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1 Modeled subglacial water flow routing supports localized intrusive heating as a possible
2 cause of basal melting of Mars' south polar ice cap

3

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12 **Key Points:**

- 13 • We calculate the subglacial hydraulic potential for the Martian south polar ice cap from
14 measured surface topography and ice thickness.
- 15 • The recently-observed area of inferred basal melt does not occupy a predicted depression
16 in the subglacial hydraulic potential surface.
- 17 • We argue this supports the hypothesis that local geothermal heating could be responsible
18 for this area of liquid.
19

20 **Abstract**

21 The discovery of a ~20 km wide area of bright subsurface radar reflections, interpreted as liquid
22 water, beneath the Martian south polar layered deposits (SPLD) in data from the Mars Advanced
23 Radar for Subsurface and Ionosphere Sounding (MARSIS) instrument, and the discovery of two
24 geologically recent potential eskers (landforms produced by subglacial melt) associated with
25 viscous flow features in Martian mid-latitudes, has suggested recent basal melting of Martian ice
26 deposits may be feasible, possibly due to locally elevated geothermal heating. Locations of
27 terrestrial subglacial lakes and major drainage axes have been successfully predicted from
28 subglacial hydraulic potential surfaces calculated from surface topography and ice thickness.
29 Here, we use surface topography from the Mars Orbiter Laser Altimeter and SPLD bed
30 elevations derived from MARSIS data to calculate the subglacial hydraulic potential surface
31 beneath the SPLD and determine whether the observed high reflectance area coincides with
32 predicted subglacial lake locations. Given the sensitivity of terrestrial predictions of lake
33 locations to basal topography, we derive over 1000 perturbed topographies (using noise statistics
34 from the MARSIS data) to infer the most likely locations of possible subglacial water bodies and
35 drainage axes. Our results show that the high reflectance area does not coincide with any
36 substantial predicted lake locations; three nearby lake locations are robustly predicted however.
37 We interpret this result as suggesting that the high reflectance area (assuming the interpretation
38 as liquid is correct) is most likely a hydraulically-isolated patch of liquid confined by the
39 surrounding cold-based ice, rather than a topographically-constrained subglacial lake.

40

41 **Plain Language Summary**

42 Mars' present-day ice deposits are generally assumed to be frozen throughout given its cold
43 climate. However, new evidence from orbital radar data suggests a possible present-day ~20 km
44 wide area of liquid water beneath Mars' south polar ice cap. Recently-discovered landforms in
45 Mars' mid-latitudes have been interpreted as eskers (landforms produced by flowing meltwater
46 beneath glaciers on Earth) and also suggest that subglacial melt may be feasible in Mars' recent
47 past. Subglacial lakes are common on Earth, and their locations have been successfully predicted
48 from ice surface topography and ice thickness, in conjunction with theories for subglacial water
49 flow. In this paper we use the surface topography and ice thickness data for Mars' south polar ice
50 cap to calculate the theoretical locations of possible subglacial lakes, and compare these with the
51 location of the possible present-day area of liquid water. The observed patch of possible liquid
52 water does not coincide with the lake locations we predict. We interpret this result as implying
53 that the liquid water is most likely to be an isolated patch of liquid, possibly caused by locally-
54 raised geothermal heating, and which is fixed in position by the surrounding frozen ice, rather
55 than the liquid forming a topographically-constrained subglacial lake.

56 1 Introduction

57 Ice sheets, glaciers and ground ice distributed between Mars' poles and mid-latitudes
58 [e.g. *Plaut et al.*, 2007; *Levy et al.*, 2014; *Souness and Hubbard*, 2012; *Head et al.*, 2003] contain
59 a total volume of water ice (estimated as $\sim 3.5 \times 10^6 \text{ km}^3$), comparable to that of all glaciers on
60 Earth excluding the East Antarctic Ice Sheet [*Levy et al.*, 2014], Mars' existing ice deposits are
61 thought to have formed ~ 100 s to Myrs ago [e.g. *Head et al.*, 2003; *Arfstrom and Hartmann*,
62 2005; *Butcher et al.*, 2017; *Conway et al.*, 2018], with landforms and models also indicating
63 episodic glaciation of different regions of Mars' surface over billions of years [e.g. *Fastook et*
64 *al.*, 2008; *Wordsworth et al.*, 2013; *Butcher et al.*, 2016]. Glaciers and ice sheets insulate a
65 planet's surface, trapping geothermal heat from the planetary interior and frictional heat
66 produced by ice flow, typically making their beds warmer than their surfaces. However, it is
67 commonly thought that Mars' climate has been cold and dry for $\sim 2\text{--}3$ Gyr, and hence that the
68 existing glaciers and ice caps have most probably been frozen throughout [e.g. *Levy et al.*, 2016],
69 although there is some evidence for spatially-limited, ephemeral supraglacial melt [e.g. *Fassett et*
70 *al.*, 2010]. The migration of ice between different reservoirs at different locations on Mars over
71 this period is thought to have taken place via sublimation and (solid) precipitation, rather than
72 melt [*Bramson et al.* 2017].

73 In contrast, recent geomorphological studies found evidence for localized melting
74 beneath two existing mid-latitude glaciers around 110–150 Myr ago (Ma), in the form of
75 subglacially-deposited eskers in Phlegra Montes and Tempe Terra [*Gallagher and Balme*, 2015,
76 *Butcher et al.*, 2017]. Furthermore, that present day glaciers and ice caps are frozen throughout
77 has also been challenged by the discovery of possible present-day liquid water beneath Mars'
78 South Polar Layered Deposits [SPLD; *Orosei et al.*, 2018], based on anomalously bright
79 subsurface reflections recorded by the MARSIS (Mars Advanced Radar for Subsurface and
80 Ionospheric Sounding) instrument in a well-defined, 20 kilometer wide zone centered at 193°E,
81 81°S. In all three locations, elevated geothermal heating (e.g. due to localized magmatic
82 intrusions) has been proposed as a possible source of the heat needed to melt the basal ice under
83 cold current and recent Martian climate conditions [*Sori and Bramson*, 2019, *Butcher et al.*,
84 2017, *Gallagher and Balme*, 2015]. In addition, perchlorate salts, which substantially lower the
85 melting point of ice to around ~ 200 to 230K depending on the concentration and species of
86 perchlorate, may be highly enriched in the basal layers of the Martian polar ice caps [*Fisher et*
87 *al.*, 2010; *Sori and Bramson*, 2019].

88 Subglacial water is far more common on Earth, occurring beneath the bulk of terrestrial
89 valley glaciers, and both the Greenland and Antarctic Ice Sheets. In the case of Antarctica, large
90 numbers of subglacial lakes, including Lake Vostok, the 6th largest (by volume) lake on Earth,
91 have been detected by observing bright reflectances from numerous ground penetrating radar
92 systems. Movement of water between subglacial lakes in large episodic drainage events has also
93 been observed using localized changes in surface topography as lakes drain and fill. There is also
94 a great deal of evidence for widespread subglacial meltwater in the landscapes occupied by ice
95 during the Quaternary (and earlier) glaciations on Earth [e.g. *Shreve*, 1985; *Storrar et al.*, 2013,
96 *Stroeven et al.*, 2016; *Clark et al.*, 2017].

97 The physics of subglacial water flow are well known. *Flowers* [2015] provides a
98 comprehensive review, but briefly, the overall pattern of subglacial water flow is governed by
99 the subglacial potential surface (calculated from the subglacial topography and ice thickness
100 distribution; *Shreve*, [1972]). Local depressions in the potential surface form the nuclei for

101 possible subglacial lakes, as water cannot escape from such depressions until it fills to the
102 elevation of the local ‘spill point’, the elevation of the lowest point in the constraining potential
103 surface. Calculations of subglacial potential using digital elevation models (DEMs) of ice
104 surfaces and ice thickness (or bed elevations) have proved very successful at predicting
105 subglacial lake locations and volumes for terrestrial ice sheets. In a study of Antarctic lake
106 locations, *Willis et al.*, [2016] show that the centers of lake locations predicted using the
107 calculated subglacial potential surface were a mean distance of 6.3 km from the centers of the
108 379 known (in 2016) lakes. *Willis et al.*, [2016] predict around 100 times as many possible lake
109 locations as the number of known lakes however. Some of this mis-match will be due to the size
110 and remoteness of Antarctica, meaning that the current lake inventory will be incomplete.
111 However, some of the mis-match will be due to errors of commission, where the potential
112 surface predicts a lake where no lake exists. Such errors may be caused by inaccuracies in the
113 surface or bed DEMs due to inaccuracies and/or resolution effects in the sensors, or the
114 interpolation of relatively sparse point data onto a regular grid.

115 Several studies have shown that quite small changes in the bed and particularly the
116 surface elevation can cause substantial changes in predicted subglacial water flow directions and
117 lake locations. *Wright et al.*, [2008] found that raising three grid cells in the surface DEM by
118 10 m led to water flow switching between two major outlets from East Antarctica, and
119 subsequently lowering another four surface DEM cells by 10 m led to another switch between
120 two other large outlet glaciers. *Le Brocq et al.*, [2009] report similar sensitivity to small (~10 m)
121 changes in surface elevation that divert flow between two major ice streams draining the Siple
122 Coast of West Antarctica.

123 Uncertainties in data will also affect the predicted location of subglacial lakes.
124 *Livingstone et al.*, [2013a] used current bed topography and modeled ice thickness distributions
125 and isostatic effects from a suite of ice-dynamic reconstructions of the geometry of the North
126 American Ice Sheets during the last glaciation on Earth to predict possible subglacial palaeo-lake
127 locations. They found that lake locations (and inferred water flow directions) were highly
128 sensitive to the configuration of the ice sheet at particular stages during ice sheet growth and
129 decay, and depended in some instances on the choice of model and the predicted ice thickness
130 distribution. They found that some predicted lakes were persistent through time, and were also
131 predicted by many of the modeled ice sheet configurations, whereas other predicted lakes were
132 present in fewer simulations, or were more transient features through time. Overall, deeper
133 predicted lakes in areas of more pronounced topography (e.g. beneath the Cordilleran Ice Sheet
134 in western North America) were more persistent. In another study, *Livingstone et al.*, [2013b]
135 investigated the impact of errors or uncertainty in the basal topography of Antarctica on
136 predicted subglacial lake locations by perturbing the bed topography using random noise with a
137 standard deviation equal to the uncertainty in the bed elevation. They performed 50 such
138 perturbations to produce a ‘persistence map’ of lakes, in which the most persistent lakes were
139 those which were predicted in the majority of the perturbed-bed experiments.

