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1 **Modelling reconstruction and boulder size-frequency distribution of a young** 2 **(< 5 Myr) landslide located in Simud Vallis floor, Mars.**

3 Pajola Maurizio¹, Mergili Martin², Cambianica Pamela¹, Lucchetti Alice¹, Brunetti Maria
 4 Teresa³, Guimpier Anthony⁴, Mastropietro Maria⁵, Munaretto Giovanni^{6,1}, Conway Susan⁴,
 5 Beccarelli Joel⁶, Cremonese Gabriele¹.

6 ¹ INAF, Astronomical Observatory of Padova, Vic. Osservatorio 5, 35122 Padova, Italy (maurizio.pajola@inaf.it)

7 ² University of Graz, Graz, Austria

8 ³ Research Institute for Geo-Hydrological Protection-Italian National Research Council, Perugia, Italy,

9 ⁴ Laboratoire de Planétologie et Géodynamique, UMR6112 CNRS, Université de Nantes, France

10 ⁵ Braunschweig TU University, Braunschweig, Germany

11 ⁶ University of Padova, Padova, Italy

12

13 **Abstract**

14 We focus on a young (~ **4.5 Ma**), 3.4 km long landslide located in the floor of Simud Vallis,
 15 Oxia Palus Quadrangle of Mars. By making use of a **2 m-scale** HiRISE DEM we reconstruct
 16 the terrain surface before the landslide and in doing so we estimate the release and deposition
 17 heights and volumes related to the different stages of the landslide. Using the *r.avaflow* software
 18 we simulate the mass movement as a multi-stage event, and obtain simulated deposits that are
 19 both spatially and longitudinally comparable to the current landslide deposits. Through two
 20 0.25 m-scale HiRISE images we identify and manually count > 130 000 boulders that are
 21 located along the landslide, deriving their size-frequency distribution and spatial density per
 22 unit area for boulders with an equivalent diameter ≥ 1.75 m. Our analyses reveal that the
 23 distribution is of a Weibull-type, suggesting that the rocky constituents fractured and
 24 fragmented progressively during the course of the mass movement, consistent with our
 25 proposed two-stage model of landslide motion.

26 **1.0 Introduction**

27 During the last 50 years multiple planetary exploration missions have collected high-resolution
 28 images of Solar System bodies, which have been sufficient to detect and identify detailed
 29 structures and scales of surface landforms, such as landslides. Using as a reference similar
 30 structures observed on Earth (terrestrial analogues) and using the same classification system,
 31 we observe that the characteristics of landslides on other solid bodies can be just as variable as
 32 those on Earth (**Hargitai and Kereszturi, 2015**).

33 So far, landslides have been observed on Mercury (Xiao and Komatsu, 2013; Blewett et al.,
 34 2013; Brunetti et al., 2015), Venus (Malin, 1992), the Moon (Pike, 1971; Lindsay, 1976; Xiao
 35 et al., 2013; Brunetti et al., 2015), Phobos (Shi et al., 2016), asteroids (Massironi, et al., 2012;
 36 Magrin et al., 2012; Duarte et al., 2019; Jawin et al., 2020), icy satellites (Schenk and Bulmer,
 37 1998; Chuang and Greeley, 2000; Singer et al., 2012, Beddingfield et al., 2021) and comets
 38 (Lucchetti et al., 2019). Among terrestrial planets, Mars has been the most studied with respect
 39 to landslides (e.g., Lucchitta, 1979; McEwen, 1989; Shaller and Komatsu, 1994; Quantin et al.,
 40 2004; Soukhovitskaya and Manga, 2006; Bigot-Cormier and Montgomery, 2007; Lucas and
 41 Mangeney, 2007, Brunetti et al., 2014; Crosta et al., 2018). **On Mars, landslides generally**
 42 **much larger than the terrestrial subaerial ones¹** are located on slopes (Sharp, 1973) in

¹ Landslides in Valles Marineris, which are among the largest found on Mars, are comparable in size to the largest submarine landslides of the Earth (Brunetti et al., 2014).

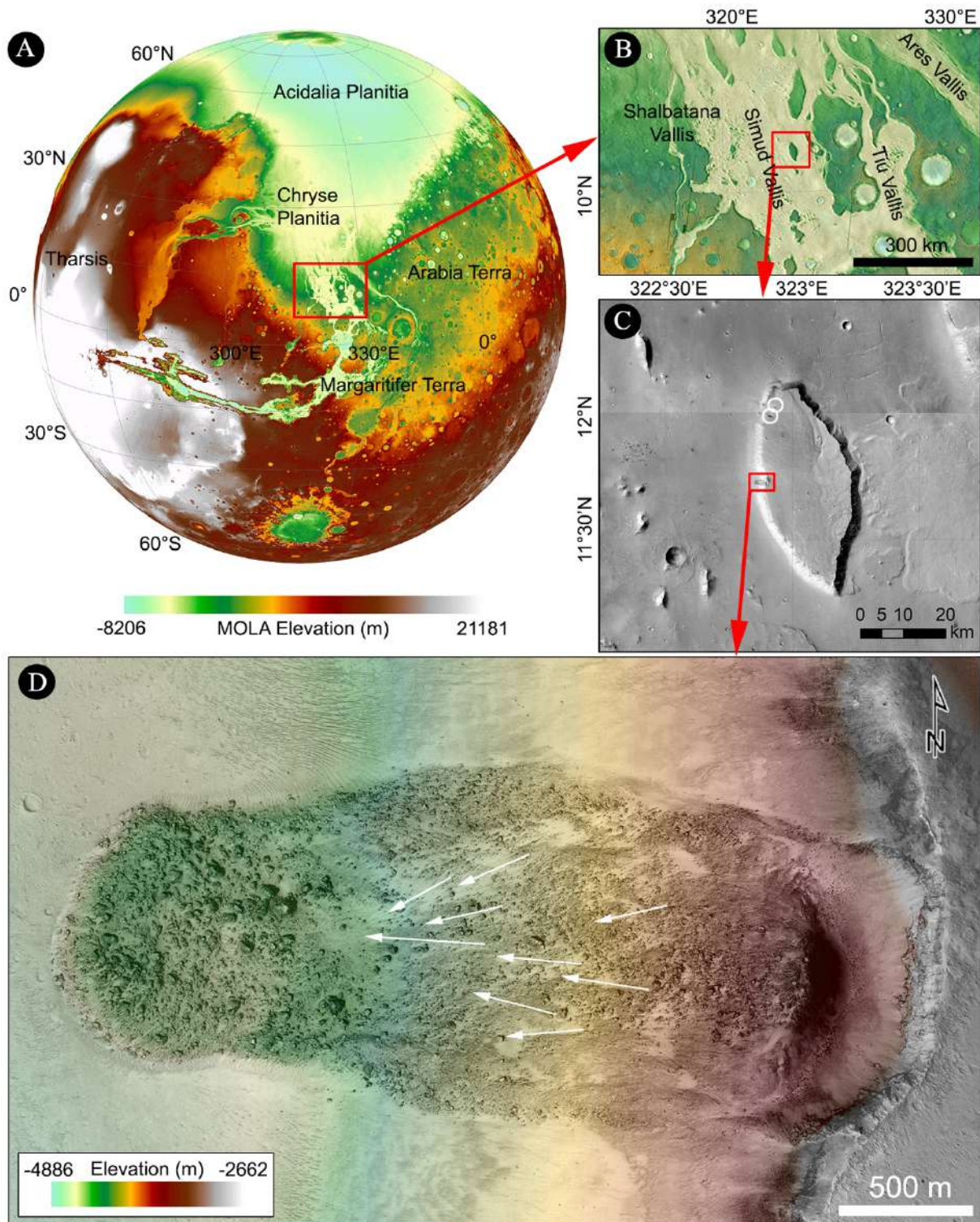
43 canyons (e.g., Valles Marineris), as well as on mountainous reliefs (e.g., Olympus Mons) and
44 crater structures.

45 As on Earth, the morphology (e.g., area, volume, runout length, drop height, width, and texture)
46 of the landslide deposit on extra-terrestrial bodies depends on factors such as the slope of the
47 topography, the collapse mechanism, the mechanical properties of the material, the presence of
48 fluids and volatiles within the sliding material, and the local environmental conditions (e.g., the
49 gravitational acceleration). Potential landslide preparatory and triggering factors, on the other
50 hand, can vary from one planet to another. For example, on Earth the presence of a thick
51 atmosphere means the main cause of landslides is rainfall, whereas on Mars the low-density
52 atmosphere means this mechanism cannot be inferred for recent landslides. Because we have
53 never directly observed any Martian landslide, their preparatory and triggering factors have to
54 be inferred from their morphology and context. Terrestrial analogues, although useful to infer
55 these factors, are limited because they are not the same size and are directly or indirectly
56 influenced by liquid water. On Mars, the landslides in Valles Marineris have been suggested to
57 be caused by meteorite impacts: either directly caused by the impact-shaking or as a
58 consequence of the decrease in the rock cohesion and potentially the subsequent increase of
59 ground water flowing through the pore fraction (Crosta et al., 2014). According to Kumar et al.
60 (2019), recent (last thousands of years) seismic shaking would be the cause of rock falls along
61 the Martian slopes. The same type of landslide observed on impact crater slopes was also
62 attributed to thermal-stress (Tesson et al., 2020). Recently, Bishop et al. (2021) suggested that
63 the occurrence of Martian landslides could be related to the subsurface cryosalt expansion. They
64 used analogue field investigations on Earth and laboratory experiments to demonstrate that
65 when salts interact with gypsum or water underground, it can cause disruptions on the surface,
66 and may trigger collapses and landslides. The wide range of hypotheses for potential causes of
67 Martian landslides shows that, the lack of direct Earth analogues makes the precise
68 identification of the preparatory and triggering factors of landslides on Mars challenging, with
69 the potential role of water and/or active tectonics being actively debated within the scientific
70 community.

