-on terrain
1142 reduces the erosive capacity of the gullies. Second, much of the erosional potential of the gullies
1143 could be taken up in stripping the pasted-on terrain. Third, the pasted-on terrain could be forming a
1144 protective layer and inhibiting the continuous weathering of the bedrock needed to initiate and
1145 grow the alcoves.



1147 **Fig. 21.** *Gullies incising into unconsolidated glacial and volcanic sediments on Earth. (A) Debris flows*
1148 *dissecting the unconsolidated volcanoclastic deposits on the side of La Fossa Cone on Vulcano Island,*
1149 *Italy. Image from Ferrucci et al. (2005). (B) Photo from 1979 of gullies incised into the west side of Mt*
1150 *St Helens from the USGS photo library, id: mhob0056. (C) Google Earth image of the same suite of*
1151 *gullies as shown in (A). (D) Google Earth image of gullies cut into moraine from glaciers in west*
1152 *Greenland (70.381N, 52.255W). (E-F) Slopes in Fåbergstølsdalen where gullies cut into the drift*
1153 *deposits and where debris flow deposits are seen in the fans in the foreground. Images from*
1154 *Ballantyne and Benn (1994), Figs. 4 and 5. (G) Gullies stripping pasted-on terrain on Mars, HiRISE*
1155 *image PSP_002066_1425.*

1156 Our previous work has shown that dense concentrations of extant VFF are anticorrelated with dense
1157 gully populations (Conway et al., 2017), yet gullies are intimately associated with pasted-on terrain
1158 and arcuate ridges (Christensen, 2003; e.g., Berman et al., 2005; Head et al., 2008; Dickson et al.,
1159 2015). In this work we have presented evidence in favour of small quantities of meltwater being
1160 involved in lowering crater wall slopes where texturally altered bedrock and pasted-on terrain are
1161 present. Does this suggest a role of water in the formation of martian gullies? Unfortunately, the
1162 findings reported in this paper do not provide any resolution to the ongoing debate between CO₂
1163 sublimation (e.g., Dundas et al., 2017, and references therein) and meltwater generation (e.g.,
1164 Conway et al., 2018b, 2018a, and references therein) for forming martian gullies. A meltwater
1165 hypothesis for this recent glacial erosion event on Mars leads to the reasonable assumption that
1166 gullies can be produced via a similar yet less widespread mechanism, which has been occurring
1167 periodically since this glacial erosion event. However, CO₂ and H₂O ices are often found together
1168 (Vincendon et al., 2010; e.g., Vincendon, 2015), and without more detailed knowledge on how CO₂
1169 sublimation might engender sediment transport assessing whether the morphological evidence fits
1170 with this transport mechanism remains difficult (Conway et al., 2018b).

1171 Finally, texturally altered bedrock and pasted-on terrain are more spatially widespread than gullies
1172 and occur at low slope angles where gullies are not found. Hence, even if gullies do not represent
1173 episodes of flowing liquid water at the surface of Mars, the wet-based glacial erosion event that we
1174 report on here does present evidence for a significant occurrence of widespread, recent melt on
1175 Mars.

1176 6. Conclusions

1177 We have found evidence for extensive crater wall erosion by very recent (~5-10 Ma) small-scale
1178 glaciers on Mars that we estimate achieved headwall retreat rates up to $\sim 10^2$ m My⁻¹. This erosion
1179 rate is of the same order of magnitude as the landscape change brought about by the youngest
1180 gullies (<1 Ma) incised into bedrock on Mars.

1181 We posit that the erosion of these crater walls is driven by small amounts of subglacial melt, possibly
1182 favoured by brines, using the following lines of evidence based on Earth analogy:

- 1183 • Recent headwall retreat rates are equivalent to wet-based rather than cold-based glaciers
1184 on Earth.
- 1185 • Formation of arcuate ridges and associated deformational features at the base of crater
1186 walls involved a component of glacial tectonism, which requires pore water.
- 1187 • The presence of texturally altered bedrock, which is indicative of ice-segregation and frost-
1188 shattering on Earth.
- 1189 • Pasted-on terrain, found topographically below the texturally altered bedrock represents
1190 glacial till deposits. We suggest the downslope lineations often found on this deposit could
1191 be glacial in origin. The till interpretation is supported by the nature of the gully erosion. This
1192 provides an alternate explanation to that the pasted-on terrain is simply a thicker version of
1193 the latitude-dependant mantle — an airfall deposit of ice nucleated on dust (e.g., Mustard et
1194 al., 2001; Kreslavsky and Head, 2002).