140 In this paper we use gridded surface and ice thickness maps for the Martian South Polar
141 Layered Deposit [SPLD; *Plaut et al.*, 2007] to calculate the subglacial hydraulic potential. From
142 this, we infer the locations of depressions in the potential surface that would be expected to
143 confine possible subglacial water bodies were basal meltwater available beneath the SPLD. We
144 also predict the main drainage axes that any possible flowing water would be expected to follow.
145 In order to address uncertainties in the bed elevation data (due to the limited resolution of the ice

146 thickness data from the MARSIS sensor, in particular) we adopt a similar approach to
 147 *Livingstone et al.*, [2013b] and perform a suite of lake location calculations using DEMs derived
 148 from the raw MARSIS data, the mean and median of all radar footprints crossing each point (cf.
 149 *Orosei et al.*, [2018]), and a set of 1000 DEMs interpolated from randomly perturbed bed
 150 elevation data, with perturbations calculated using the statistical properties of the elevation
 151 differences within the MARSIS data. We use different assumptions of the density of the inferred
 152 subglacial liquid, from pure water to saturated perchlorate brine, and we also investigate the
 153 impact of the spatial resolution of the interpolated DEMs. From this set of calculations we
 154 determine the most persistent predicted lake locations and predicted water flow paths. We then
 155 consider the implications of these predictions by comparing the predicted lake locations and
 156 persistence values with the high reflectance area (HRA) reported by *Orosei et al.*, [2018]. We
 157 use these predictions to consider the likelihood (or otherwise) of a well-connected subglacial
 158 drainage system beneath the Martian SPLD, or whether the possible water body inferred from
 159 radar reflectance [*Orosei et al.*, 2018] is more likely to be a hydraulically-isolated feature.

160 2 Methods

161 2.1 Modeling lake locations

162 We calculate subglacial hydraulic potential (ϕ) from the bed and surface elevation
 163 [*Shreve*, 1972]:

$$164 \quad \phi = \rho_w g Z + k \rho_i g H \quad [1]$$

165 where g is gravity (3.711 m s^{-2}); ρ_w is the density of the subglacial liquid, ranging from
 166 1000 kg m^{-3} for pure water to 1980 kg m^{-3} for saturated perchlorate brine [*Fisher et al.*, 2010]; Z
 167 is the bed elevation (m), ρ_i is the ice density (here taken to be 910 kg m^{-3} , given the uncertainty
 168 concerning the overall density of the Martian SPLD; [e.g. *Zuber et al.*, 2007; *Plaut et al.*, 2007;
 169 *Wieczorek*, 2008]) and H is the ice thickness (m). k is a dimensionless factor which represents the
 170 influence of ice overburden pressure on the local subglacial water pressure, with $k = 1$ implying
 171 the water pressure is at the ice overburden pressure, and $k = 0$ implying the subglacial water is at
 172 atmospheric pressure. In terrestrial systems, k is variable in time and space, especially in
 173 situations closer to the ice margin where the ice is thinner and/or heavily affected by seasonal
 174 meltwater. In the interiors of ice sheets, where the ice is thicker, and especially for Antarctica,
 175 measured water pressures are typically very close to ice overburden. Given the location of the
 176 HRA in the interior of the SPLD, and the $\sim 1500 \text{ m}$ inferred ice thickness, we assume $k = 1$ in our
 177 simulations.

178 Equation 1 can usefully be re-cast into an alternative formulation which uses Z as before,
 179 and also the ice surface elevation, S , where $S = Z + H$:

$$180 \quad \phi = (\rho_w - k \rho_i) g Z + k \rho_i g S \quad [2]$$

181 This alternative formulation highlights the typical dominance (for high k values) of the
 182 ice surface elevation on the subglacial hydraulic potential due to the small difference in the
 183 density of ice and water. For the SPLD, however, if the liquid layer is a high solute-
 184 concentration brine, ρ_w will be higher, increasing the importance of the basal topography in

185 determining the subglacial hydraulic potential. Given this, we use $\rho_w = 1980 \text{ kg m}^{-3}$ in the bulk
186 of our calculations, but we also perform a set of runs in which ρ_w is varied in 11 equal steps
187 between 1000 and 1980 kg m^{-3} . Uncertainty in the ice density [e.g. *Zuber et al.*, 2007; *Plaut et*
188 *al.*, 2007; *Wieczorek*, 2008] will have an opposing effect. A higher ice density will increase the
189 dominance of the ice surface elevation, reducing the role played by basal topography on the
190 subglacial potential, and reducing the impact of possible higher liquid density. If the ice density
191 exceeds the liquid density, this will reverse the sign of the first term in Eq. 2 (although this
192 negative value will be compensated to some extent by the increased value of the second term,
193 depending on the relative values of the ice and liquid densities, and the surface and bed
194 elevations). To assess the likely maximum impact of this effect, we also perform a run using an
195 ice density of 1200 kg m^{-3} [*Zuber et al.*, 2007; *Wieczorek*, 2008] and a liquid density of 1000 kg
196 m^{-3} .

197 We calculate lake locations using the flow accumulation algorithm developed by *Arnold*
198 [2010], as applied by *Willis et al.*, [2016] for Antarctica, using gridded ϕ values calculated from
199 the surface topography, inferred bed topography, and ρ_w value for each experiment. The
200 calculated hydraulic potential (hereafter shortened to potential) gradient allows inferred water
201 flow directions to be calculated for each DEM cell; the algorithm assigns each DEM cell an area,
202 based on its own area plus the total area upstream of the cell for which water flow-lines pass
203 through the cell. The algorithm also identifies all cells in the potential surface that are at a lower
204 potential than all their neighbours (and which therefore act as ‘dead end’ in the flow
205 accumulation algorithm) and defines them as ‘sink’ cells. Together, the water flow directions and
206 sink cells allow local subglacial catchments (a group of contiguous cells which all drain toward
207 the sink) to be determined. Sink cells also form the nucleus for possible subglacial lakes; the
208 algorithm ‘floods’ each sink cell to find the elevation of the lowest cell in the catchment
209 surrounding the sink cell over which water would spill into a lower potential downstream cell
210 (and hence, into an adjacent catchment). This spill point cell defines the maximum depth
211 (relative to the elevation of the sink cell), area, and volume of each predicted possible lake, and
212 also allows the routing algorithm to pass the total catchment area from catchment to catchment
213 downstream until the model reaches the edge of the ice cap. In this way, the algorithm builds up
214 the topology of possible subglacial water flow, linking the individual catchments and lakes
215 together into arborescent structures analogous to typical stream networks. Major drainage axes
216 appear as distinct ‘threads’ across the potential surface with large upstream area values, meaning
217 a large number of upstream cells ultimately feed their area through such cells. Upstream source
218 areas, or more isolated areas, show much lower upstream area values. Cells within a lake are
219 assigned an area equal to the total upstream area above the spill point for that lake. Lakes on the
220 main drainage axes appear therefore as ‘beads’ of high upstream area, but even lakes in upstream
221 or more isolated areas are easily identified as areas of uniform, locally-high upstream area
222 values.

223 2.2 Ice surface and bed topography

224 For the ice surface elevation, we use the south polar MOLA gridded topographic map for
225 Mars at 128 pixels per degree (460 m per pixel) resolution (Figure 1a; *Smith et al.*, [2001]) in all
226 experiments. For the bed topography, we use the MARSIS sub-spacecraft latitude and longitude,
227 and the inferred elevation of the basal reflector (used to produce the ice thickness and hence bed
228 elevation datasets from *Plaut et al.*, [2007]), supplemented with the additional data reported in

229 *Orosei et al.*, [2018], to provide the x,y,z point-cloud data needed for interpolation of the bed
230 DEMs. We perform our experiments for the 200×200 km area centered around 193°E , 81°S that
231 contains the high-reflectance signal reported by *Orosei et al.*, [2018]; Figure 1b. In order to
232 facilitate direct comparison with the results reported in *Orosei et al.* 2018, we also adopt a north-
233 down orientation in Figure 1b and subsequent figures. 86 points from *Plaut et al.* [2007] fall in
234 our study area, and allow interpolation outside the area covered by the point cloud data presented
235 in *Orosei et al.* [2018]. In order to compare our results directly with those from *Orosei et al.*,
236 [2018], we perform the bulk of our experiments at the 200 m resolution used by *Orosei et al.*,
237 [2018], but we also derive DEMs at 500 m, 1000 m, 2500 m and 5000 m resolution to test the
238 impact of DEM resolution. Following *Plaut et al.*, [2007], we use natural neighbour
239 interpolation, and apply noise to the MARSIS data from *Orosei et al.* [2018] as detailed below
240 for our perturbation analysis. We pin the edge of the study area to the elevation taken from the
241 overall SPLD gridded bed elevation map [*Plaut et al.*, 2007].

242 The vertical resolution of the MARSIS sensor is around 100–150 m [e.g. *Plaut et al.*,
243 2007], controlled ultimately by the frequency and time resolution of the sensor. Inferring bed
244 elevation from the radar returns also requires knowledge of the dielectric constant of the
245 material; we use the values (and hence elevations) reported by *Orosei et al.* [2018]. The noise we
246 apply to the data (detailed below) effectively allows us to simulate the impact of uncertainty in
247 the dielectric constant however. The Fresnel radius (ground footprint) is $\sim 3\text{--}5$ km [*Orosei et al.*,
248 2018] and varies along-track and across-track. 76,800 individual data points are available within
249 the study area, from 27 individual satellite tracks (Figure 2a). As can be seen in Figure 2a,
250 however, the inferred SPLD thickness along the tracks varies considerably over quite short
251 distances (well within the sensor footprint), which suggests that possible local variations in the
252 dielectric, sensor resolution effects, or other noise sources strongly affect the inferred elevation.
253 The along-track spacing of the points varies from ~ 30 to ~ 90 m, meaning that there is
254 considerable footprint overlap along the tracks, and several sets of tracks are also close together,
255 running nearly parallel to each other, or cross. This overlap between the footprints of the
256 individual measurement points gives us information on the local variability or uncertainty in the
257 inferred SPLD thickness in the MARSIS data. In order to estimate the statistical properties of
258 this uncertainty, and provide a measure of the noise within the data, we identify all the points
259 within the Fresnel radius of the sensor for each point, and use these to calculate the mean (Figure
260 2b) and median (Figure 2c) height differences, and the standard deviation (Figure 2d) of the
261 height differences, between each original point and the other points within the Fresnel radius.

262 Figure 3a shows the overall distribution of in-radius elevation differences, with the
263 calculated mean and standard deviation of height difference. It approximates a normal
264 distribution, though with a more marked central peak. However, some individual points deviate
265 quite markedly from their neighbours; Figure 3b-d shows the distribution for the points with the
266 maximum (Figure 3b) and minimum (Figure 3c) mean in-radius elevation difference, and the
267 point with the largest standard deviation (Figure 3d) of in-radius elevation difference. Of the
268 76,800 points, 64,496 have a standard deviation of the in-radius elevation difference smaller than
269 ± 1 standard deviation for the total set. The number of in-radius points varies between 108 and
270 2362, with a mean value of 665 in-radius points per point.

271 We create a set of bed DEMs (shown in Figure 4 at 200 m resolution) for the hydraulic
272 potential and flow routing calculations using the raw MARSIS points from *Orosei et al.*, [2018],

273 (Figure 4a); and the mean (Figure 4b), median (Figure 4c), minimum, and maximum of the
274 elevation differences within the Fresnel radius as above. The minimum and maximum
275 topographies were judged to be implausible due to the very large height differences produced in
276 some areas, and are not used further. We then use a form of Monte-Carlo analysis to create a set
277 of 1000 perturbed bed topographies by applying a random elevation change to each data point
278 drawn from a normal distribution with the calculated mean and standard deviation of the in-
279 radius elevation differences for that point. We apply the flow accumulation algorithm to each
280 perturbed DEM, which then allows us to calculate a probability value that any given pixel is
281 within a lake for the set of 1000 model runs. We also create a DEM (dubbed the mean-perturbed
282 DEM) using the mean elevation of each cell calculated from the set of individual perturbed
283 topographies (Figure 4d). We also apply the flow accumulation algorithm to the same set of
284 DEMs at reduced resolution, but focus mainly on the 200 m resolution results for direct
285 comparison with *Orosei et al.*, [2018].