71 Recently, Guimpier et al. (2021) focussed on three small and **young (with estimated ages not**
72 **exceeding 20 Ma)** Martian landslides with volumes $< 10^{10} \text{ m}^3$, with the main aim to compare
73 them with similarly sized terrestrial analogues. In particular, by using the Mars Reconnaissance
74 Orbiter's (MRO, Zurek et al., 2007) High Resolution Imaging Science Experiment images
75 (HiRISE, McEwen et al., 2007), Guimpier et al. (2021) identified and studied a landslide that
76 is located in Simud Vallis (Oxia Palus quadrangle), a large outflow channel that together with
77 Tiu Vallis once connected the Valles Marineris with the Chryse Planitia (Pajola et al., 2016a,
78 Fig. 1A, B). On Simud Vallis' floor multiple teardrop-shaped islands are present, all elongated
79 in the S-N direction of the flow (Carr and Clow, 1981) that incised the Mid-Noachian plateau
80 (Tanaka et al., 2014; Pajola et al., 2016a). On the western side of one of such landforms (Fig.
81 1C), is the Simud Vallis landslide (hereafter called *SV landslide*), Fig. 1D, characterised by
82 numerous boulders on its deposits. **In particular, by estimating the maximum age of**
83 **formation of this landslide through the use of the crater-size frequency distribution**
84 **(Michael and Neukum, 2010) technique (we underline that only one 23 m size crater is**
85 **identified on the SV deposits), Guimpier et al. (2021) derived a modelled age of $\sim 4.5 \pm 4$**
86 **Ma: hence the SV landslide formed recently, i.e. during the Late Amazonian period of**
87 **Mars geochronology.**

88 By using Earth-Mars comparative morphological analyses performed through similar resolution
89 Digital Elevation Models (DEMs), Guimpier et al. (2021) inferred that the triggering processes
90 of this landslide could have been the shaking by a meteorite impact or a marsquake and the
91 dynamics of the landslide was similar to a dry rock avalanche. As highlighted in Guimpier et
92 al. (2021), the *SHALTOP* numerical modelling (Bouchut et al., 2003; Bouchut and
93 Westdickenberg, 2004; Lucas and Mangeney, 2007) applied on the SV landslide was not
94 conclusive, and further numerical tests are needed to achieve a better representation of the mass
95 movement and its final shape in order to better understand the landslide dynamics (e.g. wet vs
96 dry flow). Guimpier et al., (2021) did not study the size-frequency distribution (SFD) of
97 boulders that characterise the landslide's surface. Yet, as shown by previous studies on the
98 Moon (Cintala and McBride, 1995; Bart and Melosh, 2010; Khrishna and Kumar, 2016; Pajola
99 et al., 2019), asteroids (Geissler et al., 1996; Thomas et al., 2001; Küppers et al., 2012; Mazrouei
100 et al., 2014; Burke et al., 2021, Schröder et al., 2021a), comets (Pajola et al., 2015; Mottola et
101 al., 2015; Pajola et al., 2016b,c; Pajola et al., 2017a,b), icy satellites (Pajola et al., 2021) and
102 Mars (Garvin et al., 1981; Grant et al., 2006; Golombek et al., 2008; Pajola et al., 2017c;
103 Mastropietro et al., 2020) the form of the SFD of blocks/boulders can provide important
104 information on how the parent rock was broken up and on the associated formative and
105 degradation processes.

106 In this work, we therefore aim to understand in detail how the initial mass moved and
107 fragmented using numerical **modelling** and by quantifying the boulder SFD, hence providing
108 insights into “recent” surface process that may still be occurring elsewhere on Mars.



109

110 **Figure 1:** A-B) The location of the Simud Vallis study area on Mars. C) The teardrop-shaped island
 111 where the SV landslide is located. **The white circles show other locations where circular-shaped**
 112 **erosion zones are identified, as well as degraded deposits at their feet.** D) The HiRISE
 113 ESP_050033_1920 image (see Dataset and Methodology, Table 1) showing the SV landslide, together
 114 with the elevation values (in metres) derived from the HiRISE DTM overlaid in transparency. **The white**
 115 **arrows indicate the set of trenches that align parallel to the local direction of motion.**

116

117

118 2.0 Datasets and Methodology

119 2.1 Imagery and DEM

120 The SV landslide was first observed with the HiRISE camera in October 2007 with a phase
121 angle of 36.2°, an incidence angle of 41.0° and an original image scale of 0.28 m/pixel. In
122 March 2017 the landslide was again imaged through HiRISE, with a phase angle of 60.9°, an
123 incidence angle of 34.0° and a spatial scale of 0.32 m/pixel. The details for each acquisition are
124 presented in Table 1, the original imagery dataset can be downloaded from
125 www.uahirise.org/PSP_005701_1920 and www.uahirise.org/ESP_050033_1920, while the
126 resulting anaglyph is presented in Supplementary Material Fig SM1. **The slightly different**
127 **observing conditions (both illumination and geometry of the acquisitions)** of the two
128 images provided the possibility to prepare a DEM of the landslide through the Ames Stereo
129 Pipeline (Moratto et al., 2010) with a spatial resolution of 2 m (Guimpier et al., 2021), Fig. 1D.
130 This product has been vertically controlled to ESA's Mars Express High Resolution Stereo
131 Camera (HRSC, Neukum and Jaumann, 2004) publicly available DEMs.

HiRISE Image ID	Acquisition Date	Pixel Scale (m)	Local Time (hh:mm)	Phase Angle (°)	Incidence Angle (°)	Emission Angle (°)
PSP_005701_1920	14/10/2007	0.25	14:12	36.2	41.0	5.0
ESP_050033_1920	30/03/2017	0.25	13:51	60.9	34.0	28.7

132

133 Table 1: HiRISE image ID, date of acquisition, map projected (equirectangular) pixel scale, local time
134 and observing angles of the images used for the DEM generation.

135 **The mentioned 2 m-scale DEM is extremely useful to both evaluate the post-collapse**
136 **surface topography and to reconstruct the pre-landslide terrain including its longitudinal**
137 **profile. These are pivotal aspects to then simulate the landslide formation and subsequent**
138 **movement. Moreover, the acquired images are characterised by shadows (if they were**
139 **absent, they would hamper the boulder identification, if they were too elongated, they**
140 **would cover nearby features) that are optimal for the detailed identification of the**
141 **boulders located in the landslide deposits, hence providing the possibility to perform a**
142 **size-frequency distribution (SFD) analysis.**

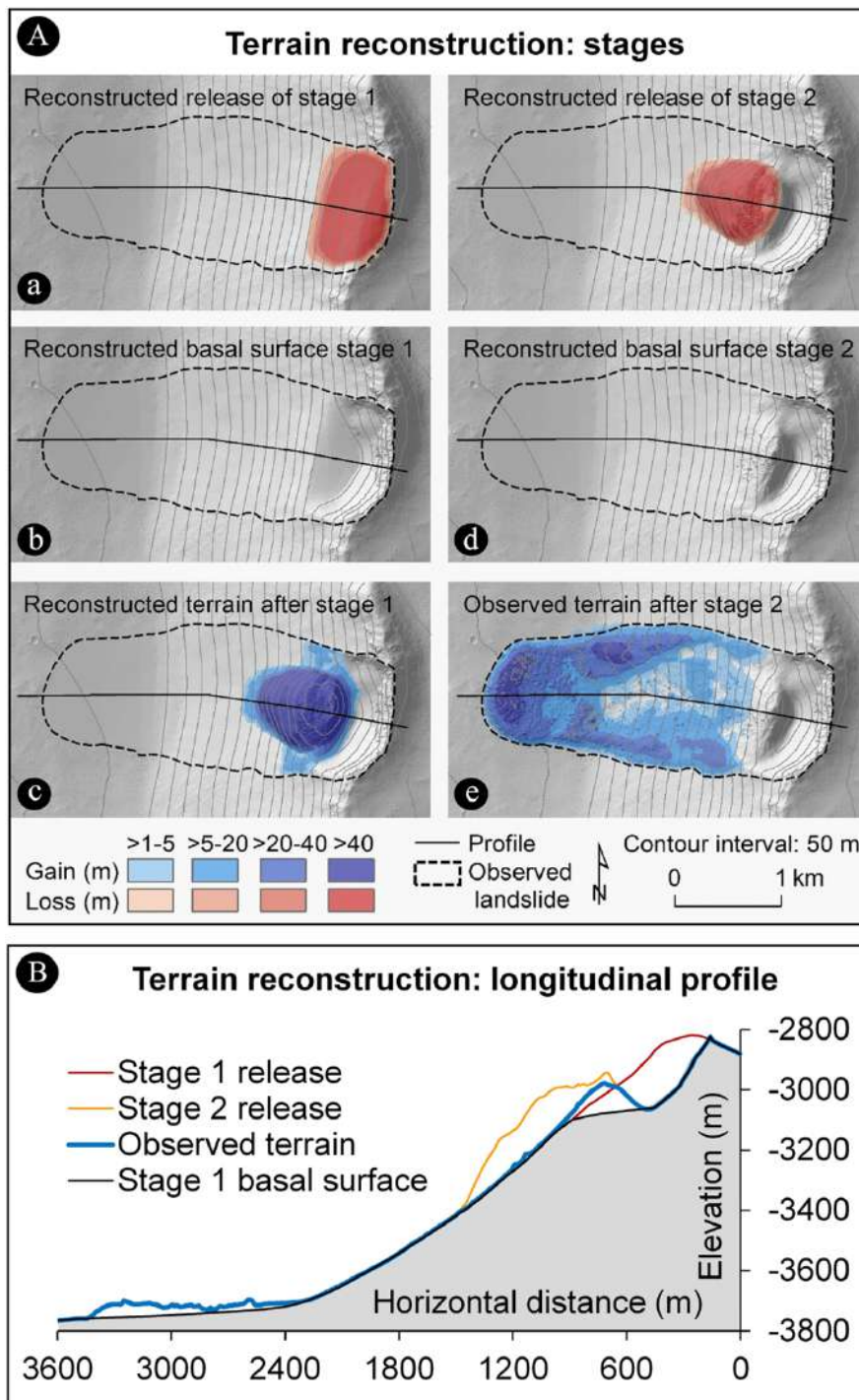
143 2.2 Terrain and event reconstruction

144 **By making use of the HiRISE 2 m-scale DEM (Fig. 1D) we first distinguish the circular-**
145 **shaped erosion zone where the Simud Vallis landslide detached, its longitudinal profile**
146 **down to the deposit zone, the sharp front, as well as the lateral levees. We then identify**
147 **the most likely sequence of mechanisms involved in the event. Based on the qualitative,**
148 **manual interpretation of visible terrain features, we suggest a two-stage model of the**
149 **event. Stage 1 represents a rotational slide, stopping and depositing still in the upper portion of**
150 **the slope (overlap of scarp and deposit). Stage 2 starts through the immediate or delayed**
151 **mobilization of part of the deposit, and propagates down the entire slope as a viscous flow-type**
152 **movement (see Discussion for a more detailed interpretation).**

