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1 **Dynamics of recent landslides (<20 My) on Mars: Insights from high-resolution**
2 **topography on Earth and Mars and numerical modelling**

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19 **Keywords** : Geomorphology – Digital Elevation Model – Landslides – Modelling – SHALTOP

20 **Abstract**

21 Landslides are common features found on steep slopes on Mars and the role of water in their
22 formation is an open question. Our study focuses on three young martian landslides whose mechanism
23 of formation is unknown and knowing their formation mechanism could give us key information on
24 recent martian climate and/or tectonics. They are less than 5 km long, and formed during the Late
25 Amazonian Epoch, with an age < 20 Ma when Mars is thought to have had a hyperarid climate. To
26 better understand the dynamics and formation mechanism of these landslides, we combine two
27 approaches: geomorphic comparison between martian and terrestrial landslides using remote sensing
28 data from the High Resolution Imaging Science Experiment (HiRISE) and the Colour and Stereo Surface
29 Imaging System (CaSSIS), and numerical modelling using a dry granular flow dynamical model. Our
30 geomorphic analysis revealed two contrasting morphologies suggesting differing dynamics and
31 formation mechanisms. Two of the three martian landslides resemble terrestrial rockslides, while the
32 third is more akin to terrestrial mudslides. The numerical modelling, although not fully conclusive,
33 broadly supports our interpretations from the morphological observations. We suggest that the two
34 landslides resembling terrestrial rockslides could have been triggered by shaking by meteorite impact
35 or marsquakes in the absence of water. On the contrary, we suggest liquid water (originating from
36 ground-ice melted by geothermal heat flux) may have been involved in the initiation of the landslide
37 resembling a terrestrial mudslide. Our results show the value of using morphological comparison
38 between martian and terrestrial landslides combined with numerical modelling to inform the
39 hypotheses of landslide-formation on Mars where in situ analysis is not usually possible.

40 **1. Introduction**

41 Large landslides were first observed on Mars in 1972 by the Mariner 9 probe, in Valles Marineris
42 (Sharp, 1973). This region is characterised by a succession of steep-sided canyons, trending East-West
43 over ~4000 km (Quantin et al., 2004a; Lucas et al., 2011; Brunetti et al., 2014; Watkins et al., 2015),
44 with more than 1400 landslides (Crosta et al., 2018) formed between Hesperian (3.5 Ga) and Late
45 Amazonian (50 Ma) (Quantin et al., 2004b). Several studies have investigated the morphology of the
46 large landslides in Valles Marineris (Lucchitta, 1979; Quantin et al., 2004a; Soukhovitskaya and Manga,
47 2006; Brunetti et al., 2014; Crosta et al., 2018), which are characterised by scarps up to several
48 kilometres wide and kilometre-deep, broad fan-shaped deposits, often wider than the scars from
49 which they originated. The role that water may have played in these landslides is the main
50 preoccupation of these previous works and is important to understand because these landslides have
51 occurred throughout Hesperian to Late Amazonian epochs so can provide information on Mars’
52 climate through time. Mass movements can also give information on the tectonic history of a planetary
53 body (Quantin et al., 2004b). The majority of previous studies of martian landslides have examined
54 landslides with volumes greater than 10^{10} m^3 , which is larger than landslides most commonly found on
55 Earth. This lack of a direct terrestrial analogue is one of the reasons that the triggering and dynamics
56 of these large landslides is still a subject of active research and the role of water and/or active tectonics
57 is unclear.

58 To our knowledge, no studies have specifically focussed on understanding ‘small’ martian
59 landslides with a volume less than 10^{10} m^3 , which have a similar scale to landslides that can be found
60 on Earth. These common terrestrial landslides are well-studied and their formation mechanisms are
61 better understood than that of their larger counterparts. This provides an opportunity to perform a
62 comparative morphological study between terrestrial analogues and martian landslides without the

63 need for scaling. We selected three relatively fresh, recent martian landslides (with potentially
64 contrasting formation mechanisms), with the least influence of secondary processes on their surfaces
65 (e.g., impact craters, aeolian features) and topographic data available, in order to increase the
66 reliability and robustness of the comparative study. By identifying similar morphologies in the martian
67 landslides and in terrestrial analogues, whose formation process is known, we can infer the processes
68 that may have been at work on Mars.