286 Figure 4 shows that the mean, median and mean-perturbed 200 m resolution topographies
287 are considerably smoother than that produced by the raw data. For all four, however, the impact
288 of some of the individual satellite tracks can still be seen, such as the straight, deep trough
289 running NNW from the S (top) edge of the area. This effect is reduced most in the mean-
290 perturbed topography, which we therefore use in the series of experiments to investigate the
291 possible impact of variable ρ_w values.

292 **3 Results**

293 Figure 5 shows the common logarithm of the flow accumulation results for the four 200
294 m resolution topographies shown in Figure 4, using ρ_w of 1980 kg m^{-3} .

295 Possible lake locations are seen as the broad patches with high, uniform flow
296 accumulation values (yellows in Figure 5). The overall drainage trend across the majority of the
297 area is towards the east-north-east (left/bottom left) in all four simulations, driven by the effect of
298 the overall slope of the SPLD surface topography in the area (Figure 1b) on the subglacial
299 hydraulic potential (Eq. 2). The effect of the surface scarps in the north-west (bottom right) of
300 the area (Figure 1b) can also be clearly seen as the thinner ice downstream of the scarps leads to
301 local low subglacial potential values which act as large possible lake locations. Away from the
302 scarps, however, where the SPLD surface topography is much flatter, the basal topography acts
303 as the major control on predicted possible lake locations. Depressions in the bed form the foci for
304 numerous possible lakes, especially in the rougher raw topography DEM (Figure 5a). However,
305 what is clear is that the area of high basal reflectance does not coincide with a substantial
306 predicted possible lake location in any reconstruction.

307 Figure 6 shows the 2500 km^2 central area around the HRA in more detail. Consistently,
308 three substantial possible lake locations are predicted nearly adjacent to the HRA; an irregularly
309 shaped lake to the south-west (upper right) of the HRA; a rounder lake to the east (left), and a
310 lake with a more variable extent to the south-east (upper left). For the raw topography DEM,
311 several smaller lake locations are predicted within the HRA. These have lower flow
312 accumulation values than the three larger adjacent lakes, however, and are topologically more
313 isolated. In topological terms, the SW lake is linked to the E lake via a drainage axis running
314 through the HRA from W to E in the mean and mean perturbed topographies (Figures 6b and d).

315 For the raw topography DEM (Figure 6a), the SW lake flows into the SE lake, and then to the E
316 lake, with the predicted drainage axis touching the southern tip of the HRA. For the median
317 topography DEM (Figure 6c), the routing is different again. The SW lake extends beyond the
318 southern (top) edge of the study area, and flow is directed into the southern edge of the SE lake
319 (outside the area shown in the figure), and then to the E lake; no substantial drainage axis crosses
320 the HRA.

321 Figure 7a shows the probability values (P) for whether any given pixel is part of a
322 possible lake from the set of 1000 perturbed topography runs. P is calculated by dividing the
323 number of runs in which a pixel is calculated to be within a lake by the total number of runs.
324 Thus, $P = 0$ for a pixel which was never calculated to be in a lake and $P = 1$ for a pixel which
325 was calculated to be in a lake in all simulations. Here, the overall position of the three large
326 adjacent predicted possible lakes is robust to perturbation using the calculated elevation variation
327 statistics, although their exact extents vary (with the edges of the lakes having lower pixel
328 probabilities), particularly for the southern extent of the SW lake. A small but persistent lake also
329 appears in the southern part of the HRA that is less apparent in the single topography
330 calculations (Figures 5 and 6). The individual satellite tracks are visible, as small predicted
331 possible lakes will form in some topographies when an individual point is randomly moved
332 downwards by a large amount, producing a deep, small depression, or when nearby points are
333 moved upwards producing a depression between them. Figure 7b shows the mean flow
334 accumulation values for the 1000 perturbed-topography runs. This again shows the persistence of
335 the main predicted possible lakes; the lake edges are less sharp than for the single-topography
336 runs, as pixels nearer the margins which are only calculated to be in lakes in some perturbed
337 topographies will have lower mean flow accumulation values. There is also some blurring of the
338 main drainage axes as calculated routing between the lakes will vary between the different
339 perturbed topographies, but again, a drainage axis running W-E across the southern part of the
340 HRA is visible, as occurs in the mean and mean-perturbed topographies (Figures 5 and 6b and
341 d).

342 Figures 7c–f show the results for the variable liquid and ice density experiments. Figure
343 7c shows the probability that any given pixel is part of a lake and Figure 7d shows the mean flow
344 accumulation values for the 11 variable liquid density runs. Figure 7e shows the results for the $r_w = 1000 \text{ kg m}^{-3}$ run, and Figure 7f the results for the $r_i = 1200 \text{ kg m}^{-3}$ and $r_w = 1000 \text{ kg m}^{-3}$ run. As
345 implied by Eq. 2, for lower-density liquid the impact of bed topography is heavily reduced, and
346 the surface slope dominates. For $r_w = 1000 \text{ kg m}^{-3}$, calculated lake occurrence in the vicinity of
347 the HRA is virtually eliminated (Figure 7e). As liquid density increases, the SE lake begins to
348 form at the lowest liquid density, and increases in extent most rapidly, as shown by the largest
349 area of high probability (Figure 7c) and the largest area of high (bright yellow) mean flow
350 accumulation values (Figure 7d). The SW and E lakes only become more extensive for inferred
351 density $> \sim 1600 \text{ kg m}^{-3}$, with smaller areas of high probability (Figure 7c) and high mean flow
352 accumulation values (Figure 7d). The drainage axis from the SW to E lake through the HRA is
353 present in all variable-density runs (Figure 7d). For the run with $r_i = 1200 \text{ kg m}^{-3}$ and $r_w = 1000$
354 kg m^{-3} (Figure 7f), the overall pattern of flow is similar to the $r_w = 1000 \text{ kg m}^{-3}$ run (Figure 7e)
355 particularly within the HRA, which shows no predicted lake occurrence. The small predicted
356 lake upstream of the location of the SE lake is caused by the increased impact of the thicker ice
357 over the deep bed depression in this area preventing water from being routed through the
358 depression, forcing water to flow around the bed depression.
359

360 Figure 8 shows the mean flow accumulation values from the set of 1000 reduced
361 resolution perturbed-topography runs (the equivalent to the 200 m resolution results shown in
362 Figure 7b). These clearly show that resolution does not affect the location of the predicted
363 possible lakes, nor the route of the drainage axis from the SW to E lake through the HRA. Only
364 at the coarsest resolution (5000 m, Figure 8d) does the pattern start to break down somewhat,
365 with the SW and E lakes remaining as separate entities, but linked by a much broader drainage
366 axis through the HRA.

367 4 Discussion

368 Mapping the locations of depressions in subglacial hydraulic potential surfaces has
369 proved very successful in terms of predicting the location of subglacial lakes and drainage axes
370 on Earth. The robust results reported here, which show that the HRA beneath the SPLD does not
371 occupy such a depression, but occurs within 10 to 20 km of three such possible depressions
372 which do not show high radar reflectances [Orosei et al. 2018], therefore seems anomalous.
373 However, there are critical differences between the vast majority of terrestrial subglacial lakes
374 and a hypothesized lake (or area of basal liquid) beneath the SPLD. Almost all known subglacial
375 lakes on Earth are fed by meltwater which flows into the lake basin from upstream, delivered by
376 an active subglacial drainage system that covers the ice sheet bed, which is at the pressure
377 melting point, over wide areas. For Antarctica, where large numbers of subglacial lakes are
378 known to exist, the main water source is basal melt. Numerical modeling suggests mean
379 subglacial melt rates around 2–5 mm yr⁻¹ beneath Antarctica, with some fast flowing areas
380 reaching several hundred mm per year [e.g. Willis et al., 2016; Pattyn 2010, Llubes et al., 2006].
381 Basal melt is generated by a combination of strain heating and geothermal heat flux; West
382 Antarctica in particular exhibits some of the highest melt rates, as it is thought to have relatively
383 high geothermal heat flux of over 120 mW m⁻² in places (versus a continent-wide average value
384 of 30-60 mW m⁻²) [Martos et al., 2017], and contains areas of very fast-moving ice which
385 generate large amounts of strain heating. These high basal melt rates lead to sometimes very
386 substantial water fluxes [over 1 km³ yr⁻¹, Willis et al., 2016] into some active lakes near the
387 margins of Antarctica, which can trigger cyclic drainage events which have been observed
388 through their impact on the surface topography of the ice sheet [e.g. Wingham et al., 2006,
389 Fricker et al., 2007]. Such conditions are therefore very different from those at the base of the
390 SPLD.

391 As well as subglacial lakes with large catchments fed by near-ubiquitous basal melt (such
392 as beneath Antarctica), subglacial lakes directly fed by localized high geothermal heat fluxes are
393 known to exist on Earth. The most widely studied is Grimsvötn, a subglacial lake beneath
394 Vatnajökull, a large ice cap in SW Iceland. This lake is known to fill due to subglacial melt
395 driven by very high geothermal heating [estimated at 35–40 W m⁻², Björnsson and
396 Guðmundsson, 1993], and then drain catastrophically when the lake volume exceeds a threshold.
397 Subglacial hydrological theory has proved rather successful at predicting the discharge
398 hydrographs of these drainage events [e.g. Nye, 1976; Spring and Hutter, 1981, Fowler, 1999],
399 and even (though with less certainty) the triggers of the drainage events [Fowler, 1999]. The key
400 finding from Grimsvötn is that the lake forms in a topographic depression beneath the ice, which
401 causes a low in the subglacial hydraulic potential. The lake surface, however, can reach a higher
402 elevation than the top of the topographic lip of the basin as the subglacial potential increases in
403 front of the lake due to thicker ice downstream. This reversed potential slope impounds the lake,

404 not the bed topography in itself. At some critical water level, however, the potential gradient
405 reverses and the lake drains catastrophically, as the large water discharge enables rapid growth of
406 an efficient subglacial drainage system. These drainage events result in catastrophic floods
407 (jökulhlaups) beyond the ice cap margin, and result in the formation of pronounced cauldrons on
408 the ice cap surface.

409 Figure 9 shows the surface and basal topography, and subglacial hydraulic potential, for a
410 transect through the HRA located on the drainage axis between the SW and E depressions
411 beneath the SPLD (shown in Figure 7b). The reverse gradient in subglacial potential that
412 impounds the two predicted lakes can be clearly seen, but no such reverse gradient exists for the
413 region occupied by the HRA.