153

154 On this basis, we reconstruct the terrain surfaces prior to the landslide and estimate the release
155 and deposition heights and volumes related to the different stages of the SV landslide. The
156 terrain reconstruction (Fig. 2A, B) is realised by manual construction of contour lines and their

157 subsequent interpolation, representing four situations (Fig. 2A a-e): (a) the terrain before stage
 158 1; (b) the basal surface of stage 1; (c) the terrain after stage 1, but before stage 2; (d) the basal
 159 basal surface of stage 2. The post-event situation (e) is represented by the existing DEM. The release
 160 height of stage 1 is derived as (a) – (b), the deposition height of stage 1 as (c) – (b), the release
 161 height of stage 2 as (d) – (c), and the deposition height of stage 2 as (e) – (d). Thereby, we build
 162 on the assumption that (i) the pre-event topography was characterized by a smooth and straight
 163 connection of the terrain elements around the landslide – **this assumption was implemented**
 164 **by connecting the corresponding contour lines on both sides of the landslide by a straight**
 165 **reconstructed contour lines, and interpolating those vector data to a DTM;** (ii) volumes
 166 were conserved during the landslide (no material has entered or left the system); (iii) during the
 167 flow-like stage 2 of the landslide, disintegration of the landslide mass generated a pore space
 168 of 20% of the volume.



170 Figure 2: Terrain reconstruction. A) Shaded relief map with release heights (loss) and deposition heights
171 (gain) of both stages. The different stages from (a) to (e) are detailed in the main text. B) Longitudinal
172 **profiles**. Note that horizontal and vertical scales are different.

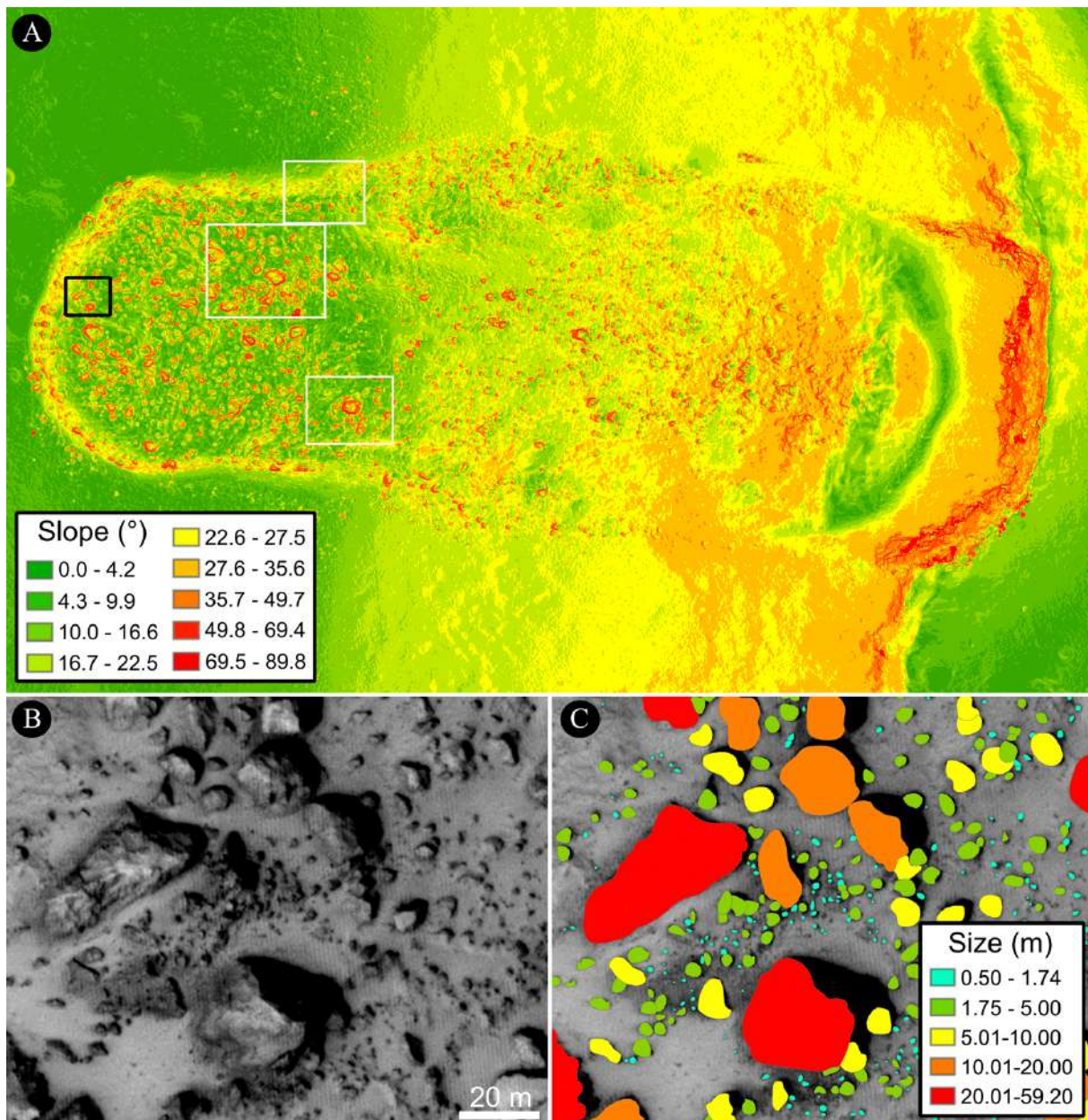
173 *2.3 Simulation of stage 2*

174 **Guimpier et al. (2021) modelled the SV landslide (called Chryse Chaos landslide, ChrC,**
175 **in that work) with the numerical simulation software *SHALTOP*. Results showed some**
176 **moderate correspondence between the observed and the simulated flow areas and**
177 **deposits, and the authors concluded that their results are not fully conclusive regarding**
178 **the mechanism. *SHALTOP* represents a comparatively simple flow model, where the**
179 **dynamics of the movement strongly depend on the basal friction angle as the dominant**
180 **material parameter, but where viscosity cannot be considered. *SHALTOP* has been**
181 **successfully applied to processes such as lahars (Peruzzetto et al., 2021), but its rheological**
182 **assumptions might be not fully suitable for the SV landslide with a presumably rather**
183 **viscous behaviour. Further, Guimpier et al. (2021) assumed a relatively confined release**
184 **area in the centre of the stage 1 deposit which, in our opinion, is not necessarily fully**
185 **supported by observations. Therefore, we re-evaluate the dynamics of the SV landslide**
186 **using the mass flow simulation framework *r.avaflow* (Mergili et al., 2017; Mergili and**
187 **Pudasaini, 2021), with our reconstructed release area and the assumption of a more**
188 **viscous flow, taking into account the gravitational acceleration on Mars of $3.72 \text{ m}\cdot\text{s}^{-2}$. Thereby,**
189 **the multi-stage flow model of Pudasaini and Mergili (2019) is applied. For the sake of**
190 **simplicity, we reduce the flow mass to one stage of a viscosity-dominated viscoplastic**
191 **material, which is assumed to consist of fragments of basaltic rock with a density of 3000**
192 **$\text{kg}\cdot\text{m}^{-3}$. The Pudasaini and Mergili (2019) model considers the grain density (i.e. the**
193 **density of the basalt in this case), and not the bulk density including the pore space. The**
194 **flow is assumed to stop as soon as its kinetic energy drops below 5% of the maximum kinetic**
195 **energy reached throughout the motion. Friction and viscosity parameters are optimized through**
196 **a step-wise iterative procedure, targeting at the best possible reproduction of the travel distance**
197 **as well as the extent and thickness of the observed landslide deposit. Simulations are run at a**
198 **raster cell size of 10 m.**

199

200 *2.4 Boulder identification and SFD*

201 We imported the map projected HIRISE images and the **2 m-scale** DEM into the ArcGIS
202 environment, which is commonly used for planetary landform analysis (Pajola et al., 2017c;
203 Mastropietro et al., 2020, Burke et al., 2021). Following the Burrough and McDonnel (1998)
204 methodology we computed the surface slope of the area (Fig. 3A), using it to identify both the
205 extent of the SV landslide as well as those of the largest boulders. We then visually identified
206 the landslide boulders based on their shape and slope boundary (Fig. 3A), defining them as
207 positive reliefs detectable through the presence of a nearby elongated shadow (Fig. 3B).
208 Afterwards, their outline was manually approximated by a polygon (Fig. 3C), extracting their
209 areas and returning the linear metric size as the diameter of a circle with the same area as the
210 polygon.



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Figure 3: A) Slope map of the SV landslide surface and boulders. The slope values (in degrees) are grouped in coloured bins. The black box outlines the extension of Fig. 3B and C, **while the black boxes outline the extension of Fig. SM2 A-F**. B) HiRISE uninterpreted closeup image showing the presence of multiple boulders. C) Same as B, but with the identified boulders grouped by size.

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The solar incidence angle and the height of the boulder **cause** uncertainties in the determination of the boulder shape because of the shadows. For example, at constant height, the smaller is the incidence angle, the smaller is the shadow. Other sources of error include the ability to reliably distinguish boulders from non-boulders features, such as small hills, mounds or portions of bedrock escarpments (Golombek et al., 2008). As previously done for HiRISE images, we decided to consider as statistically meaningful only those boulders with a diameter larger than 1.75 m (Pajola et al., 2017c; Mastropietro et al., 2020), i.e. 7 HiRISE pixels, hence reducing the possibility of feature misinterpretation. **We underline that some of the largest identified boulders are characterised by deep cracks and fractures (see the Discussion section of the manuscript) which could lead to the erroneous count of multiple boulders, instead of single ones. The continuity of the overall boulder shape means that we can determine if nearby**

229 chunks belong to the same feature, and so we decided to count the fractured boulders as
230 single ones, hence not affecting the resulting diameters (Supplementary Material Fig SM2
231 A-D). Another point to be considered is the fact that there are some boulders in the
232 landslide deposit that are mantled by dust. The correct size identification is then difficult,
233 because we can only measure the exposed part of the boulder, and we cannot infer what
234 lies underneath (Supplementary Material Fig SM2 E, F). Nevertheless, we believe that
235 given that the smallest sizes we identified (i.e. those boulders that should be most affected
236 by a dust blanket) are not used for the SFD analysis, we assume that the obtained results
237 are not significantly affected by this aspect. After the identification of the boulders, we
238 determined the area over which they are found to 3.03 km². This is done in order to obtain the
239 boulder density per unit area. We then derived the boulder SFD, using a log-log plot, where the
240 *x* axis is the boulder size in meters and the *y* axis is the cumulative number of boulders per km².
241 Afterwards, the data are fitted with multiple curves (power-law, exponential-law and Weibull)
242 that are commonly used for boulder SFD fitting studies on different planetary and minor bodies
243 surfaces (Golombek and Rapp, 1997; Michikami et al., 2008, Küppers et al., 2012; Pajola et al.,
244 2019; Schröder et al., 2021b), in order to evaluate which one better represents the data.