69 In addition to this comparative morphological study, we use the thin-layer numerical code SHALTOP
70 to simulate the landslide dynamics, assuming it is a dry granular flow. In spite of their simplifying
71 assumptions and the uncertainty on initial and boundary conditions (see Section 2.3, and Delannay et
72 al., 2017), thin-layer numerical models have previously been successful in reproducing the runout and
73 approximate deposit morphology for a wide range of landslides on Earth and Mars. Using seismic data
74 to reconstruct the dynamics of some terrestrial landslides, it was shown that thin-layer models can
75 also reproduce these dynamics (Moretti et al., 2012, 2020; Levy et al., 2015; Yamada et al., 2016, 2018).

76 We therefore employ a double approach, using both morphological and numerical methods, to better
77 constrain the mechanism of formation of these ‘small’ martian landslides and to understand their
78 dynamics and hence, the potential role of liquid water and/or active tectonics.

79 First, in Section 2, we describe the data and the methods used to carry out this study, including the
80 morphological analysis, age-estimation using crater size-frequency analysis and the numerical model
81 used to carry out the simulations. In Section 3, we present the results from the morphological analysis,
82 age estimation and numerical simulations. In Section 4 we first compare our results with those for
83 other martian landslides presented in the literature, then discuss the potential emplacement
84 mechanisms of the martian landslides and finally we assess the likelihood of the different hypotheses
85 that could explain the formation of these three martian landslides.

86 **2. Methodology**

87 In this section, we elaborate the data and methods used to carry out the geomorphological analyses
88 of the martian and terrestrial landslides. We also describe the crater counting method used to estimate
89 the age of formation of martian landslides and then the method used to perform the numerical
90 modelling.

91 **2.1. Geomorphological analysis**

92 In order to analyse the geomorphology of the martian and terrestrial landslides, we made
93 measurements using the tools provided by the ArcGIS software. We first detail the data that we used
94 during this study, then the analysis methods used for the martian and terrestrial landslides.

95 **2.1.1. Datasets for martian landslides**

96 We analysed the martian landslides at two scales: 1:1,000,000 scale for the geographical and geological
97 context, and 1:4,000 scale to identify the key geomorphological structures. The analysis was performed
98 in sinusoidal projection respectively centred at 325 °E, 322 °E and 78 °E for the Capri Chasma, Chryse
99 Chaos and Nilosyrtris Mensae landslides.

100 At 1:1,000,000 we used the published geological and structural maps of Mars (Tanaka et al., 2014) and
101 the Mars Orbiter Laser Altimeter (MOLA) Digital Elevation Model (DEM), with a resolution of

102 463 m/pixel (Smith et al., 2001). We used images from ConTeXt imager (CTX, Malin et al., 2007)
103 onboard the Mars Reconnaissance Orbiter (MRO) with a resolution of ~6 m/pixel and from the Colour
104 and Stereo Surface Imaging System (CaSSIS, Thomas et al., 2017) onboard the Trace Gas Orbiter (TGO)
105 with a resolution of 4 m/pixel.

106 For the 1: 4,000 scale analysis we used images from MRO High Resolution Imaging Science Experiment
107 (HiRISE, McEwen et al., 2007) with a resolution of 25 to 50 cm/pixel. We also used DEMs with a
108 resolution of 2 m/pixel produced from HiRISE stereo observations using the Ames Stereo Pipeline
109 (Moratto et al., 2010). These were vertically controlled to ESA's Mars Express High Resolution Stereo
110 Camera (HRSC, Neukum and Jaumann, 2004) publically available DEMs.

111 Finally, near the Nilosyrtris Mensae landslide we made additional analyses of surface composition using
112 Compact Reconnaissance Imaging Spectrometer (CRISM, Murchie et al., 2007) with a spatial resolution
113 of ~19 m/pixel (*see supplementary material, S1- Spectral analysis*).