1195 Our results suggest that the accelerated crater wall slope reduction was brought about in a pulse of
1196 erosion, which may have occurred sometime 5-10 Ma coincident with a significant climate shift
1197 brought about by a change in Mars' mean orbital obliquity. This shift alone probably cannot explain
1198 the generation of basal pore water, and we suggest that the influence of volcanic greenhouse gases
1199 and/or the influence of salts as additional potential contributing factors. Gullies seem to postdate
1200 this event and have caused significant reworking of the pasted-on terrain on the steepest slopes.
1201 The magnitude of the glacial erosion is the same as that brought about by the youngest modern
1202 gullies, and hence evidence of gullies predating the glacial erosion is likely difficult to find. Although
1203 our results cannot directly advocate that gullies are produced by meltwater runoff, the nature of this
1204 wet glacial event does provide strong evidence for meltwater generation in Mars' recent history.

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Table 1 Summary of the craters included in this study, elevation data, and their ages^a

Crater details		Digital terrain model (DTM) information							Dating						
		Latitude	Longitude	Diameter (km)	Features	Type	Method	Credit	Image1	Resolution (m)	Image2	Resolution (m)	Convergence (°)	Vertical error (m)	Source
Zumba	28.65°S	226.90°E	3	talus only	unmodified	SS	OU	PSP_002118_1510	0.3	PSP_003608_1510	0.28	18.1	0.2	Hartmann et al. (2010), Schon et al. (2012)	(0.1-0.8) Ma
Corintho	16.95°N	141.70°E	14	talus only	unmodified	SS	UoA	PSP_003611_1970	0.3	PSP_004244_1970	0.3	18.1	0.2	Golombek et al. (2014b)	(0.1-3.0) Ma
Qara	16.41°N	209.7°E	3	talus only	unmodified	SS	UoA	PSP_005837_1965	0.3	PSP_005837_1965	0.32	20.1	0.2	de Haas et al. (2015b)	5.3 (4-8) Ma
Kenge	16.37°S	102.96°E	6	talus only	unmodified	SS	LPG	ESP_011893_1635	0.3	ESP_012315_1635	0.27	13.1	0.2	Figure 12K	12.6 (5-30) Ma
Topola	15.82°N	267.8°E	8.0	talus only	unmodified	SS	LPG	ESP_023662_1960	0.6	ESP_023372_1960	0.55	25.5	0.3	Figure 12J	15 (8-100) Ma
Yelwa	31.11°N	212.39°E	8	talus only	unmodified	ASP ²	LPG	ESP_020539_2115	0.6	ESP_020117_2115	0.58	16.7	0.4	Figure 12D	20 (5-100) Ma
Kilmia	24.09°S	59.50°E	7	talus only	unmodified	SS	UoA	ESP_026861_1555	0.3	ESP_027349_1555	0.27	24.4	0.1	Figure 12Q	1.1 (0.6-3) Ga
Crater A	50.19°N	175.51°E	2	small bedrock gullies, facing slope	gullied	SS	UoA	ESP_025366_2305	0.3	ESP_025498_2305	0.35	25.2	0.2	Figure 12C	0.1 (0.01-0.7) Ma
Istok	45.11°S	274.2°E	5	bedrock gullies	gullied	SS	OU	PSP_006837_1345	0.3	PSP_007127_1345	0.26	20.1	0.1	Johnsson et al. (2014)	0.19 (0.1-1.0) Ma
Galap	37.66°S	192.93°E	6	bedrock gullies	gullied	SS	OU	PSP_003939_1420	0.3	PSP_003939_1420	0.29	21.7	0.2	de Haas et al. (2015b)	6.5 (5-9) Ma
Jaisalmer	33.47°N	84.15°E	14	pasted on only, facing slope	glaciated	SS	NHM	PSP_009098_2140	0.3	PSP_009164_2140	0.32	19	0.2	Figure 12B	0.5 (0.4-50) Ma
Crater B	27.35°S	59.13°E	7	pasted-on only, facing slope	glaciated	SS	NHM	ESP_014400_1525	0.3	ESP_020703_1525	0.27	13.9	0.2	Figure 12F	2.15 (0.5-15) Ma
Crater C	36.56°N	155.46°E	9	pasted-on, arcuate ridges, facing slope	glaciated	SS	NHM	PSP_010559_2170	0.6	ESP_016005_2170	0.6	15.8	0.4	Figure 12I	2.43 (0.3-5) Ma