245

246 3.0 Results

247 The SV landslide forms a 3.4 km long feature that is located at 11°43'N, 322°54'E, in a
248 region that Tanaka et al. (2014) described as “Hto” unit, i.e. an Hesperian transition
249 outflow unit constituted by fluvial deposits of both Tiu and Simud Valles. A more detailed
250 geological analysis done by Pajola et al. (2016a) reported multiple evolutionary stages
251 (occurring from after Middle Noachian to Late Amazonian), with possible flow inversions
252 and ponding. In particular, the SV landslide is situated on the west-flank of a flat-topped
253 teardrop-shaped island (also called a mesa) in the middle of the Simud Vallis' floor, rising
254 up to ~950 m above it (Fig. 1B, D). The island's top has a Middle Noachian modelled age
255 and it is characterised by the presence of friable sediments, impact debris, as well as
256 volcanic material (Tanaka et al., 2014; Pajola et al., 2016a). As detailed in the
257 morphometric section of Guimpier et al. (2021), the erosion zone of the studied landslide
258 has a maximum length of 450 m, a maximum width of 1150 m and its slope ranges between
259 27° and 85° (Fig. 3A). In addition, the transport zone has a total length of 1700 m, a
260 maximum width of 1300 m and a mean slope of 23°. Finally, the deposit zone has a total
261 length of 1100 m and a maximum width of 950 m. It presents slope values between ~2° to
262 < 10°, while its front scarp slope ranges between 18° and 29°.

263 The studied landslide and its circular-shaped erosion zone are not the only ones present
264 on the flanks of the teardrop-shaped island. Indeed, two other similar eroded shapes are
265 present in the north-western flanks of the Simud mesa (Fig. 1C, white circles).
266 Nevertheless, the studied SV landslide is the only one with clearly recognisable boulders
267 across the deposit without significant aeolian mantling and the appearance of the boulders
268 is fresh.

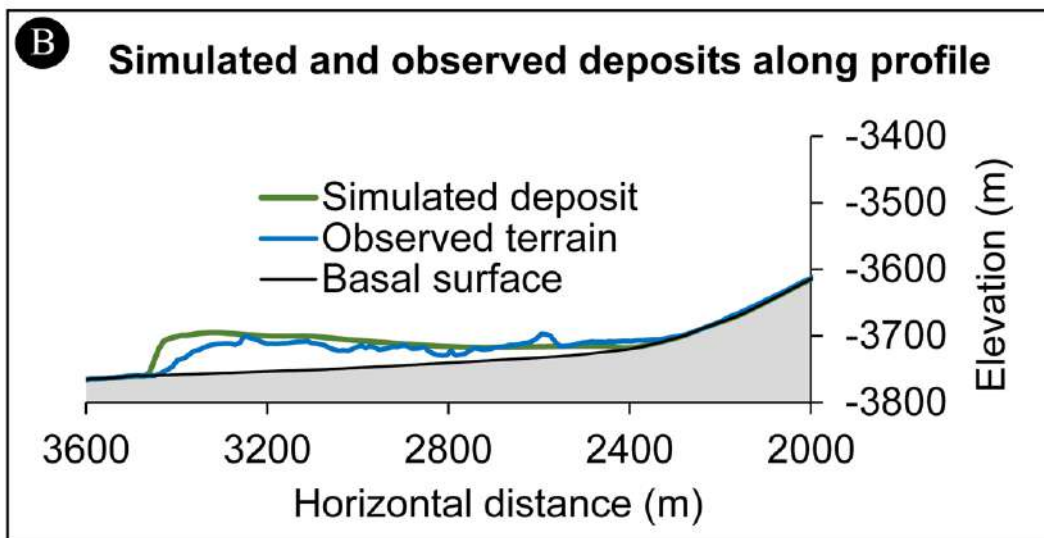
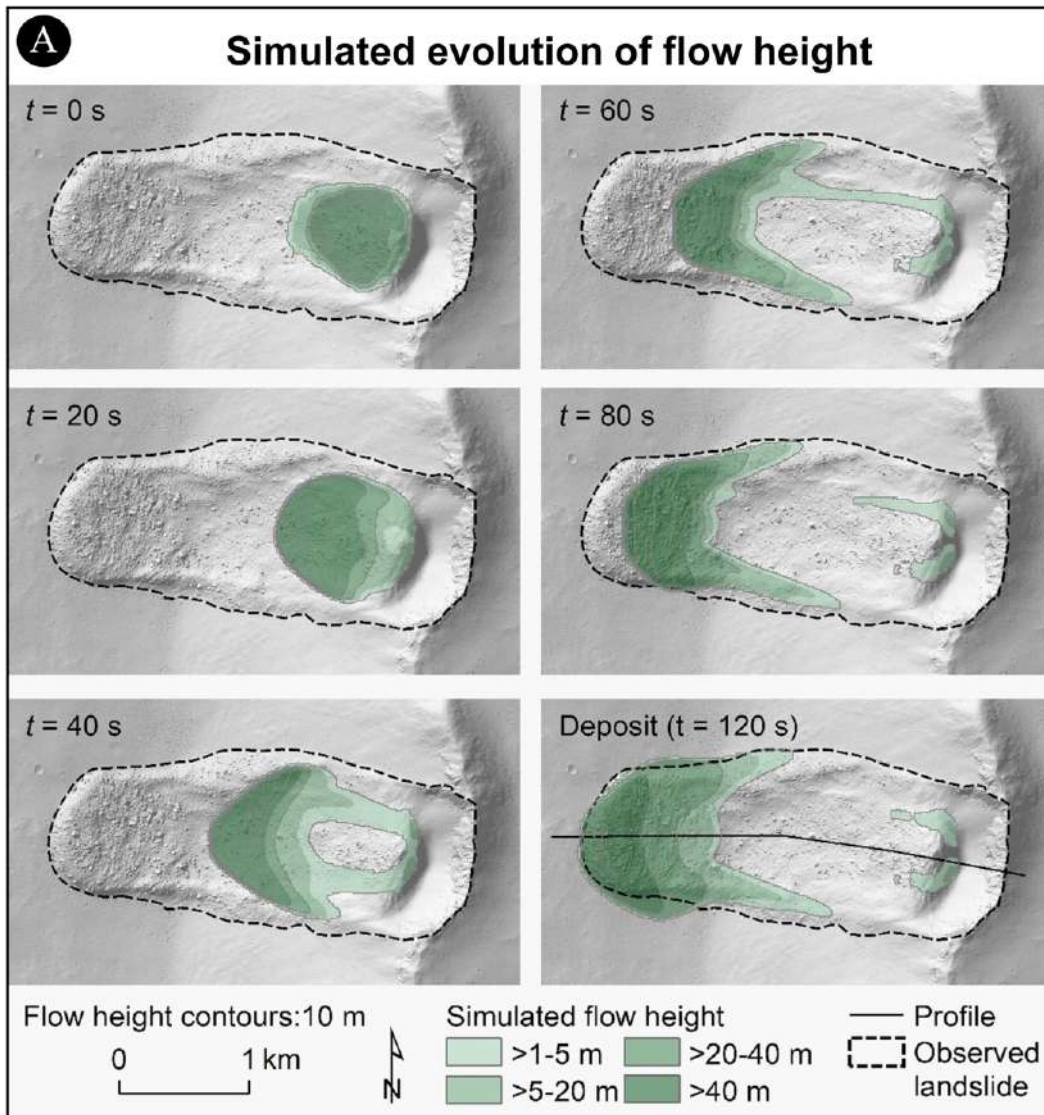
269 3.1 Terrain and event reconstruction

270 Fig. 2A and B shows the reconstructed release and deposition heights of both stages of the SV
271 landslide. For stage 1, we suggest a rotational slide with a total volume of $54.0 \cdot 10^6$ m³. We
272 assume a deforming, but not disintegrating moving mass, characterized by identical release and
273 deposition volumes. **Deformation of the sliding mass was certainly necessary to conform to**
274 **the terrain (a completely rigid block would not have been able to move in the way shown**

275 **by available evidence), and is also indicated by the shape of the remaining intermediate**
276 **deposit. If there had been immediate disintegration (i.e. the transformation from a slide-**
277 **dominated movement to a flow-dominated movement), it would have been expected that**
278 **the mass would have immediately moved down to the base of the slope (as it did in stage**
279 **2), and would not have formed the clearly visible intermediate deposit.** $13.4 \cdot 10^6 \text{ m}^3$ of the
280 stage 1 deposit remain in place after stage 2 (shown in the post-event DEM). This means that a
281 stage 2 release volume of $40.6 \cdot 10^6 \text{ m}^3$ has to be imposed in order to achieve the stage 1
282 deposition of $54.0 \cdot 10^6 \text{ m}^3$. The post-event DEM reveals a deposited volume of $48.7 \cdot 10^6 \text{ m}^3$
283 for stage 2. This means a volume increase of 20% due to the generation of pore space during
284 the flow, corresponding to the target value defined in the methods section. This confirms the
285 overall plausibility of the reconstruction, even though some details (the exact shape of the stage
286 1 deposition / stage 2 release mass in particular) remain uncertain.