114 **2.1.2. Datasets for terrestrial landslides**

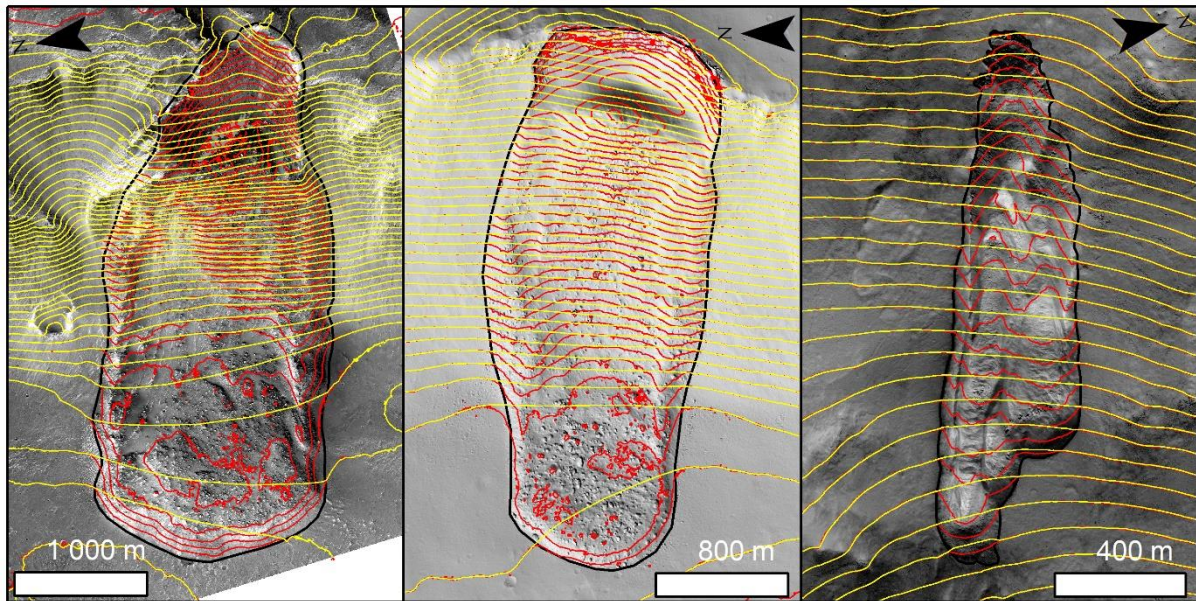
115 We used three terrestrial analogues, comprising landslides located near Abisko in Sweden, near Seattle
116 in the USA and near Hól mavík in Iceland. We chose these three landslides because of their similar
117 morphology to the martian landslides studied here. The dynamics and formation of terrestrial
118 landslides is better understood than those on Mars, hence comparison between martian and terrestrial
119 landslides is intended to provide additional information on the dynamics of the formation of the
120 martian landslides of our study. We therefore used a similar resolution of data as for the martian
121 landslides.

122 For the landslide located near Abisko in Sweden at 68°12' N, 19°2' E, a 2 m/pixel DEM derived from
123 airborne laser altimetry was provided by the Swedish Land Survey, Geographical Sweden Data (GSD)-
124 Elevation Data. For the landslide located near Seattle and Mount Rainier in Washington State in the
125 USA at 46°59' N, 121°40' W, a 3 m/pixel DEM was used as provided by the Washington Lidar Portal
126 (<https://lidarportal.dnr.wa.gov>). For the landslide located near Hól mavík in Iceland (65°42' N,
127 21°42' W), we used archived aerial images from the Land Survey of Iceland (LMI) at 30 cm/pixel and
128 control points derived from the ArcticDEM (Porter et al., 2018) to reconstruct the topography using
129 the commercial software AgiSoft Photoscan multiview photogrammetry resulting in a final DEM with
130 a ground sampling of 60 cm/pixel.

131 **2.1.3. General analysis**

132 **1:1M-scale analysis.** At this scale we identified the presence or absence of similar landslides in the
133 area surrounding our studied landslides. This scale also allowed us to identify faults and impact craters
134 surrounding our landslides. Using the geological map of Tanaka et al. (2014) we were able to identify
135 the units in which our landslides are located.

136 **1:4k-scale image analysis.** At this scale HiRISE images were used. We determined the size of the
137 boulders (we use the term 'block' to refer to clasts > 4 m in length; Blair and McPherson (1999)) on the
138 landslide deposits by measuring the long and short axes in planview and did not include the shadow.
139 We made visual observations on the texture of the landslide deposits at the metre-length scale (e.g.,
140 rough, smooth, ripples) and made a visual assessment of the density of blocks on each landslide. To
141 determine the height of the small structures, such as ridges, we extracted perpendicular topographic
142 cross sections from the DEMs.