414 In addition to local melt, Grimsvötn is thought to also receive melt from upstream [e.g.
415 *Fowler, 1999*]; the ice cap is also at the pressure melting point, so once discharge from the lake
416 is underway, no thermal barrier exists to prevent drainage. This also contrasts strongly with the
417 situation beneath the SPLD. If locally-raised geothermal heating is necessary to raise the basal
418 ice to the melting point [*Sori and Bramson, 2019*], the rate of filling of any putative lake would
419 be far slower. This is because of the lower likely geothermal heating, more rapid loss of heat
420 through the much colder ice within the SPLD, and the lack of any upstream catchment to feed
421 additional melt into the hypothesized lake. The inferred liquid beneath the SPLD therefore seems
422 to be trapped by a thermal ‘dam’, rather than by variations in the potential surface. This perhaps
423 makes it less likely that the HRA is a true subglacial lake, in the sense of being constrained by
424 topography.

425 The geophysical model results presented by *Sori and Bramson* [2019] show geothermal
426 heating rising rapidly (over ~0.5 Myr) after the intrusion of a magma chamber beneath the HRA,
427 and then falling more slowly (over the subsequent 1–2 Myr). If the HRA is indeed geothermally-
428 heated liquid, it could therefore currently be on the ‘rising limb’ of heating, and getting larger; at
429 its peak extent (though perhaps this is the least likely scenario due to the shorter duration of peak
430 heating); or on the ‘falling limb’, and shrinking. If the former is the case, and the extent of basal
431 melt increases in the future such that the area expands to encompass our predicted depressions in
432 the potential surface, then we can speculate whether such an occurrence would allow a ‘true’
433 subglacial lake, pinned by the topography, to form. We can also speculate if a rapid drainage
434 event of this lake would occur if it reached some critical volume threshold. In the latter case, and
435 especially if a drainage event had occurred in the past, it could be the case that some topographic
436 signature on the surface of the SPLD might be manifest. Whilst Icelandic subglacial volcanism
437 can produce cauldrons several hundred meters deep for ice typically up to ~500 m thick,
438 numerical modeling [e.g. *Evatt and Fowler, 2007*] and observations [e.g. *Wingham et al., 2006,*
439 *Fricker et al., 2007, 2009*] suggest changes in surface topography of a few metres for lakes of
440 similar size to the HRA for the thicker ice, and pressure-driven drainage events, known to occur
441 for Antarctic subglacial lakes. Such topographic signatures would be difficult to detect on the
442 SPLD, but could perhaps be detected in derivatives of the highest resolution MOLA elevation
443 data. Currently no such anomalous topography can be discerned on the ice surface over the
444 HRA.

445 Our results do not suggest that the explanation of the HRA being caused by the presence
446 of liquid water [*Orosei et al., 2018*] is any more or less likely due to it not being in a depression

447 within the calculated subglacial potential surface. Our results also suggest that spatially-limited
448 geothermal heating seems unlikely to produce an active subglacial drainage system capable of
449 sustaining flow across parts of the bed, but rather that spatially-limited heating may produce melt
450 which is pinned *in-situ* by the surrounding frozen base. This suggests that more extensive basal
451 melting, driven by a combination of a larger area of higher geothermal heating, thicker ice and
452 warmer surface temperatures (all of which were more probable earlier in Mars' history, and in
453 agreement with previous palaeo-environmental reconstructions [e.g. *Fastook et al.*, 2012;
454 *Scanlon et al.*, 2018]), would be required for the occurrence of an active subglacial drainage
455 system capable of moving large quantities of sediment, and therefore of forming the ancient
456 [\sim 3.5–3.6 Ga; *Kress and Head* 2015; *Bernhardt et al.*, 2013] putative eskers in the south polar
457 Dorsa Argentea Formation and Argyre Planitia [e.g., *Butcher et al.*, 2016; *Bernhardt et al.*,
458 2013]. In the case of the younger (110–150 Ma) esker systems observed in Mars' mid-latitudes
459 [*Gallagher and Balme*, 2015, *Butcher et al.*, 2017], however, widespread subglacial melt seems
460 less likely given the colder Martian climate in the Amazonian. Mid-latitude glaciers are thought
461 to have been more extensive during periods of high obliquity, however [e.g. *Baker and Head*
462 2015], which would tend to warm the basal ice directly, and could lead to higher strain heating
463 due to larger basal shear stress [*Butcher et al.*, 2017]. Coupled with locally elevated geothermal
464 heat, this would make it more likely for the basal ice to reach the melting point in the area above
465 the geothermal anomaly, especially if salts lowered the melting point of the basal ice. Were the
466 melting to occur over a large enough area, this liquid might begin to flow down the potential
467 surface or become impounded in depressions in the subglacial hydraulic potential surface as
468 topographically-constrained subglacial lakes. The number and extent of such depressions in the
469 potential surface would be increased if the basal liquid consisted of high-density brine. The
470 possible growth of any such lakes during the onset-to-peak phase of heating could lead to them
471 filling the depressions in the potential surface, possibly reaching the threshold size needed for
472 catastrophic drainage to occur. This would seem likely to mobilise large quantities of sediment,
473 leading to possible esker formation in a transient subglacial drainage system. Such a mechanism
474 could help explain the occurrence, size, and also the rarity, of modern mid-latitude Martian
475 eskers.

476 **5 Conclusions**

477 Our results show that the HRA reported by *Orosei et al.*, [2108] beneath the Martian
478 SPLD does not occupy a depression in the subglacial hydraulic potential surface as calculated
479 using subglacial hydraulic theory and the surface and basal topographies. This finding is robust
480 to random perturbations of the bed DEM using the mean and standard deviation of elevation
481 difference (calculated from the set of points within the Fresnel radius for each point), to different
482 interpolation resolutions, and to different assumed densities for the inferred liquid layer. For the
483 latter case, a higher assumed density increases the effect of the bed topography on the subglacial
484 potential surface, as expected from subglacial hydraulic potential theory (Eq. 2, *Shreve*, [1972]),
485 and makes predicted lake locations both larger in extent, and more numerous. Three substantial
486 depressions in the potential surface, with areas between one-third and three-quarters of the area
487 of HRA, are predicted to occur within \sim 20 km of the HRA, and a major drainage axis in the
488 potential surface crosses the HRA in over 90% of the perturbed topographies, a topography
489 calculated using the mean in-radius elevation, and a topography calculated from the mean
490 elevation of the set of 1000 perturbed topographies. The areas occupied by these depressions do

491 not show high radar reflectance in the MARSIS data, and therefore show no sign of liquid water
492 [*Orosei et al. 2018*].

493 The methodology we develop here to acknowledge the uncertainty in radar-derived bed
494 elevation data adds confidence to our findings. Flow accumulation calculations require the
495 gradient in the bed elevation to be calculated, and as such are very sensitive to noise in the point-
496 cloud data used to produce interpolated DEMs. Assessing the local variation in the inferred
497 topography using the inherent spatial overlap between the individual sensor measurements
498 (given the large footprint on the MARSIS instrument relative to the measurement frequency)
499 allows us to calculate the statistical properties of the bed elevation estimates around each original
500 point, and hence of the uncertainties in the estimated bed elevation. These statistics then allow
501 realistic random noise values to be added to the point-cloud data in a form of Monte-Carlo
502 analysis, which allows the calculation of both the mean elevation of pixels in the individual
503 perturbed topographies, and the mean values from any analysis using the individual perturbed
504 topographies. For analyses that require the gradient of the topography to be calculated, this
505 procedure adds valuable robustness to the calculated values.

506 We suggest that if the HRA is due to the presence of liquid water, or at least water-
507 saturated basal sediments, the location of the HRA seems likely to be due to locally-forced in-
508 situ basal melting, rather than the melt occurring elsewhere and the resulting liquid then
509 collecting in an area of low hydraulic potential as is typical for terrestrial subglacial lakes. This
510 supports the argument made by *Sori and Bramson* [2019] that localized geothermal heating is the
511 heat source required for melt to occur. Given that the subglacial potential surface in the HRA
512 does not show any substantial depressions, but instead seems to include a major predicted
513 drainage axis, the area of inferred basal melt seems to be pinned in-situ, presumably by the
514 surrounding cold-based ice which acts as an effective aquiclude and prevents movement of the
515 liquid down the subglacial potential surface. We suggest that the HRA does not represent a ‘true’
516 subglacial lake, pinned by topography, but that if it does represent liquid water, it is more likely
517 to be a brine-enriched sludge, or to consist of shallow (but spatially extensive) brine pools as
518 suggested by *Orosei et al.* [2018].

519

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531 repository maintained by *Orosei et al.* [2018] at <http://doi.org/10.5281/zenodo.1285179>. MOLA
532 data are available from the PDS Geosciences node at: [http://pds-](http://pds-geosciences.wustl.edu/missions/mgs/megdr.html)
533 [geosciences.wustl.edu/missions/mgs/megdr.html](http://pds-geosciences.wustl.edu/missions/mgs/megdr.html).

535 **References**

- 536 Arfstrom, J., & Hartmann, W. K. (2005). Martian flow features, moraine-like ridges, and gullies:
 537 Terrestrial analogs and interrelationships. *Icarus*, *174*(2), 321–335.
 538 <https://doi.org/10.1016/j.icarus.2004.05.026>
- 539 Arnold, N. (2010). A new approach for dealing with depressions in digital elevation models
 540 when calculating flow accumulation values. *Progress in Physical Geography: Earth and*
 541 *Environment*, *34*(6), 781–809. <https://doi.org/10.1177/0309133310384542>
- 542 Baker, D. M. H., & Head, J. W. (2015). Extensive Middle Amazonian mantling of debris aprons
 543 and plains in Deuteronilus Mensae, Mars: Implications for the record of mid-latitude
 544 glaciation. *Icarus*, *260*, 269–288. <https://doi.org/10.1016/j.icarus.2015.06.036>
- 545 Bernhardt, H., Hiesinger, H., Reiss, D., Ivanov, M., & Erkeling, G. (2013). Putative eskers and
 546 new insights into glacio-fluvial depositional settings in southern Argyre Planitia, Mars.
 547 *Planetary and Space Science*, *85*, 261–278. <https://doi.org/10.1016/j.pss.2013.06.022>
- 548 Björnsson, H., & Guðmundsson, M. T. (1993). Variations in the thermal output of the subglacial
 549 Grímsvötn Caldera, Iceland. *Geophysical Research Letters*, *20*(19), 2127–2130.
 550 <https://doi.org/10.1029/93GL01887>
- 551 Bramson, A. M., Byrne, S., & Bapst, J. (2017). Preservation of Midlatitude Ice Sheets on Mars:
 552 Mars Midlatitude Ice Sheet Preservation. *Journal of Geophysical Research: Planets*,
 553 *122*(11), 2250–2266. <https://doi.org/10.1002/2017JE005357>
- 554 Butcher, F. E. G., Conway, S. J., & Arnold, N. S. (2016). Are the Dorsa Argentea on Mars
 555 eskers? *Icarus*, *275*, 65–84. <https://doi.org/10.1016/j.icarus.2016.03.028>
- 556 Butcher, F. E. G., Balme, M. R., Gallagher, C., Arnold, N. S., Conway, S. J., Hagermann, A., &
 557 Lewis, S. R. (2017). Recent basal melting of a mid-latitude glacier on Mars. *Journal of*
 558 *Geophysical Research: Planets*, *122*(12), 2445–2468.
 559 <https://doi.org/10.1002/2017JE005434>
- 560 Clark, C. D., Ely, J. C., Greenwood, S. L., Hughes, A. L. C., Meehan, R., Barr, I. D., et al.
 561 (2018). BRITICE Glacial Map, version 2: a map and GIS database of glacial landforms
 562 of the last British-Irish Ice Sheet. *Boreas*, *47*(1), 11–e8. <https://doi.org/10.1111/bor.12273>
- 563 Conway, S. J., Butcher, F. E. G., de Haas, T., Deijns, A. A. J., Grindrod, P. M., & Davis, J. M.
 564 (2018). Glacial and gully erosion on Mars: A terrestrial perspective. *Geomorphology*,
 565 *318*, 26–57. <https://doi.org/10.1016/j.geomorph.2018.05.019>
- 566 Evatt, G. W., & Fowler, A. C. (2007). Cauldron subsidence and subglacial floods. *Annals of*
 567 *Glaciology*, *45*, 163–168. <https://doi.org/10.3189/172756407782282561>
- 568 Fassett, C. I., Dickson, J. L., Head, J. W., Levy, J. S., & Marchant, D. R. (2010). Supraglacial
 569 and proglacial valleys on Amazonian Mars. *Icarus*, *208*(1), 86–100.
 570 <https://doi.org/10.1016/j.icarus.2010.02.021>
- 571 Fastook, J. L., Head, J. W., Marchant, D. R., & Forget, F. (2008). Tropical mountain glaciers on
 572 Mars: Altitude-dependence of ice accumulation, accumulation conditions, formation
 573 times, glacier dynamics, and implications for planetary spin-axis/orbital history. *Icarus*,
 574 *198*(2), 305–317. <https://doi.org/10.1016/j.icarus.2008.08.008>
- 575 Fastook, J. L., Head, J. W., Marchant, D. R., Forget, F., & Madeleine, J.-B. (2012). Early Mars
 576 climate near the Noachian–Hesperian boundary: Independent evidence for cold
 577 conditions from basal melting of the south polar ice sheet (Dorsa Argentea Formation)
 578 and implications for valley network formation. *Icarus*, *219*(1), 25–40.