287 *3.2 Simulation of stage 2*

288 The stage 2 flow is numerically reconstructed with the *r.avaflow* software. Empirical
289 optimization of the key model parameters results in an internal friction angle of 40° , a basal
290 friction angle of 15° , a kinematic viscosity of $162 \text{ m}^2\text{s}^{-1}$, and a yield strength of 40 Pa. Fig. 4A
291 illustrates the simulated evolution of the flow height. The general patterns of the flow, including
292 the formation of a sharp front and lateral levees, is plausibly reproduced by the simulation. The
293 height of the deposit is slightly overestimated in the frontal part, particularly when considering
294 that the model result does not include the assumed 20% pore space of the observed deposit (Fig.
295 4B). The most notable deviation between simulation results and observation consists in the
296 minor southward turn of the mass in the relatively flat deposition area, where the terrain very
297 slightly drops southward by approximately 0.8° . The observed deposit does not show any sign
298 of turning southward. This phenomenon is considered in more detail in the Discussion section.



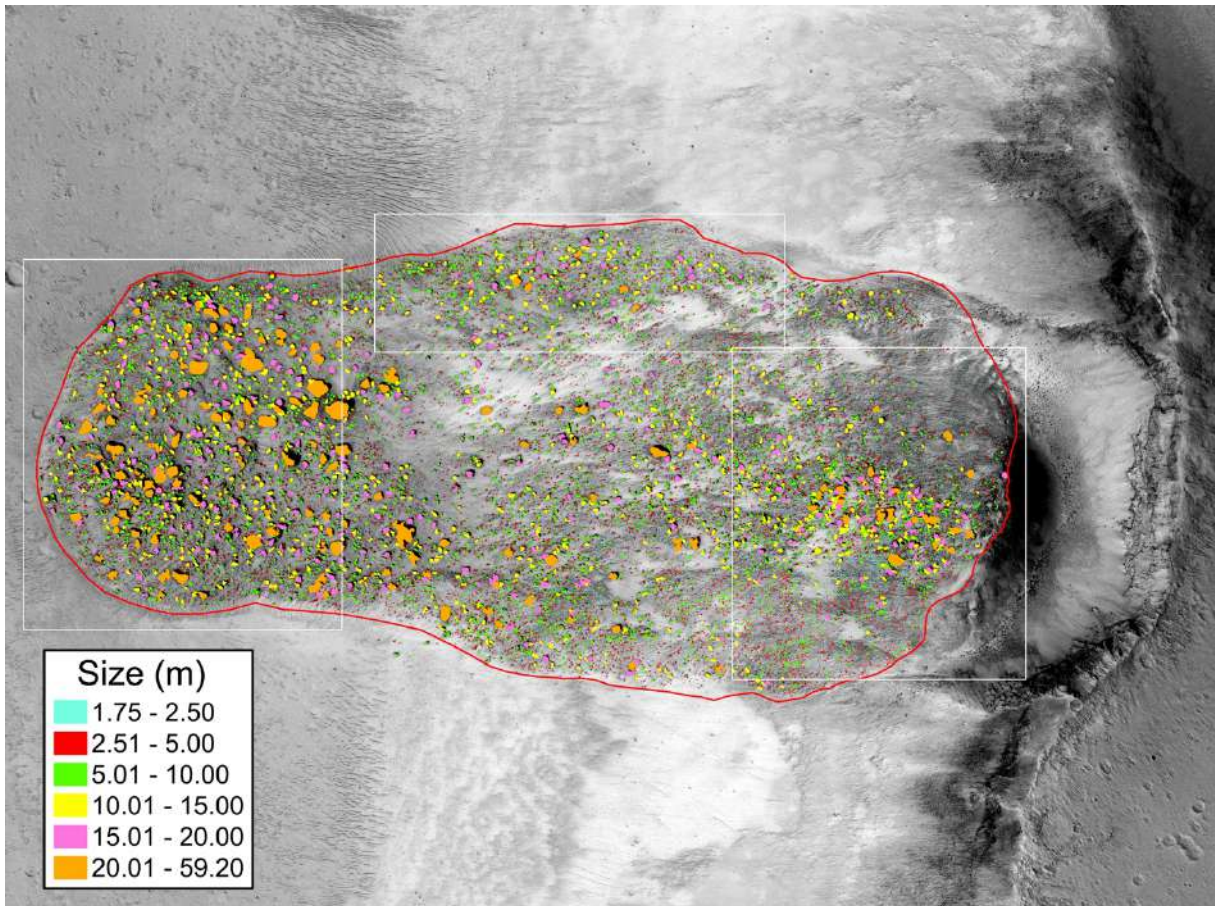
299

300 Figure 4: Simulated evolution of the flow height of the second stage of the SV Landslide. A) Shaded
 301 relief map with flow heights at different points in time. B) Longitudinal profile showing the comparison
 302 of the observed and the simulated stage 2 deposits. Note that horizontal and vertical scales are different.
 303 Simulated flow heights and deposit exclude pore space in both A and B.

304

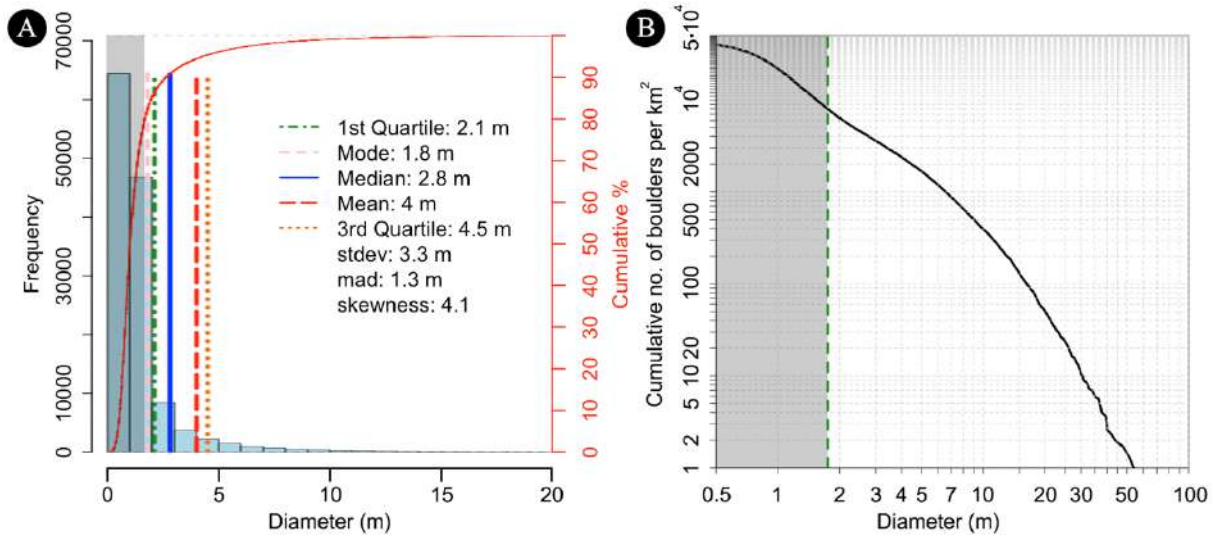
305 *3.3 Boulder SFD*

306 We manually identified 130 597 boulders at the surface of the SV landslide (Fig. 5), with a
307 minimum size ~ 0.5 m and a maximum size reaching 59.2 m. As explained in the Methodology
308 section all boulders < 1.75 m are excluded from the statistics, in order to avoid possible feature
309 misinterpretation (Fig. 6A). The total number of identified boulders ≥ 1.75 m is 24 073, **with**
310 **an areal density** of 7945 km^{-2} for a size of 1.75 m (Fig. 6B). The density value decreases to
311 4595 km^{-2} at a size of 2.5 m, 1674 km^{-2} at 5 m and 391 km^{-2} for 10 m size boulders.



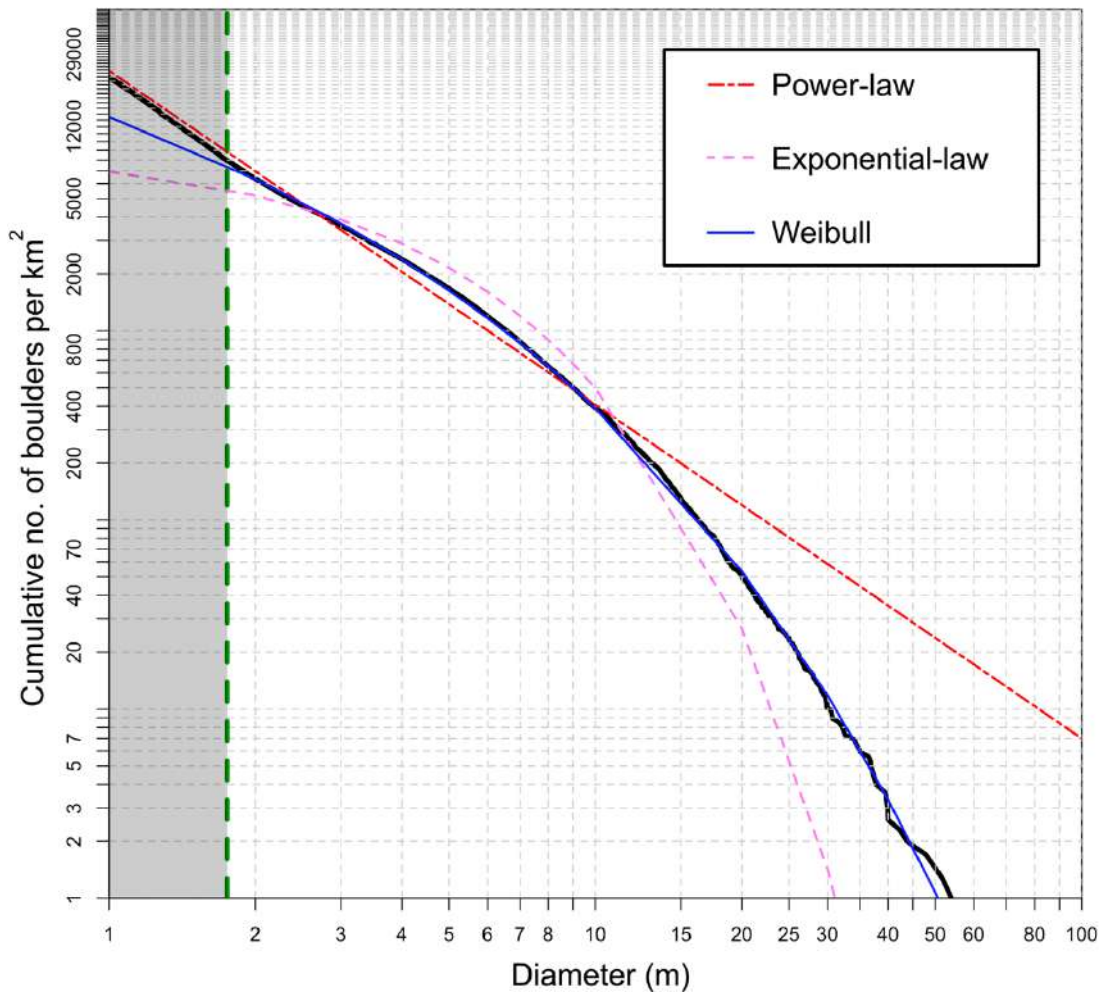
312

313 Figure 5: The spatial distribution of all boulders ≥ 1.75 m identified in the SV landslide. The boulder
314 sizes (in metres) are grouped in coloured bins. The red polyline identifies the area (3.03 km^2) where we
315 counted the boulders, while the white boxes outline the close-up views that are presented in
316 Supplementary Material Fig. SM3.