Talu	40.3°S	20.1°E	9	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	OU	ESP_0111672_1395	0.3	ESP_0111817_1395	0.26	15.7	0.2	de Haas et al. (2018)	13 (10–22) Ma
Nybyen	37.03°S	343.36°E	6.0	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	UoA	ESP_011436_1425	0.3	PSP_006663_1425	0.25	14.5	0.2	Figure 12G	110 (5–300) Ma
Taltal	39.5°S	234.4°E	10	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	LPG	ESP_031259_1400	0.5	ESP_037074_1400	0.5	5.9	1	de Haas et al. (2018)	220 (100–400) Ma
Taltal	39.51°S	234.17°E	10	gullies, pasted-on, arcuate ridges, facing slope	glaciated	ASP ¹	LPG	ESP_021989_1400	0.3	ESP_021712_1400	0.25	25.5	0.1	"	"
Dechu	42.24°S	202.19°E	0.0	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	UoA	ESP_023546_1375	0.6	ESP_023612_1375	0.51	19.4	0.3	Figure 12D	305 (80–600) Ma
Niquero	38.77°S	193.99°E	10.0	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	NHM	ESP_030443_1410	0.5	ESP_030021_1410	0.51	23.5	0.3	Figure 12H	1.3 (0.5–2.5) Ga
Corozal	38.7°S	159.4°E	8.0	gullies, pasted-on, arcuate ridges, facing slope	glaciated	SS	UoA	PSP_006261_1410	0.3	ESP_014093_1410	0.29	28.7	0.1	de Haas et al. (2018)	1.4 (0.8–2.5) Ga
Nqutu	37.9°S	169.6°E	20.0	gullies, pasted-on, arcuate ridges	glaciated	SS	Birk.	PSP_004085_1420	0.3	PSP_004019_1420	0.25	20.4	0.2	de Haas et al. (2018)	1.8 (1–3.5) Ga

^a Abbreviations: SS = SocetSet, ASP = Ames Stereo Pipeline, LPG = Laboratoire de Planétologie et Géodynamique, Birk. = Birkbeck University of London, NHM= Natural History Museum, London, UoA = University of Arizona, OU=Open University. In the method column 1 denotes that the ASP DTM was adjusted to HRSC H4234_0000 and 2 to H1525_0001.

Table 2 Summary of results for each site, including number of classes, slopes, exposed bedrock length, retreat distance, and retreat rate^a