143 *Figure 1 – Contour lines in red are derived from the original HiRISE DEM of each martian landslides of this study and contour*
 144 *lines in yellow are estimated pre-landslide contours. All these contour lines have an interval of 25 meters. (a) Capri Chasma*
 145 *landslide, CCh, HiRISE: ESP_035831_1760; (b) Chryse Chaos landslide, ChrC, HiRISE: PSP_005701_1920; (c) Nilosyrtris Mensae*
 146 *landslide, NM, HiRISE: ESP_027480_2075. Credits NASA/JPL/UofA.*

147 **Topographic reconstruction.** In order to understand the spatial distribution of erosion and deposition,
 148 we generated a thickness map for each of the landslides. We had to make a reconstruction of the initial
 149 topography before the landslide occurred because no data are available from before the landslides
 150 formed. We took the difference between the reconstructed topography and the observed topography
 151 to generate the thickness map.

152 We estimated the initial topography by adapting the procedure described in Conway and Balme (2014)
 153 and de Haas et al. (2015) and took as guide the existing topography (Lucas and Mangeney, 2007; Lucas
 154 et al., 2011, 2014; Conway and Balme, 2014; Coquin et al., 2019). We used the DEMs to derive elevation
 155 contours for our landslides at 25 m intervals. We digitised the landslide boundary using slope and
 156 slope-aspect maps (derived from the DEM), as well as orthoimages. Slopes were calculated in each
 157 DEM cell as the average slope in a 3 × 3 cells neighbourhood around the central cell. The normal to the
 158 plane returns the aspect of the central cell. We then manually drew pre-landslide ‘reconstructed’
 159 contours within the landslide boundary. We used the contours outside the landslide to guide our
 160 placement of the reconstructed contours and connected them manually to the two intersections of
 161 each contour line with the landslide boundary with a smooth curve, as shown in *Figure 1*.

162 The boundary line of the landslide was converted to point features at 2 m intervals and attributed with
 163 the elevation values of the DEM. The reconstructed contours in turn were converted to point features
 164 at 2 m intervals, attributed with the contour elevation value. These point features and associated
 165 elevations were then gridded into a ‘reconstructed DEM’ using the ArcGIS Natural Neighbour
 166 interpolation algorithm. The difference between this reconstructed DEM and the original DEM results
 167 in a thickness map, with positive values indicating areas of deposition and negative values indicating
 168 erosion. This reconstruction was also used to estimate the volume of deposition zone of the landslide.
 169 We summed the pixels with positive values and multiplied by the pixel size in the deposition zone to
 170 get an estimation of the deposition volume.

171 **Landslide morphometrics.** We used the thickness map to divide the landslides into erosion, transport
172 and deposition “zones”. The “erosion zone” is where the thickness map has predominantly negative
173 values across the whole width of the landslide, the “transport zone” is where both positive and
174 negative values are found across the width and where predominantly positive values are found across
175 the whole width, this corresponds to the “deposit zone”. We measured the maximum width of each
176 of the zones perpendicular to the general slope of the surrounding escarpment and their length parallel
177 to the general slope. We calculated the aspect ratio of the erosion zone by taking the ratio of its width
178 to its length. We calculated the area by delimited the landslide boundary line by summing the number
179 of DEM pixels within it and multiplying by their area.

180 **Topographic profile analysis:** We extracted two types of topographic profiles to derive additional
181 morphometric characteristics: i) longitudinal profiles along the full length of the landslide and ii) profiles
182 at selected key positions within and outside the landslide.

183 We used the longitudinal profiles to analyse the variations in elevation and slope angle (computed with
184 a 20 point running mean) within the landslide. We placed a topographic long profile along the centre
185 of the landslide extending from ~60 m above the scar to ~60 m below the toe of the deposits. We used
186 the same topographic long profile to calculate Heim’s ratio which is the ratio between the total drop
187 height H and the runout distance $\Delta L'$, both measured from the top of the scar to the toe of the deposit.
188 This ratio has been previously calculated for martian and terrestrial landslides in the literature (Legros,
189 2002; Quantin et al., 2004a; Lucas et al., 2014; Brunetti et al., 2014; Crosta et al., 2018) and is used to
190 assess the mobility of a mass movement. The lower the ratio is, the longer the runout of the landslide
191 is compared to the drop height and the greater its mobility.