317

318 Figure 6: A) Frequency histogram of all boulders identified in the red area outlined in Fig. 5. The grey
 319 shadowed area shows the boulders that have been identified, but their size is < 1.75 m. The main
 320 statistical properties of the right-skewed distribution are computed **only for size values ≥ 1.75 m**. B)
 321 Log-log plot showing the cumulative number of boulders per km². As for A, the grey shadowed area
 322 indicates all boulders < 1.75 m.



323

324 Figure 7: Cumulative number of boulders per km² with the three different fitting curves computed in the
 325 1.75 m – 59.2 m size range. As for Fig. 6, the grey shadowed area shows boulders < 1.75 m, not
 326 considered by the multiple fits.

327 We then fitted three different curves on the cumulative number of boulders per km²: a power-
328 law, an exponential-law and a Weibull distribution, for the size range 1.75 m – 59.2 m. For each
329 case we used the corresponding R² value to evaluate the quality of the fit.

330 In particular, since in our computation we solely include boulders larger than a certain size in
331 the fit, we could use the Weibull left-truncated distribution with the cumulative form of Wingo
332 (1989):

$$333 \quad N(> d) = N \exp[-\alpha(d_i^\beta + d_{min}^\beta)],$$

334 where N is the number of boulders larger than d_{min}, while α and β are the Weibull parameters
335 estimated from the boulder sizes $d_i > d_{min}$. The resulting best fits are presented in Fig. 7. The
336 power-law fitting curve is characterised by an index of -1.77 and an R² value of 0.97, while the
337 exponential-law curve has an exponent of -0.29 and a R² of 0.93. On the contrary, the Weibull
338 best fitting curve which returns a R² value of 0.99 is characterised by an α value of 1.09 and a
339 β value of 0.57.

340

341 **4.0 Discussion**

342 The identification and mapping of landslides on Mars uses the same visual criteria adopted for
343 terrestrial landslides when interpreting aerial photographs or satellite images. These criteria
344 include shape, size, tone, texture, object pattern, site topography, and setting² (Guzzetti et al.,
345 2012). Mass movements are classified based on their morphological similarity with terrestrial
346 landslide types (Cruden and Varnes, 1996; Hungr et al., 2014). The main types of mass
347 movements recognized on Mars are: (i) slides (RS), including rock slides and deep-seated
348 slides, (ii) complex/compound slides (CL), (iii) rock avalanches (RA), (iv) debris flows (DF),
349 including debris avalanches and shallow debris slides, and (v) rock glacier-like features (RG)
350 (Brunetti et al., 2014).

351 Based on the visual criteria, the SV landslide can be interpreted as a complex landslide (CL)
352 involving both a rotational slide and a flowing component. Moreover, the landslide deposit
353 exhibits a set of trenches (**white arrows in Fig. 1D**) that align parallel to the local direction of
354 motion. This could be indicative of a likely rapid-to fast-moving flow mechanism for the
355 emplacement of failed materials (Pierson and Costa, 1987).

356 The post-event terrain, the derived reconstruction of the pre-event terrain, as well as the release
357 mass support the hypothesis of a rotational slide (slump) during stage 1. There is little evidence
358 to conclude or even to speculate on the velocity of the stage 1 event, which might have been
359 anything from extremely slow to extremely rapid, according to the classification of Cruden and
360 Varnes (1996). However, it appears very likely that a major fraction of the stage 1 landslide
361 deposited on the steep slope beneath the circular-shaped landslide scar (Fig. 2A-B). This opens
362 the question on the mechanisms and timing of the transition from stage 1 to stage 2.

² The pattern of an object is its spatial arrangement in a repeated or typical order or form. The site topography is the position with reference to its surroundings. The setting refers to regional and local characteristics (lithological, geological, morphological, climatic, etc.) in relation to the surroundings. Site topography and setting are used to infer rock type and structure, attitude of bedding planes, and the presence of faults and other tectonic or structural features (Guzzetti et al., 2012).

363 In principle, the evidence supports two idealised scenarios, or a combination of both:

- 364 I. The material mobilised in the stage 1 rotational landslide entered the steep downslope
365 area. It was continuously remobilised through the stage 2 flow process, continuing as
366 long as material was supplied from upslope, starting either immediately or after a
367 threshold of load was reached;
- 368 II. The stage 1 landslide material deposited on the steep slope remained in place until a
369 threshold was reached, and then moved at once, forming the stage 2 flow.

370 The steep-fronted, largely unstructured deposition lobe of the stage 2 flow supports scenario II.
371 However, there seem to be at least two generations of levees: smaller linear features are located
372 within the area bounded by large levees (see Fig. 3A and Fig. 5). This might indicate a multi-
373 stage process characterized by an unknown number of surges, with at least one smaller flow
374 surge after the main one. Each surge might have been triggered by the exceedance of a load
375 threshold exerted by stage 1 material pushed onto the steep slope, making a slowly moving
376 stage 1 movement more likely.

377 We now consider the potential velocity of stage 2 of the SV Landslide:

- 378 i. Movement as an extremely rapid avalanching flow. The runout plane is very gently
379 inclined towards SW, as shown by the contour lines in Fig. 2A. Any flow which is not
380 moving extremely rapidly would be expected to turn toward the direction of steepest
381 inclination when moving from the W-facing steep slope onto the SW-facing runout
382 plane. Stage 2 of the SV Landslide, in contrast, maintained its flow direction. These
383 findings support the hypothesis of an extremely rapid, quite suddenly stopping flow,
384 with too much inertia to quickly change flow direction;
- 385 ii. Movement as a slow-rapid velocity flow. This hypothesis is supported by the fact that
386 the lobate shape and steep frontal escarpments are consistent with a highly viscous,
387 **“honey-like” type** of flow, such as slow-moving features on Earth, for example earth
388 flows or rock glaciers (Fig. 8A). Some other similar structures on Earth, however, are
389 related to processes of unknown velocity (Fig. 4B) or to presumably extremely rapid
390 flow-type landslides (Fig. 4C). **Consequently, the available evidence does not allow**
391 **a conclusive interpretation in terms of landslide velocity. The simulation results**
392 **indicate the plausibility of an extremely rapid flow, but do not exclude the**
393 **possibility of a slower movement.**



395

396 **Fig. 8.** Selected features of comparable size and pattern on Earth. A) Active rock glaciers in the Rushan
 397 Range (Pamir, Tajikistan). B) Ancient feature in the Vakhani Valley (Hindukush, Afghanistan), probably
 398 representing **the remnants of a landslide**. C) Tongue of a presumably extremely rapid debris avalanche
 399 near Ishkashim (Hindukush, Afghanistan). Photos: M. Mergili, 14 and 15 August 2009.

400

401 The best-fit *r.avaflow* simulation (Fig. 4), characterized by a high degree of empirical adequacy,
 402 presumes an extremely rapid stage 2 flow, which is suddenly released from the stage 1 deposit.
 403 All simulation attempts towards progressive release and/or a flow of lower velocity failed to
 404 reproduce the deposition patterns, as the simulated flows were clearly turning towards the
 405 direction of steepest slope in the deposition area – a behaviour even slightly visible in the best-
 406 fit result for an extremely rapid flow, clearly pointing towards the hypothesis II (Fig. 4A). **The**
 407 **values of the flow parameters (basal and internal friction, viscosity, yield strength) leading**
 408 **to the best correspondence of the *r.avaflow* simulation results with the observations (area,**
 409 **patterns of deposition height) are the outcome of an empirical optimisation procedure.**
 410 **We underline that these optimised parameter values should not be considered as**
 411 **physically correct material characteristics, as the optimisation outcome may be affected**
 412 **by possible equifinality issues, as well as by simplifications of the model assumptions.**
 413 **Hence, simulations on other Martian landslides with comparable characteristics will be**
 414 **necessary to draw conclusions on the general validity of that set of parameter values.**
 415 **Nevertheless, if we compare our results with those obtained using a granular flow model**
 416 **presented by Guimpier et al. (2021), it is possible to notice that our simulated dynamical**
 417 **evolution returns a final deposit that is more similar to the observation in terms of the**
 418 **longitudinal profile, as well as for what concerns the formation of the lateral levees and the**
 419 **frontal shape. However, we note that *r.avaflow* is primarily designed for extremely rapid flows,**

420 and, to our knowledge, there is also no other software available to properly simulate slow-
421 rapidly moving flows, impeding a more rigorous numerical analysis of the corresponding
422 scenarios.

423 As it is possible to see from Fig. 5, the identified boulders are not homogeneously distributed
424 along the landslide. As predicted by landslide particle size segregations (Crosta et al., 2007),
425 where large particles generally move to the front and to the top surface, while the smaller
426 particles accumulate at the bottom of the flow, the SV landslide's higher boulder spatial density
427 is located at the forefront, where the biggest boulders with sizes larger than 20 m are also
428 present. The lateral levees of the landslide are also two areas where the boulder density
429 increases (Supplementary Material Fig. SM3), while the central part of the landslide is generally
430 characterised by a depletion of boulders. Another location where the density of boulders
431 increases is the rear side of the landslide, close to the reconstructed area where stage 2 might
432 have started. It is likely that such boulders are the underneath remnant of the landslide mass
433 that detached during stage 1.