Crater name	Number of classes				Mean inner wall slope (°)			Mean outer wall slope (°)		
	Inner wall	Outer wall	Gullied	Glaciated + gullied/glaciated	Pasted-on	Unmodified	Glaciated	Unmodified		
Zumba	gu(0),gg(0),gl(0),po(30),un(0)	un(30),gl(0)				39.25±1.09		9.43±4.19		
Corintho	gu(0),gg(0),gl(0),po(23),un(0)	un(23),gl(0)				40.66±5.59		21.54±9.96		
Qara	gu(0),gg(0),gl(0),po(26),un(0)	un(26),gl(0)				35.89±2.04		7.92±2.97		
Kenge	gu(0),gg(0),gl(0),po(32),un(0)	un(32),gl(0)				37.47±0.65		14.17±5.48		
Topola	gu(0),gg(0),gl(0),po(33),un(0)	un(33),gl(0)				38.45±1.15		19.38±3.30		
Yelwa	gu(0),gg(0),gl(0),po(34),un(0)	un(34),gl(0)				39.01±1.14		19.11±4.00		
Kilmia	gu(0),gg(0),gl(0),po(36),un(0)	un(36),gl(0)				26.53±2.15		10.64±5.18		
Crater A	gu(9),gg(0),gl(0),po(10),un(0)	un(19),gl(0)	39.68±1.29			37.06±2.54		26.70±3.97		
Istok	gu(20),gg(0),gl(0),po(22),un(0)	un(42),gl(0)	36.39±1.97			41.60±5.41		13.44±4.89		
Galap	gu(22),gg(0),gl(0),po(22),un(0)	un(44),gl(0)	33.35±1.48			37.25±1.94		16.18±7.16		
Jaisalmer	gu(0),gg(0),gl(10),po(16),un(13)	un(23),gl(16)		33.61±1.55	32.88±2.63	34.37±1.70	1.55±6.61	19.96±9.09		
Crater B	gu(0),gg(0),gl(5),po(28),un(10)	un(28),gl(15)		32.59±1.06	36.47±2.64	34.04±3.08	1.06±4.46	16.76±6.23		
Crater C	gu(0),gg(0),gl(0),po(12),un(16)	un(2),gl(26)		30.28±3.32		34.35±1.03	3.32±7.47	23.02±11.57		
Taltu	gu(0),gg(21),gl(0),po(12),un(0)	un(21),gl(12)		30.16±1.34		34.71±1.74	1.34±4.12	20.50±7.41		
Nybyen	gu(6),gg(15),gl(0),po(17),un(0)	un(20),gl(18)	37.61±2.21			36.88±2.12	1.36±4.62	20.72±5.85		
Taltal E	gu(0),gg(14),gl(0),po(16),un(0)	un(17),gl(13)		28.77±2.86		31.93±2.02	2.86±4.11	18.24±4.06		
Taltal W	gu(0),gg(0),gl(0),po(10),un(12)	un(12),gl(10)		24.75±2.14		33.06±1.79	2.14±5.79	17.80±5.99		
Dechu	gu(18),gg(45),gl(0),po(9),un(0)	un(22),gl(50)	32.18±3.14	27.83±2.87		29.93±1.79	2.87±5.64	4.04±3.15		
Niquero	gu(0),gg(13),gl(0),po(15),un(4)	un(17),gl(15)		27.79±2.88		34.48±1.47	2.88±2.58	17.44±11.21		
Corozal	gu(0),gg(19),gl(0),po(23),un(0)	un(17),gl(25)		26.68±2.11		31.04±2.11	2.11±3.09	16.12±6.37		
Nqutu	gu(0),gg(18),gl(0),po(0),un(0)	un(18),gl(0)		24.14±2.84				12.85±6.95		

Crater name	Exposed rock length (m)			
	Gullied	Glaciated + gullied/glaciated	Pasted-on	Unmodified
Zumba				138(+186/-37)
Corintho				245(+304/-132)
Qara				227(+97/-68)
Kenge				202(+215/-108)
Topola				212(+206/-68)
Yelwa				401(+376/-136)
Kilmia				610(+362/-512)
Crater A	118(+52/-15)			134(+47/-36)
Istok	338(+146/-185)			146(+123/-41)
Galap	508(+207/-113)			316(+203/-133)
Jaisalmer		438(+217/-254)	356(+335/-147)	264(+278/-67)
Crater B		453(+215/-161)	277(+108/-64)	361(+285/-173)
Crater C		192(+223/-73)		238(+257/-71)
Taltu		431(+260/-188)		423(+157/-131)
Nybyen	283(+151/-108)	333(+450/-179)		306(+188/-92)
Taltal E		253(+351/-187)		439(+350/-182)
Taltal W		121(+187/-29)		384(+245/-199)
Dechu	522(+213/-170)	454(+393/-329)		512(+156/-158)
Niquero		290(+604/-271)		297(+208/-179)
Corozal		714(+386/-270)		593(+406/-320)
Nqutu		583(+173/-291)		