579 <https://doi.org/10.1016/j.icarus.2012.02.013>

580 Fisher, D. A., Hecht, M. H., Kounaves, S. P., & Catling, D. C. (2010). A perchlorate brine
581 lubricated deformable bed facilitating flow of the north polar cap of Mars: Possible
582 mechanism for water table recharging. *Journal of Geophysical Research*, *115*, E00E12,
583 <https://doi.org/10.1029/2009JE003405>

584 Flowers, G. E. (2015). Modelling water flow under glaciers and ice sheets. *Proceedings of the*
585 *Royal Society A: Mathematical, Physical and Engineering Sciences*, *471*(2176),
586 20140907–20140907. <https://doi.org/10.1098/rspa.2014.0907>

587 Fowler, A. C. (1999). Breaking the seal at Grímsvötn, Iceland. *Journal of Glaciology*, *45*(151),
588 506–516. <https://doi.org/10.1017/S0022143000001362>

589 Fricker, H. A., & Scambos, T. (2009). Connected subglacial lake activity on lower Mercer and
590 Whillans Ice Streams, West Antarctica, 2003–2008. *Journal of Glaciology*, *55*(190),
591 303–315. <https://doi.org/10.3189/002214309788608813>

592 Fricker, H. A., Scambos, T., Bindschadler, R., & Padman, L. (2007). An active subglacial water
593 system in West Antarctica mapped from space. *Science*, *315*(5818), 1544–1548.
594 <https://doi.org/10.1126/science.1136897>

595 Gallagher, C., & Balme, M. (2015). Eskers in a complete, wet-based glacial system in the
596 Phlegra Montes region, Mars. *Earth and Planetary Science Letters*, *431*, 96–109.
597 <https://doi.org/10.1016/j.epsl.2015.09.023>

598 Head, J. W., Mustard, J. F., Kreslavsky, M. A., Milliken, R. E., & Marchant, D. R. (2003).
599 Recent ice ages on Mars. *Nature*, *426*(6968), 797–802.
600 <https://doi.org/10.1038/nature02114>

601 Kress, A. M., & Head, J. W. (2015). Late Noachian and early Hesperian ridge systems in the
602 south circumpolar Dorsa Argentea Formation, Mars: Evidence for two stages of melting
603 of an extensive late Noachian ice sheet. *Planetary and Space Science*, *109–110*, 1–20.
604 <https://doi.org/10.1016/j.pss.2014.11.025>

605 Le Brocq, A. M., Payne, A. J., Siegert, M. J., & Alley, R. B. (2009). A subglacial water-flow
606 model for West Antarctica. *Journal of Glaciology*, *55*(193), 879–888.
607 <https://doi.org/10.3189/002214309790152564>

608 Levy, J. S., Fassett, C. I., Head, J. W., Schwartz, C., & Watters, J. L. (2014). Sequestered glacial
609 ice contribution to the global Martian water budget: Geometric constraints on the volume
610 of remnant, midlatitude debris-covered glaciers. *Journal of Geophysical Research:*
611 *Planets*, *119*(10), 2188–2196. <https://doi.org/10.1002/2014JE004685>

612 Levy, J. S., Fassett, C. I., & Head, J. W. (2016). Enhanced erosion rates on Mars during
613 Amazonian glaciation. *Icarus*, *264*, 213–219. <https://doi.org/10.1016/j.icarus.2015.09.037>

614 Livingstone, S. J., Clark, C. D., & Tarasov, L. (2013a). Modelling North American palaeo-
615 subglacial lakes and their meltwater drainage pathways. *Earth and Planetary Science*
616 *Letters*, *375*, 13–33. <https://doi.org/10.1016/j.epsl.2013.04.017>

617 Livingstone, S. J., Clark, C. D., Woodward, J., & Kingslake, J. (2013b). Potential subglacial lake
618 locations and meltwater drainage pathways beneath the Antarctic and Greenland ice
619 sheets. *The Cryosphere*, *7*(6), 1721–1740. <https://doi.org/10.5194/tc-7-1721-2013>

620 Llubes, M., Lanseau, C., & Rémy, F. (2006). Relations between basal condition, subglacial
621 hydrological networks and geothermal flux in Antarctica. *Earth and Planetary Science*
622 *Letters*, *241*(3–4), 655–662. <https://doi.org/10.1016/j.epsl.2005.10.040>

623 Martos, Y. M., Catalán, M., Jordan, T. A., Golynsky, A., Golynsky, D., Eagles, G., & Vaughan,
624 D. G. (2017). Heat Flux Distribution of Antarctica Unveiled. *Geophysical Research*

625 *Letters*, 44(22), 11,417–11,426. <https://doi.org/10.1002/2017GL075609>

626 Nye, J. F. (1976). Water flow in glaciers: jökulhlaups, tunnels and veins. *Journal of Glaciology*,
627 17(76), 181–207. <https://doi.org/10.1017/S002214300001354X>

628 Orosei, R., Lauro, S. E., Pettinelli, E., Cicchetti, A., Coradini, M., Cosciotti, B., et al. (2018).
629 Radar evidence of subglacial liquid water on Mars. *Science*, eaar7268.
630 <https://doi.org/10.1126/science.aar7268>

631 Pattyn, F. (2010). Antarctic subglacial conditions inferred from a hybrid ice sheet/ice stream
632 model. *Earth and Planetary Science Letters*, 295(3–4), 451–461.
633 <https://doi.org/10.1016/j.epsl.2010.04.025>

634 Plaut, J. J., Picardi, G., Safaeinili, A., Ivanov, A. B., Milkovich, S. M., Cicchetti, A., et al.
635 (2007). Subsurface radar sounding of the south polar layered deposits of Mars. *Science*,
636 316(5821), 92–95. <https://doi.org/10.1126/science.1139672>

637 Scanlon, K. E., Head, J. W., Fastook, J. L., & Wordsworth, R. D. (2018). The Dorsa Argentea
638 Formation and the Noachian-Hesperian climate transition. *Icarus*, 299, 339–363.
639 <https://doi.org/10.1016/j.icarus.2017.07.031>

640 Shreve, R. L. (1972). Movement of water in glaciers. *Journal of Glaciology*, 11(62), 205–214.
641 <https://doi.org/10.1017/S002214300002219X>

642 Shreve, R.L., 1985. Late Wisconsin ice-surface profile calculated from Esker paths and types.
643 Katahdin Esker System, Maine. *Quaternary Research* 23, 27–37.
644 [https://doi.org/10.1016/0033-5894\(85\)90069-9](https://doi.org/10.1016/0033-5894(85)90069-9).

645 Smith, D. E., Zuber, M. T., Frey, H. V., Garvin, J. B., Head, J. W., Muhleman, D. O., et al.
646 (2001). Mars Orbiter Laser Altimeter: Experiment summary after the first year of global
647 mapping of Mars. *Journal of Geophysical Research: Planets*, 106(E10), 23689–23722.
648 <https://doi.org/10.1029/2000JE001364>

649 Sori, M. M., & Bramson, A. M. (2019). Water on Mars, with a grain of salt: local heat anomalies
650 are required for basal melting of ice at the south pole today. *Geophysical Research*
651 *Letters*, 46(3), 1222–1231. <https://doi.org/10.1029/2018GL080985>

652 Souness, C., & Hubbard, B. (2012). Mid-latitude glaciation on Mars. *Progress in Physical*
653 *Geography: Earth and Environment*, 36(2), 238–261.
654 <https://doi.org/10.1177/0309133312436570>

655 Spring, U., & Hutter, K. (1981). Numerical studies of Jökulhlaups. *Cold Regions Science and*
656 *Technology*, 4(3), 227–244. [https://doi.org/10.1016/0165-232X\(81\)90006-9](https://doi.org/10.1016/0165-232X(81)90006-9)

657 Storrar, R. D., Stokes, C. R., & Evans, D. J. A. (2013). A map of large Canadian eskers from
658 Landsat satellite imagery. *Journal of Maps*, 9(3), 456–473.
659 <https://doi.org/10.1080/17445647.2013.815591>

660 Stroeven, A. P., Hättestrand, C., Kleman, J., Heyman, J., Fabel, D., Fredin, O., et al. (2016).
661 Deglaciation of Fennoscandia. *Quaternary Science Reviews*, 147, 91–121.
662 <https://doi.org/10.1016/j.quascirev.2015.09.016>

663 Tanaka, K.L., Skinner, J.A., Jr., Dohm, J.M., Irwin, R.P., III, Kolb, E.J., Fortezzo, C.M., Platz,
664 T., Michael, G.G., and Hare, T.M., (2014), Geologic map of Mars:
665 U.S. Geological Survey Scientific Investigations Map 3292, scale 1:20,000,000,
666 pamphlet 43 p., <http://dx.doi.org/10.3133/sim3292>

667 Wieczorek, M. (2008). Constraints on the composition of the Martian south polar cap from
668 gravity and topography. *Icarus*, 196(2), 506–517.
669 <https://doi.org/10.1016/j.icarus.2007.10.026>