434

435 From the boulder SFD analysis we find that the power-law curve adequately fits the SFD at the
436 smallest sizes in the range 1.75 - 3.2 m, with a power-law index of -1.77, nevertheless it
437 underestimates the values from 3 to 10 m, as well as severely overestimates the data > 10 m
438 (Fig. 7). In contrast, the exponential-law fit underestimates values between < 3 m and > 10 m,
439 while overestimating the sizes that are found in the middle. For this reason, this curve can hardly
440 be considered as a good fitting one for the observed data. The Weibull curve, instead, show a
441 very good fit to the data in the whole size range considered (Fig. 7).

442

443 From a formative perspective, the power-law SFD would indicate a single-event fragmentation
444 (for example during impact cratering) that leads to a branching tree of cracks that have a fractal
445 character (Turcotte et al., 1997, Schröder et al., 2021b). Whereas, the Weibull distribution is
446 thought to result from sequential fragmentation (Brown & Wohletz 1995) and it is largely used
447 in fracture and fragmentation theory (Grady and Kipp, 1987; Brown and Wohletz, 1995;
448 Turcotte, 1997; McSaveney, 2002). In addition, the Weibull distribution (Weibull, 1951) is
449 often used to describe the particle distribution that is derived from grinding experiments (Rosin
450 and Rammler, 1933). The good fit between the boulder SFD and the Weibull curve may
451 therefore suggest that while the SV landslide formed and was moving downward as a single
452 event, the rocky constituents may have been ground in a sequential way (McSaveney, 2002),
453 rather than in a sudden, single-event fragmentation (as, for example, boulders that are formed
454 during impacts), hence resulting in the distribution that we see today.

455 **Some of the largest identified boulders show exposed deep cracks and fractures dividing**
456 **them into separate chunks (Supplementary Material Fig. SM2 A-D). As shown by Eppes**
457 **et al. (2015), simple diurnal insolation can drive subcritical fracture growth in Martian**
458 **rocks, that can afterwards result in polygonal pattern of cracking and hence a**
459 **disintegration of the boulder into pieces. Moreover, de Hass et al. (2013) estimated that**
460 **the boulder local breakdown rate due to the presence of (metastable) liquid water during**
461 **the last glacial period of Mars is of 3.5 m/Ma. So, despite the young age of the SV landslide**
462 **($\sim 4.5 \pm 4$ Ma, Guimpier et al., 2021) favouring the preservation of the original boulders**
463 **SFD, it could be affected by secondary weathering processes.**

464 **We do not think that solar-induced thermal stress can explain all the observed fractures**
465 **because we do not observe the mentioned developed polygonal cracking in the boulders**
466 **(at least at the 0.25 m spatial scale) and the cracks we identify are randomly oriented,**
467 **which should not be the case for solar-induced cracks. Moreover, the boulders in the SV**
468 **deposit look particularly fresh with sharp shapes, and only a small proportion (a few tens)**
469 **of the total number is affected by fractures, hence pointing to recent boulder formation**

470 from the parent bedrock. Concerning the local boulder breakdown rate, de Haas et al.
471 (2013) mentioned that this value is site specific and it was identified on a small alluvial fan
472 system in eastern Promethei Terra, which is a different morphological setting to the one
473 studied here. In addition, it is not clear if metastable liquid water enhancing weathering
474 rates was ever present on the SV landslide deposit given its equatorward latitude.
475 For all such mentioned reasons, we maintain that the derived SFD reflects the original
476 one, nevertheless, we cannot rule out that an in-situ form of weathering is actually
477 occurring in the SV deposit, at least on the biggest boulders.

478
479 We highlight that both the simulation reported in this paper, as well as the boulder SFD
480 study do not analyse the triggering causes of the landslide, which is a difficult matter to
481 identify and unambiguously prove (Crosta et al., 2014; Kumar et al., 2019; Bishop et al.,
482 2021). Indeed, as presented by Guimpier et al. (2021), both seismic shaking from a nearby
483 meteorite impact, or from a crustal marsquake could have been triggers for this landslide,
484 but other processes such as those related to thermal stress cannot be ruled out, either
485 (Tesson et al., 2020). However, one could hypothesize as predisposing factor the presence
486 of a structural weakness in the landslide area, as evidenced by the local irregularity of the
487 mesa edge. Indeed, Guimpier et al. (2021) suggested that the raised topography near to
488 the landslide scarp could be the remnants of a rim of an ancient impact crater that formed
489 before the incision of the Simud Vallis outflow channel. If this is the case, it could point
490 out at a location of pre-existing weakness. Moreover, the occurrence of localised darker-
491 than-the-surroundings colour in close proximity to the circular eroded area could hint at
492 a different mineralogical composition, too. Whether this suggests the presence of hydrated
493 materials or not is a matter of debate, since this area is affected by the presence of dust
494 and no unambiguous hydrated mineral signatures have been identified here (Pajola et al.,
495 2016a). But if so, this could support the hypothesis of a reduced local strength. Eventually,
496 structural weakness in the mesa could also be hypothesized due to the deep excavated
497 trenches visible at the back of the landslide scarp (Fig. 1C).

498 499 **Conclusions**

500 We have studied a young ($\sim 4.5 \pm 4$ Ma), 3.4 km long landslide located on the floor of Simud
501 Vallis on Mars that was previously analysed by Guimpier et al. (2021). This previous work
502 **included** a *SHALTOP* numerical modelling of the landslide movement, but acknowledged that
503 the results obtained were not conclusive since the simulated deposits differed from the
504 **observed** landform appearance.

505 Based on the visual criteria, the Simud Vallis landslide can be interpreted as a complex landslide
506 involving both a rotational slide (stage 1) and a flow (stage 2). Using a **2 m-scale** HiRISE DEM
507 we have reconstructed the initial terrain surface, allowing us to estimate the release and
508 deposition heights and volumes related to the different stages of the landslide. For stage 1, we
509 suggest a rotational slide with a total volume of $54.0 \cdot 10^6$ m³. Most of this initial volume was
510 then involved in the stage 2 flow, whereas $13.4 \cdot 10^6$ m³ of it remained perched in the source
511 area. The post-event DEM revealed a stage 2 deposition volume of $48.7 \cdot 10^6$ m³, indicating a
512 volume increase of 20% with respect to the stage 2 release value, due to the generation of pore
513 space during the flow. This confirms the overall plausibility of the reconstruction, even though
514 the exact shape of the stage 1 deposition and the stage 2 release mass remain uncertain.

515 The stage 2 flow has been numerically reconstructed with the *r.avaflow* software. The general
516 patterns of the flow, including the formation of a steep frontal scarp and lateral levees, is
517 reproduced by the simulation. Nevertheless, we highlight that the height of the modelled deposit
518 is slightly overestimated in the frontal part, since our model result does not include the assumed
519 20% pore space of the observed deposit. Moreover, when compared to the **observed** deposit,
520 our simulation results show a very slight southward turn of the mass in the relatively flat
521 deposition area. Our best-fit simulation suggests an extremely rapid stage 2 flow, which is
522 suddenly released from the stage 1 deposit. This is supported by the fact that the landslide
523 deposit exhibits a set of trenches that align parallel to the local direction of motion, indicative
524 of a likely rapid-to fast-moving flow mechanism for the emplacement of failed materials. **The**
525 **simulated impact area based on *r.avaflow* clearly shows a higher degree of correspondence**
526 **to the observed impact area of the Simud Vallis landslide than the simulated impact area**
527 **based on *SHALTOP* (Guimpier et al., 2021). We attribute this improved model**
528 **performance to a more appropriate definition of the release mass, as well as to the**
529 **assumption of a viscous flow, which is most probably more realistic than the assumption**
530 **of a purely frictional flow, as it was applied by Guimpier et al. (2021) based on the scope**
531 **of the *SHALTOP* software.**

532 By using two 0.25 m-scale HiRISE images we have manually identified and counted > 130 000
533 boulders located along the landslide. As predicted by landslide particle size segregations, the
534 identified boulders are not homogeneously distributed along the landslide. The highest spatial
535 density of boulders is located at the front of the deposit, where the biggest boulders with sizes
536 larger than 20 m are also present. The boulder density increases also inside the lateral levees of
537 the landslide as well as where the remnants of the mass that detached during stage 1 is present.
538 After deriving the boulder size-frequency distribution in the size range 1.75-59.2 m (the total
539 number of boulders ≥ 1.75 m being 24 073), we have identified that the best fitting curve is the
540 Weibull distribution which results from sequential fragmentation and it is often used to describe
541 the particle distribution derived from grinding experiments. This suggests that while the Simud
542 Vallis landslide formed and was moving downslope, it could have ground the rocky constituents
543 in a sequential way, rather than in a sudden, single-event fragmentation.

544 **The analyses done in this paper did not aim to identify the triggering causes of the**
545 **landslide. Nevertheless, we could hypothesize as the landslide predisposing factor the**
546 **presence of structural weaknesses in the collapsed area, as i) the irregularity of the mesa**
547 **edge that can hint at an ancient impact crater that formed before the incision of the Simud**
548 **Vallis outflow channel, ii) the occurrence of the deep excavated trenches visible at the**
549 **back of the landslide scarp, as well as iii) the possible presence of a different local**
550 **mineralogy, all favouring a reduced local strength.**

551
552 **Eventually, we underline that the characterisation of the boulder SFD, coupled with the**
553 **modelling of the landslide movement is a novel approach that can be used for future**
554 **analyses of mass movements and boulder fragmentation that occurred in the Late**
555 **Amazonian period of Mars.**