Crater name	Retreat distance (m)			
	Unmodified compared to equatorial	Glacial compared to equatorial	Gullied compared to equatorial	Glacial compared to same crater
Zumba	3.42(+4.6/-0.91)			
Corintho	0			
Qara	30.94(+13.17/-9.22)			
Kenge	17.18(+18.27/-9.14)			
Topola	11.05(+10.72/-3.52)			
Yelwa	12.03(+11.28/-4.07)			
Kilmia	246.5(+146.08/-207.02)			
Crater A	13.21(+4.61/-3.54)		1.27(+0.44/-0.34)	0
Istok	0		17.57(+14.75/-4.92)	24.86(+20.87/-15.38)
Galap	29.16(+18.7/-12.24)		67.78(+43.46/-28.45)	42.55(+27.28/21.66)
Jaisalmer	48.57(+51.01/-12.39)	90.39(+44.87/-52.45)		12.24(+6.08/-7.1)
Crater B	69.96(+55.23/-33.44)	107.34(+51.06/-38.12)		24.36(+11.59/-8.65)
Crater C	43.85(+47.3/-13)	58.11(+67.56/-21.99)		27.96(+32.5/-10.58)
Taltu	73.28(+27.19/-22.73)	132.01(+79.56/-57.48)		69.42(+41.84/-30.23)
Nybyen	31.91(+19.57/-9.6)	81.42(+109.9/-43.72)		52.1(+70.32/-27.97)
Taltal E	112.45(+89.54/-46.49)	87.05(+120.85/-64.58)		30.1(+41.79/-22.33)
Taltal W	85.67(+54.68/-44.28)	54.38(+84.14/-12.89)		35.28(+54.58/-8.36)
Dechu	160.07(+48.84/-49.51)	167.97(+145.28/-121.54)		37.69(+32.6/-27.27)
Niquero	53.51(+37.46/-32.28)	107.41(+224.08/-100.41)		67.38(+140.58/-62.99)
Corozal	166.82(+114.31/-90.02)	285.58(+154.34/-108.21)		117.85(+63.69/-44.65)
Nqutu		271.06(+80.59/-135.17)		

Crater name	Retreat rate (m/My)				
	Unmodified compared to equatorial	Glacial compared to equatorial	Gullied compared to equatorial	Glacial compared to same crater	Gullied compared to same crater
Zumba	7.6(+10.22/-2.01)				
Corintho	0				
Qara	5.84(+2.48/-1.74)				
Kenge	1.36(+1.45/-0.73)				
Topola	0.74(+0.71/-0.23)				
Yelwa	0.6(+0.56/-0.2)				
Kilmia	0.22(+0.13/-0.19)				
Crater A	132.14(+46.09/-35.43)		12.71(+4.43/-3.41)		0
Istok	0		92.46(+77.62/-25.91)		35.78(+13.93/-41.74)
Galap	4.49(+2.88/-1.88)		10.43(+6.69/-4.38)		49.37(+35.72/-54.57)
Jaisalmer	97.14(+102.03/-24.78)	180.77(+89.74/-104.89)		24.49(+12.16/-14.21)	
Crater B	32.54(+25.69/-15.56)	49.92(+23.75/-17.73)		48.73(+23.18/-17.31)	
Crater C	18.04(+19.47/-5.35)	23.91(+27.8/-9.05)		55.92(+65.01/-21.16)	
Taltu	5.64(+2.09/-1.75)	10.15(+6.12/-4.42)		138.84(+83.68/-60.46)	
Nybyen	0.29(+0.18/-0.09)	0.74(+1/-0.4)		104.2(+140.64/-55.95)	
Taltal E	0.51(+0.41/-0.21)	0.4(+0.55/-0.29)		60.2(+83.57/-44.66)	
Taltal W	0.39(+0.25/-0.2)	0.25(+0.38/-0.06)		70.56(+109.16/-16.72)	
Dechu	0.52(+0.16/-0.16)	0.55(+0.48/-0.4)		75.38(+65.2/-54.55)	
Niquero	0.04(+0.03/-0.02)	0.08(+0.17/-0.08)		134.77(+281.16/-125.98)	
Corozal	0.12(+0.08/-0.06)	0.2(+0.11/-0.08)		235.7(+127.38/-89.31)	
Nqutu		0.15(+0.04/-0.08)			

^a Abbreviations: gu = gullied, gl = glaciated, un = un modified. For slopes standard deviations are provided as the uncertainty values and for bedrock length the variance is given. This variance is propagated into the retreat distance and retreat rate calculations. Zero retreat rates are given where the mean slope of the bedrock is greater than or equal to the comparison slope (opposite wall, or equatorial slope).