670 Willis, I. C., Pope, E. L., Leysinger Vieli, G. J.-M. C., Arnold, N. S., & Long, S. (2016).

671 Drainage networks, lakes and water fluxes beneath the Antarctic ice sheet. *Annals of*
672 *Glaciology*, 57(72), 96–108. <https://doi.org/10.1017/aog.2016.15>
673 Wingham, D. J., Siegert, M. J., Shepherd, A., & Muir, A. S. (2006). Rapid discharge connects
674 Antarctic subglacial lakes. *Nature*, 440(7087), 1033–1036.
675 <https://doi.org/10.1038/nature04660>
676 Wordsworth, R., Forget, F., Millour, E., Head, J. W., Madeleine, J.-B., & Charnay, B. (2013).
677 Global modelling of the early Martian climate under a denser CO₂ atmosphere: Water
678 cycle and ice evolution. *Icarus*, 222(1), 1–19.
679 <https://doi.org/10.1016/j.icarus.2012.09.036>
680 Wright, A. P., Siegert, M. J., Le Brocq, A. M., & Gore, D. B. (2008). High sensitivity of
681 subglacial hydrological pathways in Antarctica to small ice-sheet changes. *Geophysical*
682 *Research Letters*, 35(17). <https://doi.org/10.1029/2008GL034937>
683 Zuber, M. T., Phillips, R. J., Andrews-Hanna, J. C., Asmar, S. W., Konopliv, A. S., Lemoine, F.
684 G., et al. (2007). Density of Mars' south polar layered deposits. *Science*, 317(5845),
685 1718–1719. <https://doi.org/10.1126/science.1146995>
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688 **Figure Captions**

689 **Figure 1. a.** Map of the SPLD, showing the region containing the HRA argued to be subglacial
690 liquid by *Orosei et al.*, [2018]. The surface topography is taken from the MOLA topographic
691 dataset [*Smith et al.*, 2001]. The black contour shows the outline of the late Amazonian polar cap
692 (1Apc) unit [*Tanaka et al.*, 2014]; the black square delineates the area investigated in *Orosei et*
693 *al.*, [2018], and which we investigate in this study. **b.** MOLA surface topography of the study
694 area highlighted by the black square in a. The red contour shows the location of the HRA
695 reported by *Orosei et al.*, [2018]. Color scale for both maps shows MOLA elevation in m; X and
696 Y axes in B are distances in km from the center of the study area.

697 **Figure 2.** MARSIS points for the 200×200 km area containing the HRA (red contour) reported
698 by *Orosei et al.*, [2018], shown in Figure 1. **a.** Unadjusted (raw) bed elevation (m). **b-d.** Mean
699 (**b**), median (**c**) and standard deviation (**d**) of elevation difference (m) between each point and the
700 set of points within the sensor Fresnel radius. X and Y axis units are km from the center of the
701 area; color scale units are meters. Dots are scaled to the Fresnel radius of ~ 4 km.

702 **Figure 3.** Frequency distributions of the elevation differences for each point and its neighbours
703 within the Fresnel radius for that point. **a.** Overall distribution for all points. **b.** The point with
704 the maximum mean elevation difference. **c.** The point with the minimum mean elevation
705 difference. **d.** The point with the maximum standard deviation of elevation difference. X axis
706 units are meters of elevation difference; Y axis units are point counts. Red line shows the best-fit
707 normal distribution.

708 **Figure 4.** Gridded 200 m topography of study area sub-ice bed based on: **a.** raw MARSIS radar
709 data (Figure 2a, *Orosei et al.*, [2018]; **b-c:** mean (**b**) and median (**c**) elevation of all points in the
710 Fresnel radius of each original point (Figure 2b-c); **d.** mean elevation of the 1000 perturbed
711 topographies (see text). X and Y axis units are km from the center of the area; color scale shows
712 elevation in meters. The red contour shows the HRA reported by *Orosei et al.*, [2018].

713 **Figure 5.** Calculated flow accumulation values at 200 m resolution for: **a.** Interpolated raw
714 topography. **b.** Mean in-radius topography. **c.** Median in-radius topography. **d.** Mean-perturbed
715 topography (see text). X and Y axis units are km from the center of the area; color scale units are
716 the common logarithm of the upstream area flowing into each cell in km^2 . The red contour shows
717 the HRA reported by *Orosei et al.*, [2018].

718 **Figure 6.** Calculated flow accumulation grids at 200 m resolution for the central region around
719 the HRA: **a.** Interpolated raw topography. **b.** Mean in-radius topography. **c.** Median in-radius
720 topography. **d.** Mean-perturbed topography (see text). X and Y axis units are km from the center
721 of the area; color scale units are the common logarithm of the upstream area flowing into each
722 cell in km^2 . The red contour shows the HRA reported by *Orosei et al.*, [2018].

723 **Figure 7. a-b.** Results of the 1000 perturbed-topography experiments at 200 m resolution **a.**
724 Probability value that any given DEM cell is within a lake for the set of 1000 perturbed-
725 topography runs. **b.** Mean upstream area for the flow accumulation algorithm results. **c-f.** Results
726 of the variable liquid and ice density experiments using the 200 m mean-perturbed topography
727 (shown in Figure 4d). **c.** Probability value that any given DEM cell is within a lake for the

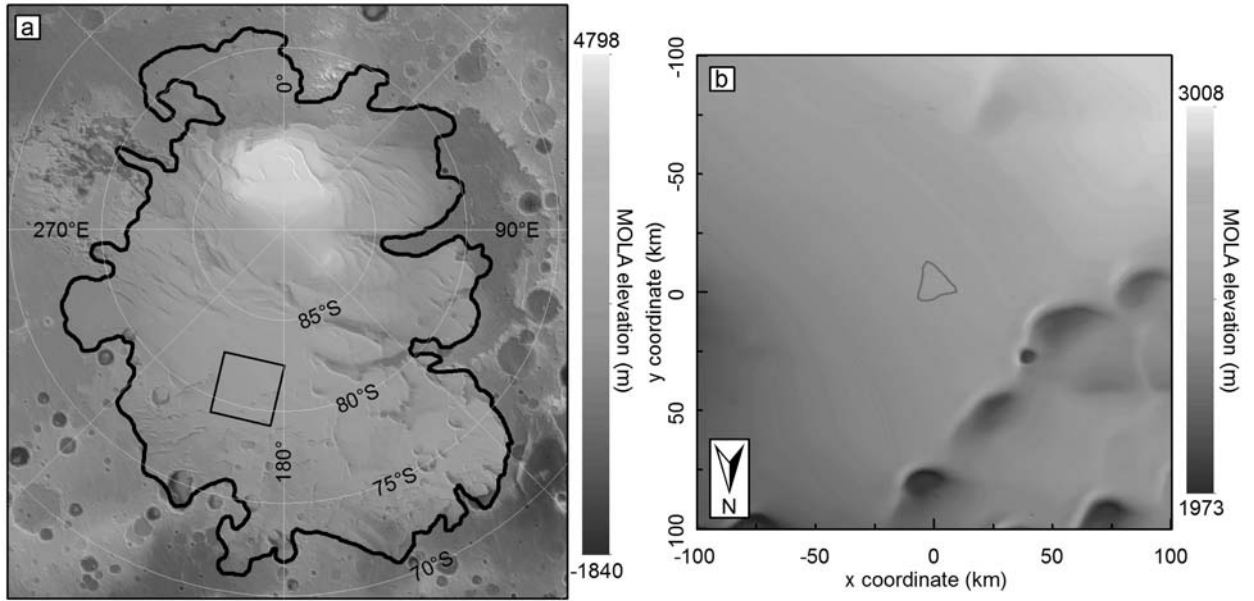
728 variable liquid density experiments. **d.** Mean upstream area for the flow accumulation algorithm
729 runs. **e.** Calculated flow accumulation grid for the $\rho_w = 1000 \text{ kg m}^{-3}$ experiment with standard ρ_i .
730 **f.** Calculated flow accumulation grid for the $\rho_i = 1200 \text{ kg m}^{-3}$, $\rho_w = 1000 \text{ kg m}^{-3}$ experiment.
731 Color scale units for a and c are probability values; $P = 0$ shows a pixel was never calculated to
732 be in a lake; $P = 1$ shows a pixel was calculated to be in a lake in all simulations. Color scale
733 units for b, d, e and f are the common logarithm of the upstream area flowing into each cell in
734 km^2 . X and Y axis units are km from the center of the area. The red contour shows the HRA
735 reported by *Orosei et al.*, [2018]. Black line labeled X–Y in Panel B shows the location of the
736 transect shown in Figure 9.

737 **Figure 8.** Mean upstream area for the set of 1000 reduced resolution perturbed topography runs.
738 **a.** 500 m resolution. **b.** 1000 m resolution. **c.** 2500 m resolution. **d.** 5000 m resolution. Color
739 scale units are the common logarithm of the upstream area flowing into each cell in km^2 . X and
740 Y axis units are km from the center of the area. The red contour shows the HRA reported by
741 *Orosei et al.*, [2018].

742 **Figure 9.** Cross-section through the HRA along transect X–Y shown in Figure 7b. Basal
743 topography is from the 200 m mean-perturbed DEM. The black arrow shows the extent of the
744 HRA. Note the discontinuity in the left-hand Yaxis to improve visibility of the basal topography.

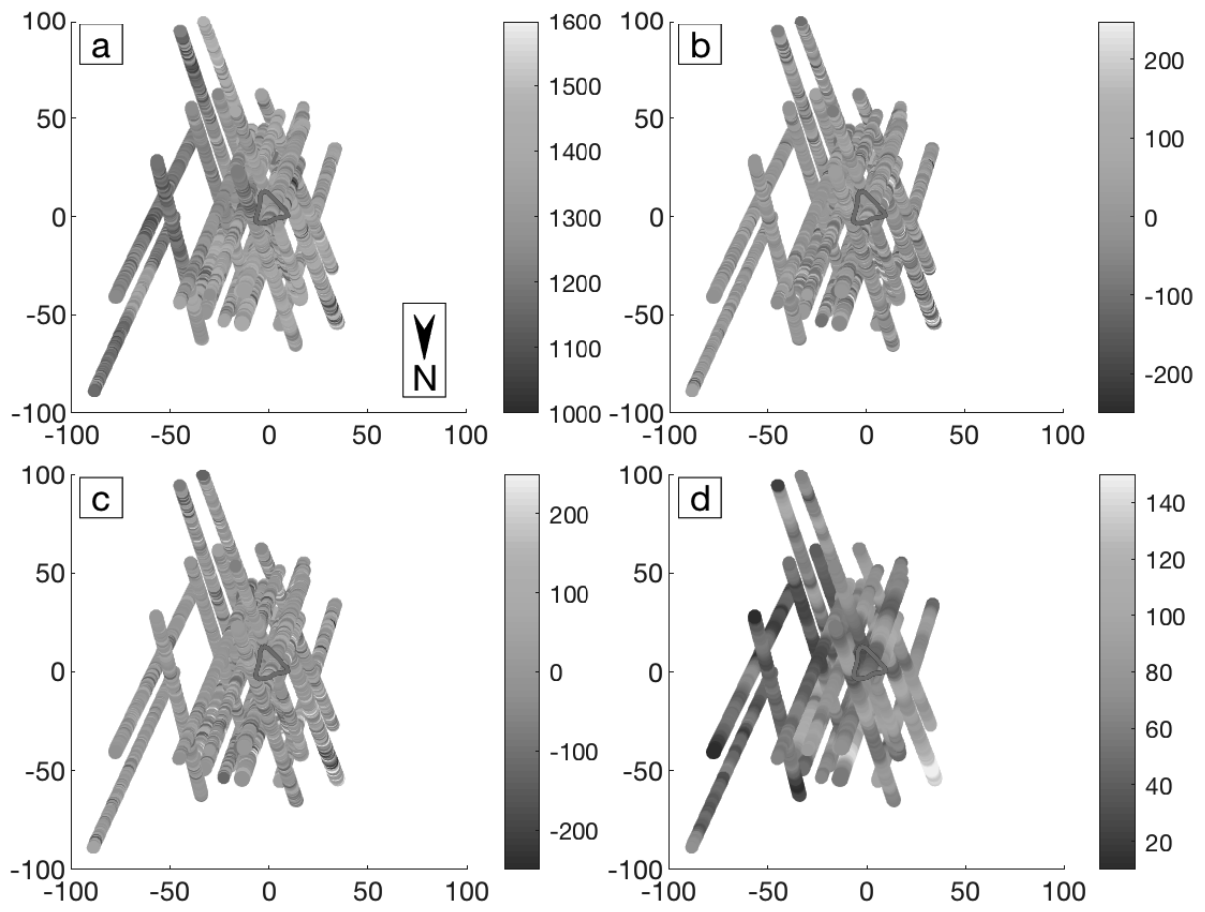
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747 Figure 1:



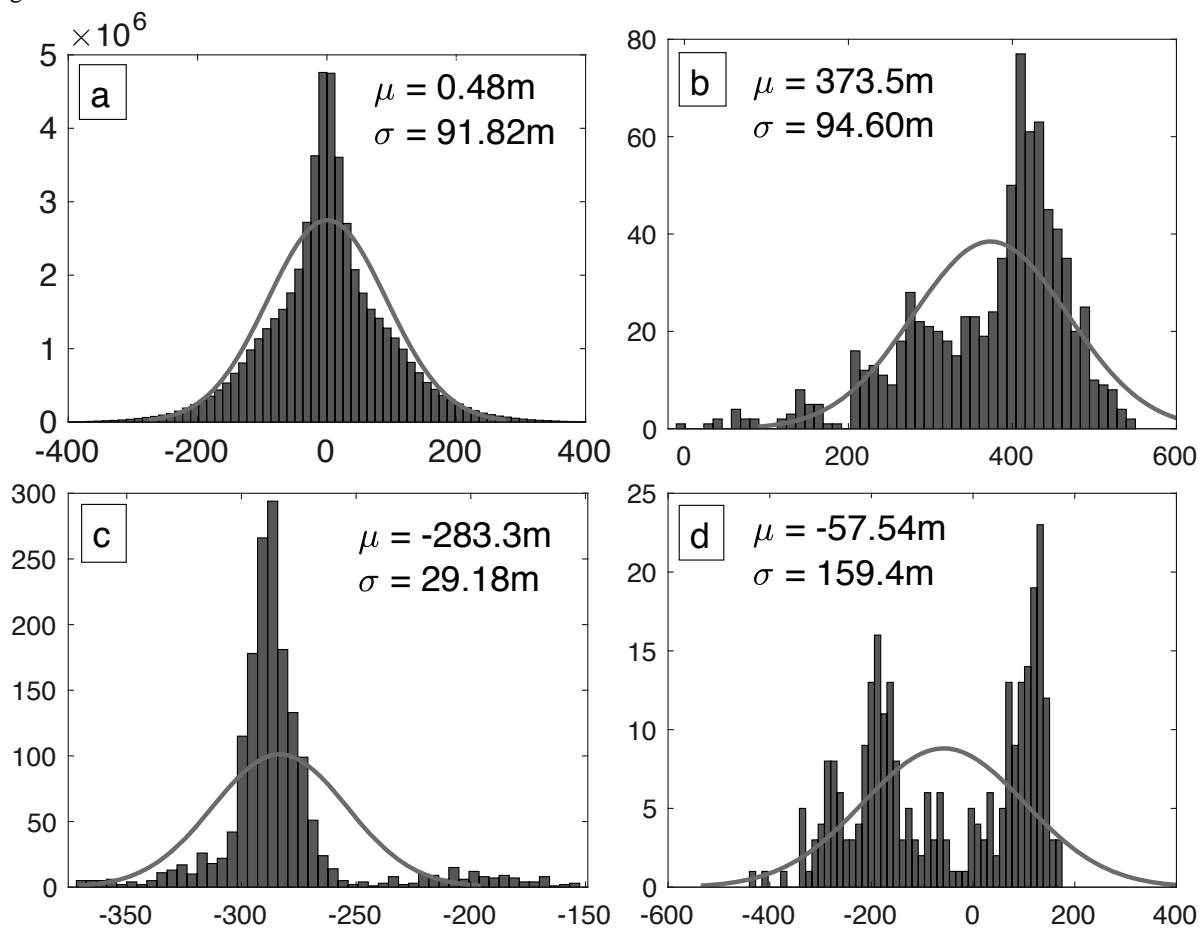
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750 Figure 2



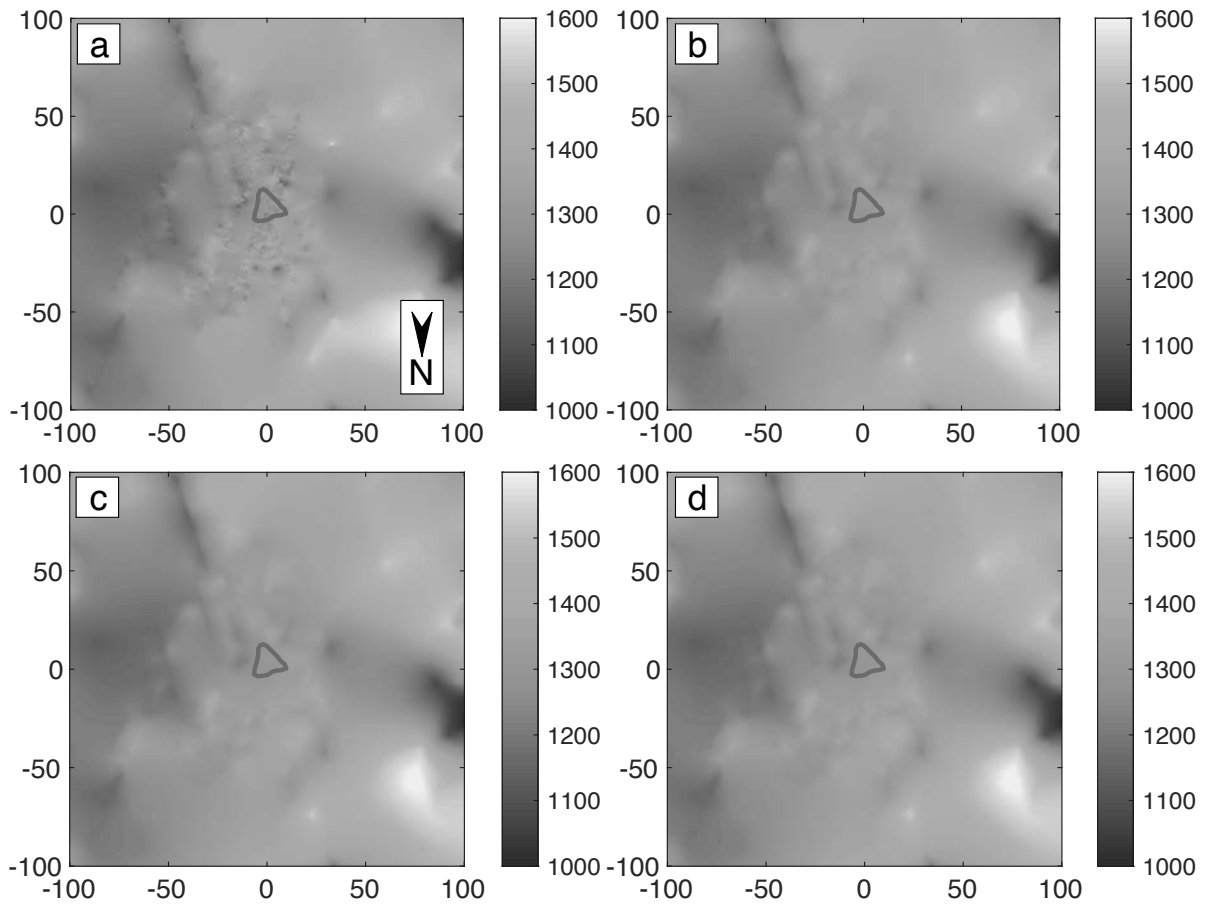
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754 Figure 3



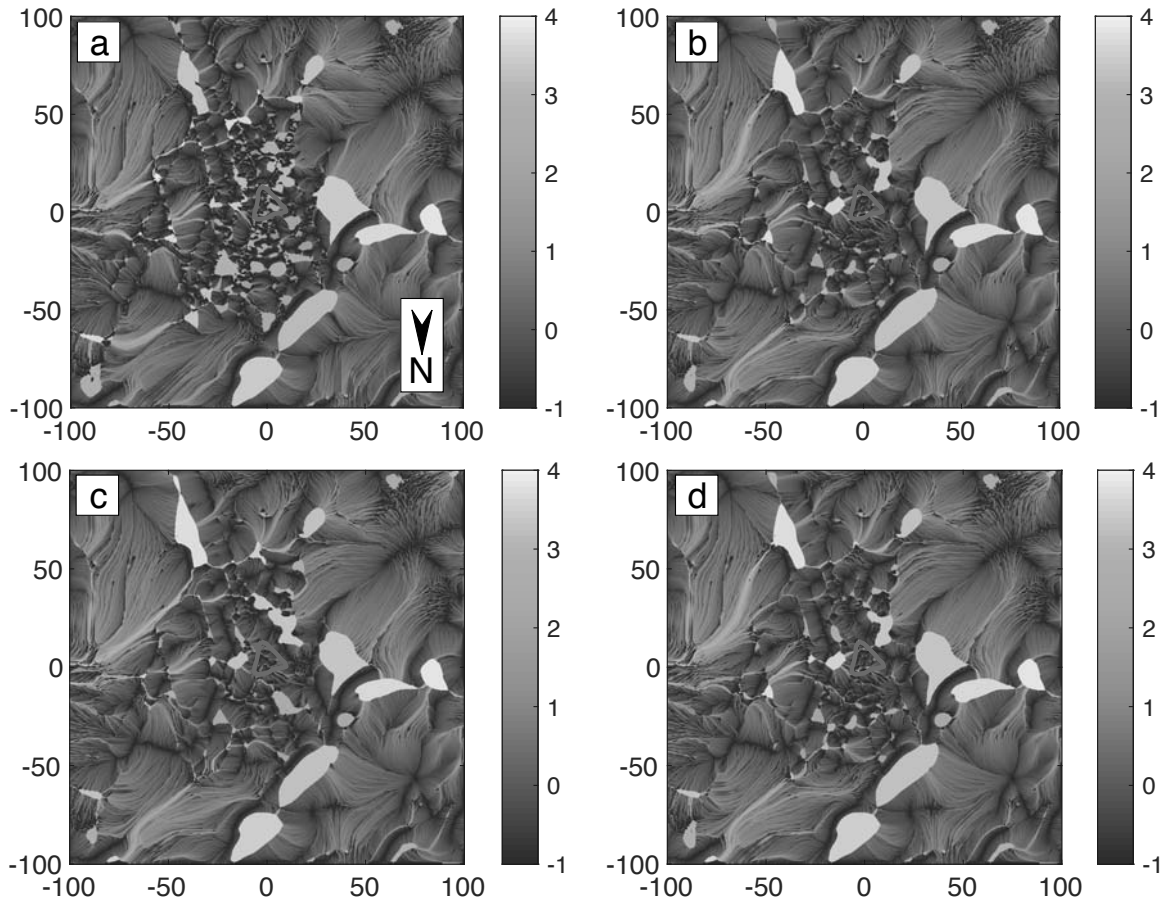
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758 Figure 4