556

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565

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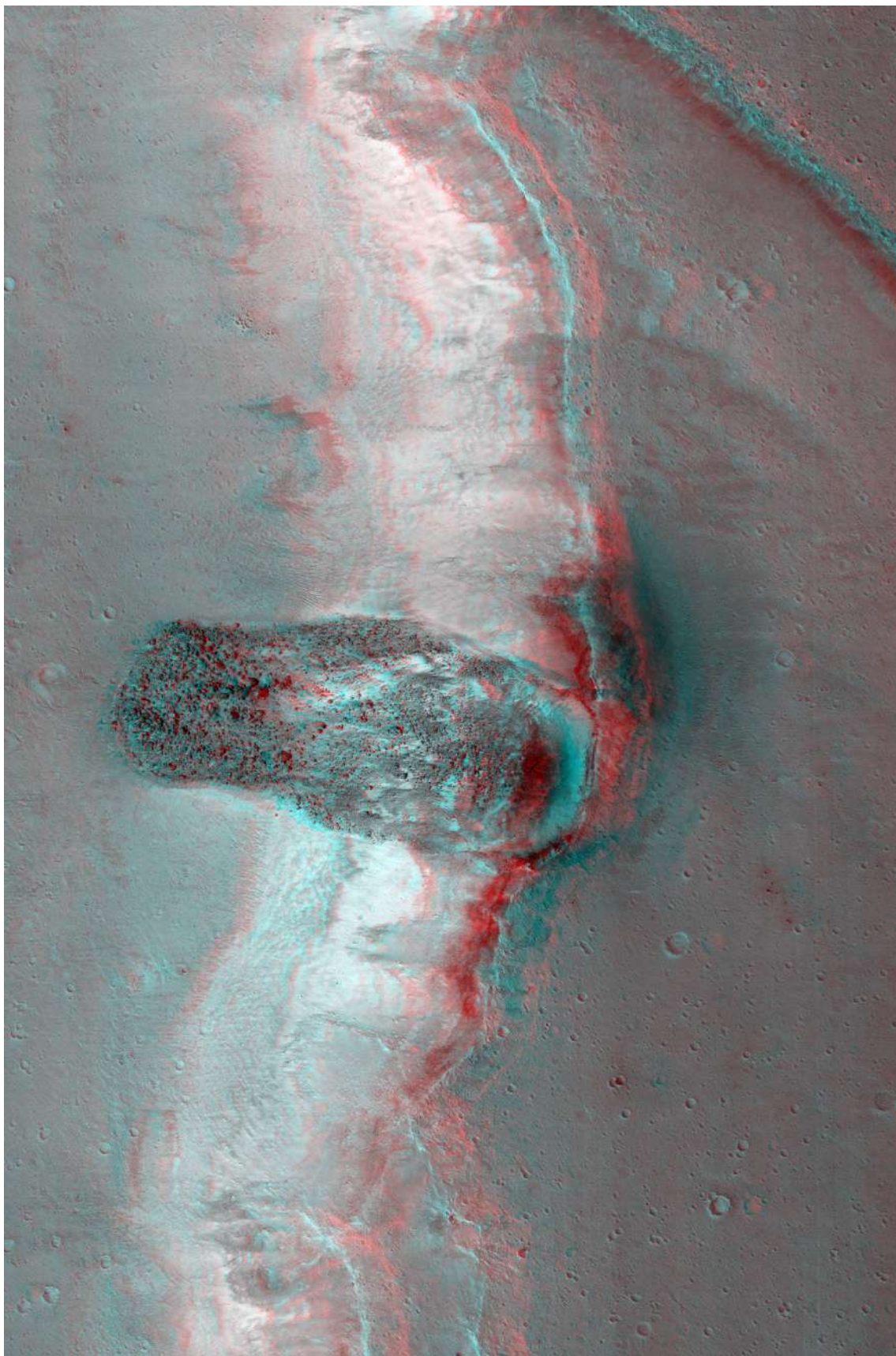
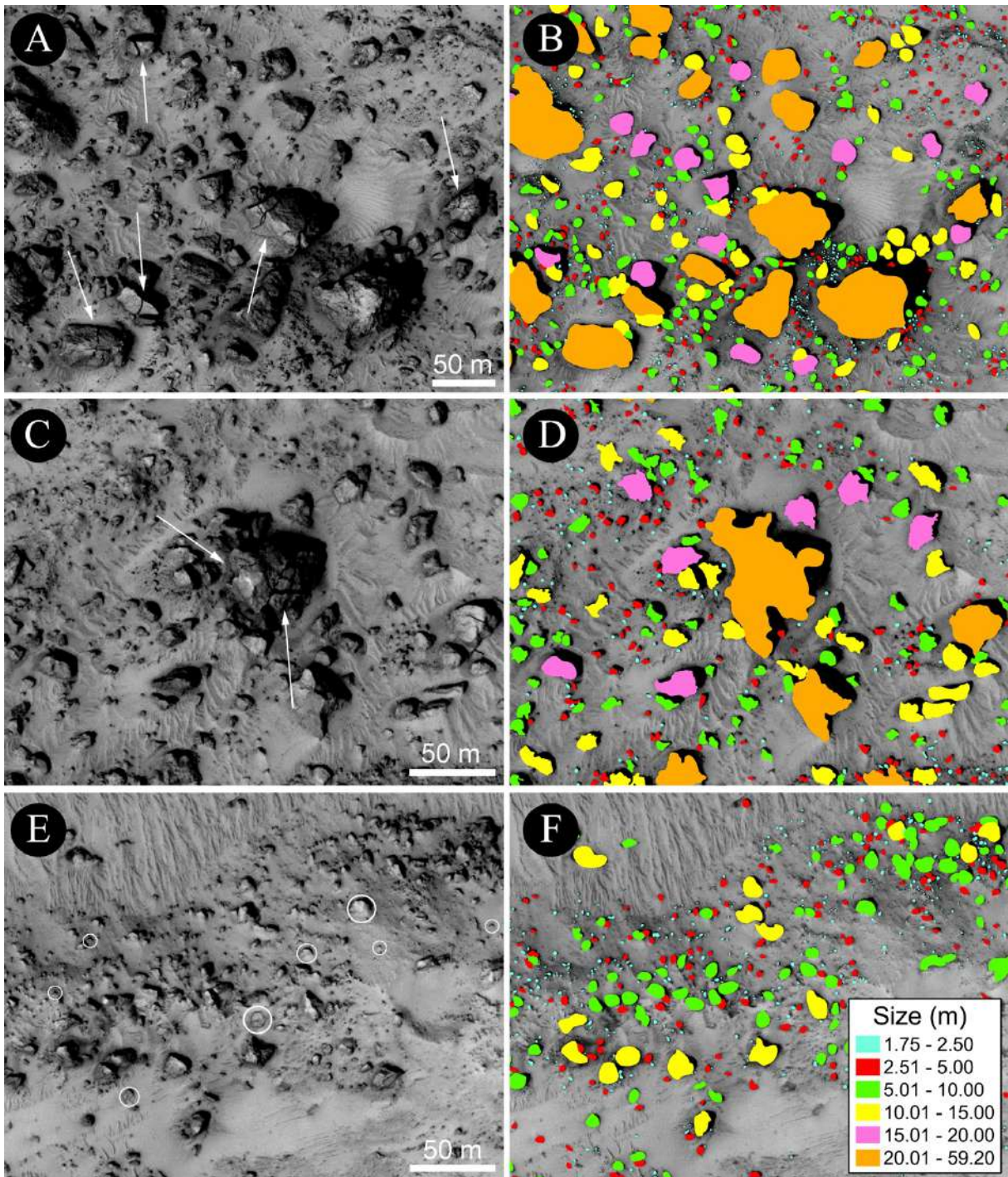


Figure SM1: The anaglyph obtained with the two HiRISE images presented in Table 1.



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834 Figure SM2: A-D) Close-up figures showing the fractured boulders (indicated with white arrows)
 835 mentioned in the main text. As it is possible to see all chunks belonging to the same fractured boulder
 836 are identified as a single boulder. E-F) The boulders (identified with white circles) that are partially
 837 covered by dust.

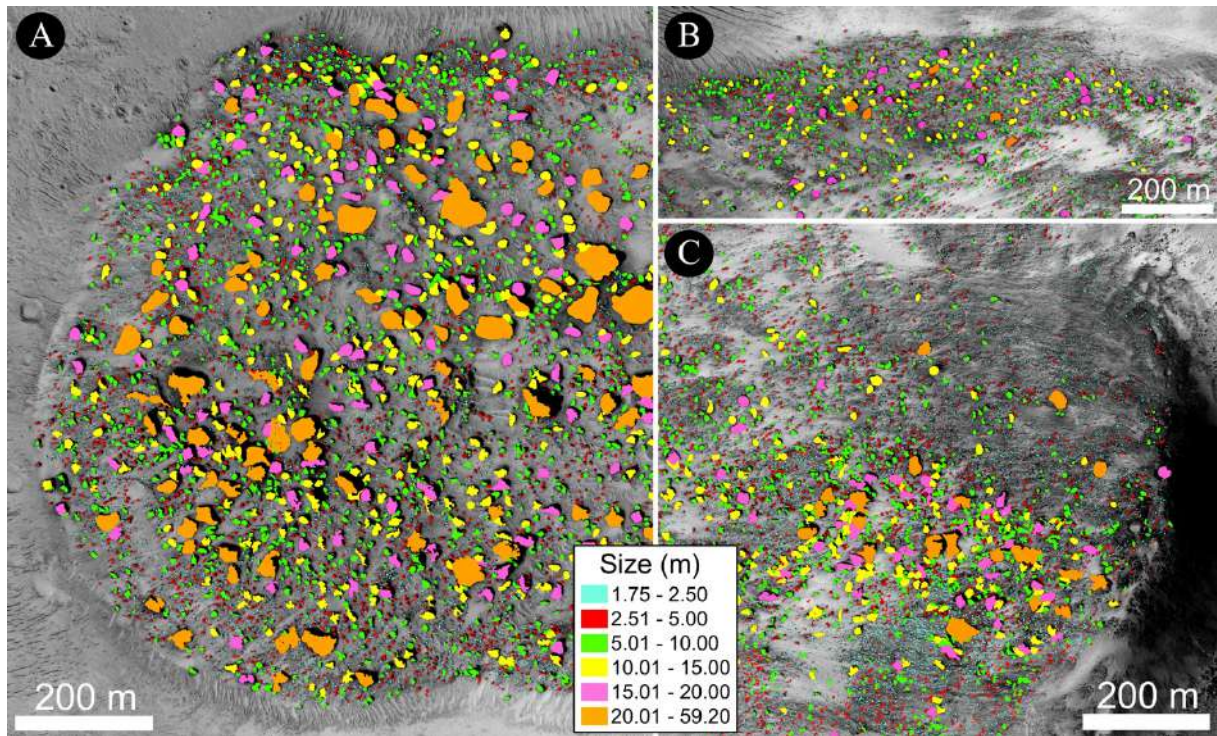
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844 Figure SM3: Close-up figures showing the boulders ≥ 1.75 m presented in Fig. 5. A) Frontal part of the
 845 SV landslide. B) Upper lateral levee of the landslide. C) Rear side of the SV landslide.