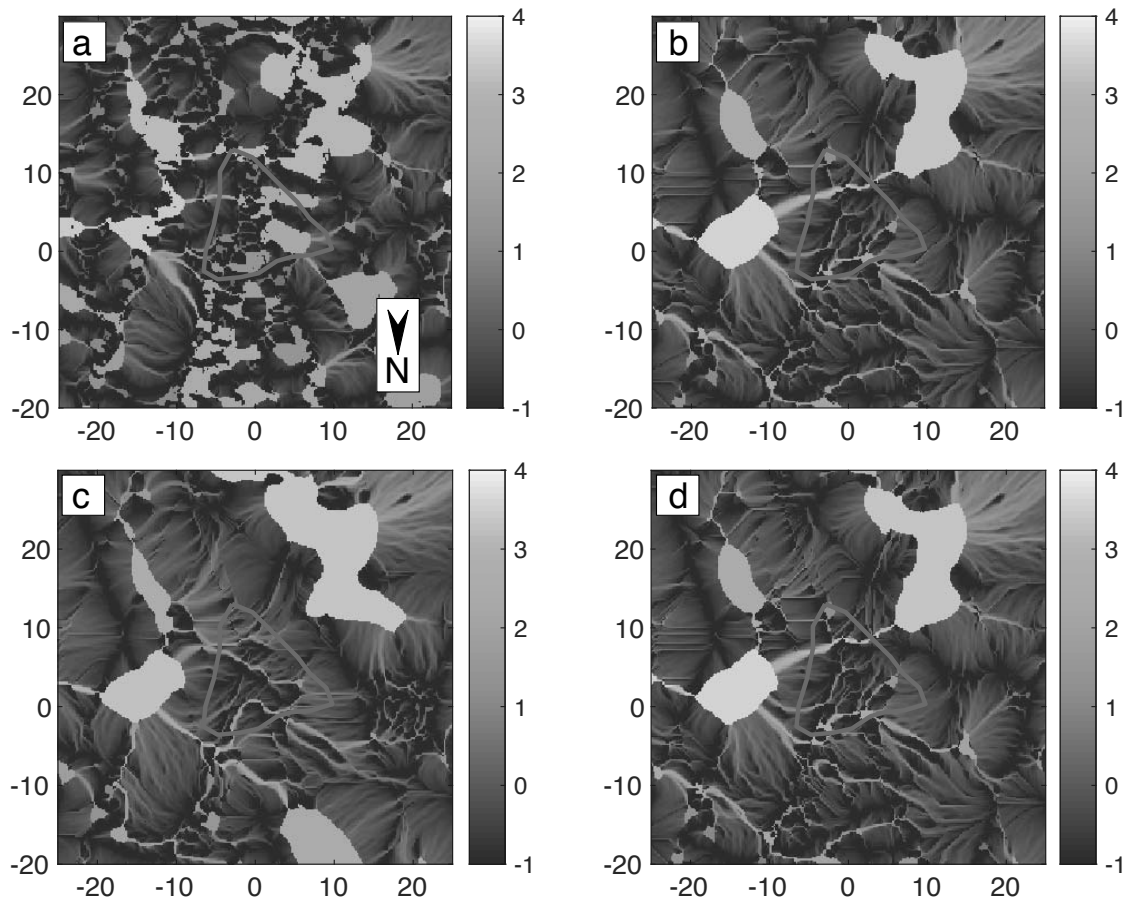


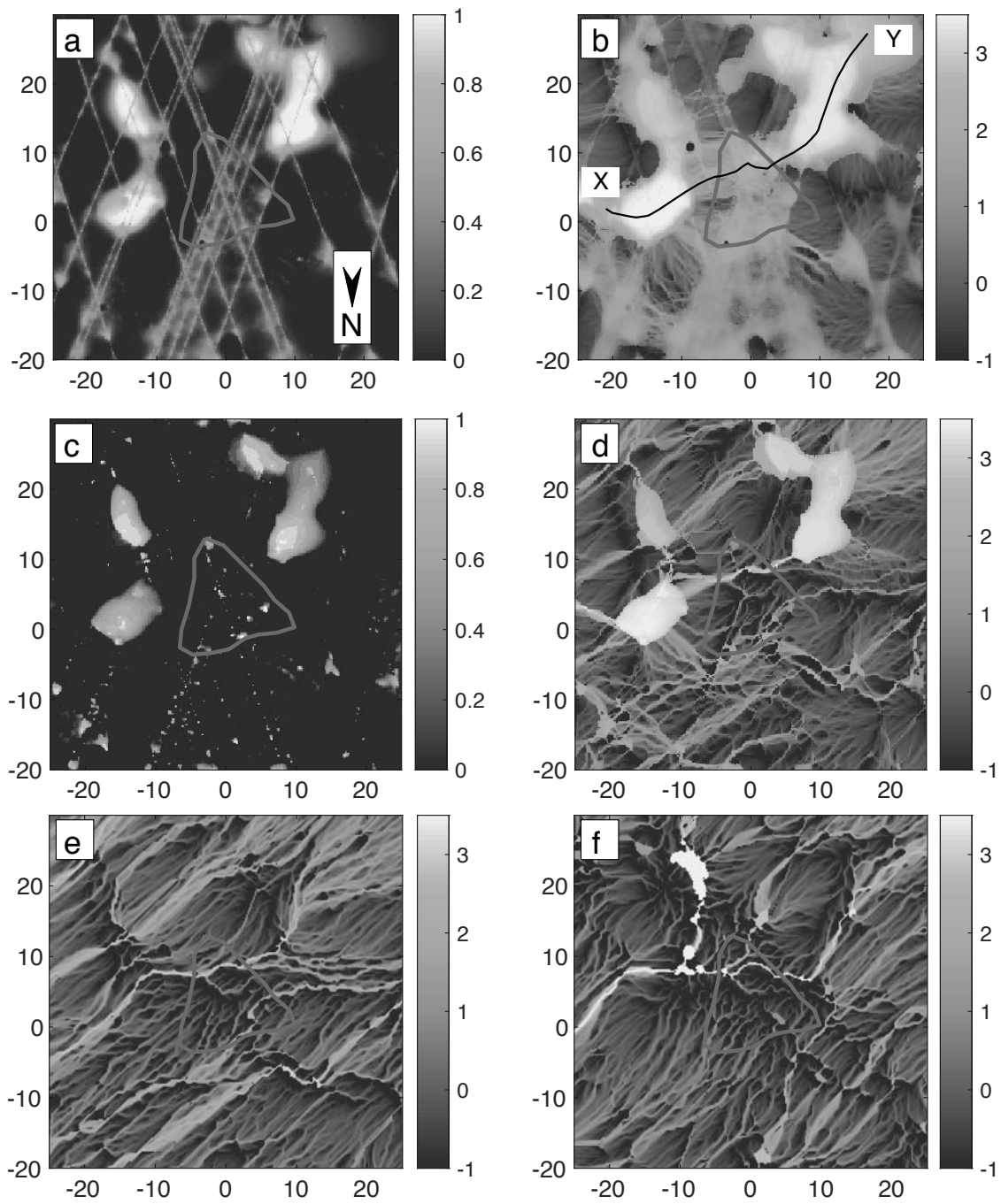
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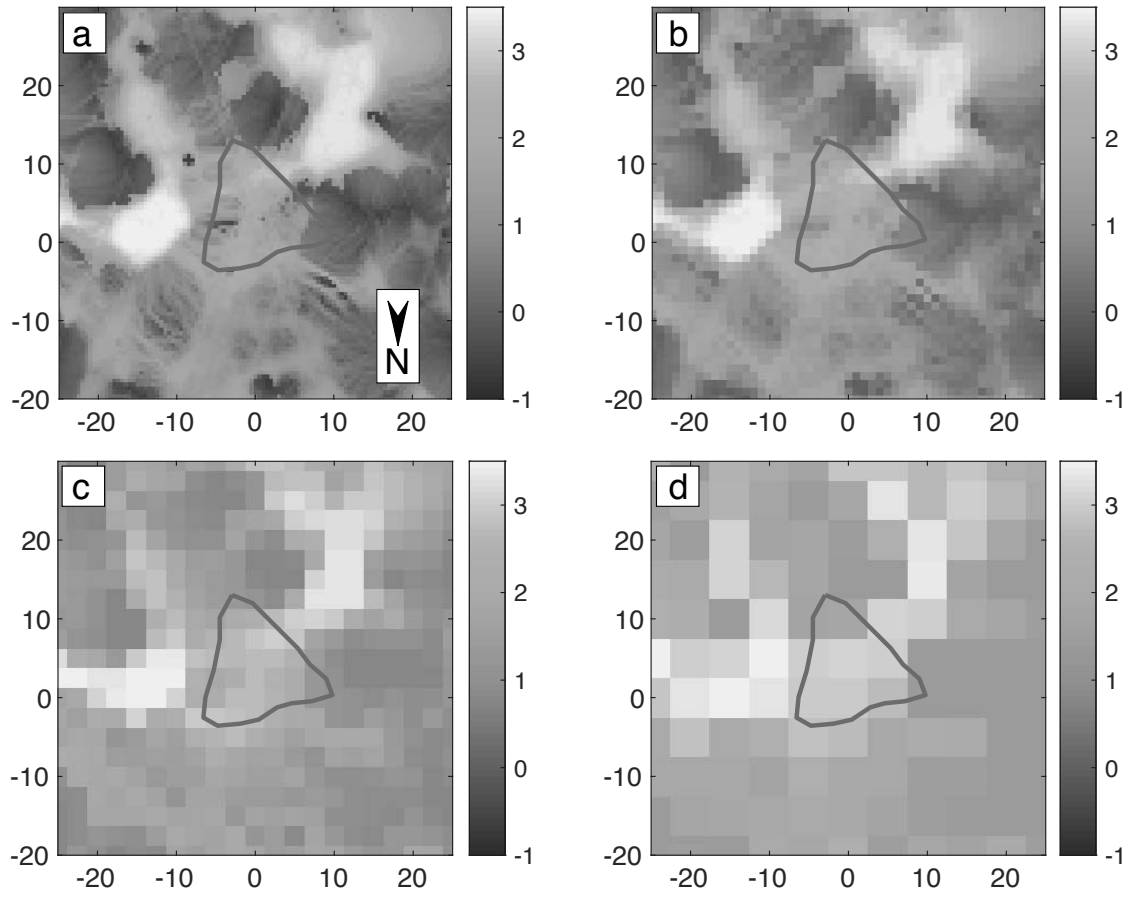
762 Figure 5



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778 Figure 9.