192 Three topographic profiles were extracted for the levees, the steepest parts of the erosion zone, the
193 transport zone and the front scarp (see *Figure S2* in the supplementary material for the precise position
194 of these topographic profiles for each landslide). The cross-section along levees are used to compute
195 the height and width of each levee, as well as the slope between the base and the crest of the levee.

196 For the steepest part of the erosion zone, we placed three profile lines in the centre and at the edges
197 of the erosion zone perpendicular to the contour lines. For the transport zone three profiles were
198 placed perpendicular to the contour lines at horizontal intervals of between 200 and 300 metres. For
199 the front scarp one profile was placed in the central part and two near the edges of the front scarp.
200 These profiles were used to calculate the average slope of these features.

201 We extracted one elevation profile ~300 m outside the boundary landslide (*Figure S2*), and
202 perpendicular to the contour lines, in order to estimate the slope angles of the terrain before the
203 landslide occurred and the slope angle of the deposit zone. This profile extended from the top to the
204 bottom of three zones defined above. Here we take only one profile, because the further profiles are
205 placed from the landslide, the less likely they are to be representative of the pre-landslide surface. For
206 each zone we calculated the slopes, with the methodology as described above.

207 For each zone of the landslide, we report the means and ranges of the slope values calculated from
208 the profiles.

209 **2.2. Age estimations of martian landslides**

210 In order to estimate when the martian landslides formed, we used the crater size-frequency
211 distribution method to obtain model ages. We used ArcGIS CraterTools (Kneissl et al., 2011) and

212 Craterstats2 (Michael and Neukum, 2010; Michael et al., 2012). This method exploits models which
213 describe how the bolide production function varies over time, enabling the size-frequency distribution
214 of impact craters present on a planetary surface to be linked to a modelled surface age (Michael and
215 Neukum, 2010; Michael et al., 2012; Michael, 2013; Michael et al., 2016).

216 In the deposition zone (described in section 2.1), we digitized craters to compute their distribution in
217 size. The deposit zone area and the size-frequency distribution of the superposed craters are then used
218 as inputs for Craterstats2 (Michael and Neukum, 2010; Michael et al. 2012). Due to the small surface
219 area of the landslides studied here and the low number of impact craters on their surface, there are
220 large margins of error in these estimations (e.g., Warner et al., 2015).

221 **2.3. Numerical modelling**

222 Numerical modelling has already been successfully used to better understand landslides on Mars and
223 compare them with terrestrial ones (Soukhovitskaya and Manga, 2006; Lucas and Mangeney, 2007;
224 Mangold et al., 2010; Lucas et al., 2011, 2014). Following previous work (Lucas and Mangeney, 2007;
225 Lucas et al., 2011, 2014; Brunet et al., 2017), we model our martian landslides with the numerical code
226 SHALTOP and compare the simulation results with the observed deposits.

227 **2.3.1. Model input preparation**

228 In SHALTOP simulations, the topography on which the modelled landslide propagates is the
229 reconstructed topography (*see section 2.1.3*), but with only the deposits removed, and the erosion
230 zone remains unchanged – we call this the “scar topography” to differentiate it from the reconstructed
231 topography described above. The initial mass is given by the difference between the reconstructed
232 topography and the scar topography (Fig.2). Reconstruction of the topography and of the initial mass
233 is challenging and can significantly affect the results (Lucas et al., 2011; Moretti et al., 2015; Peruzzetto
234 et al., 2019).

235 We smoothed the reconstructed topography twice to remove small artefacts related to the
236 reconstruction method (e.g., subtle elevation steps at the landslide boundary). In the first pass, we
237 calculated the mean elevation value within a 10-pixel square moving window. In the second, we
238 calculated the mean value within a 6-pixel radius circular moving window. In order to avoid
239 overestimating the travel distances produced in the model, we added back roughness after smoothing,
240 as follows. First we chose a typical sample area of the terrain outside the landslide of the same size as
241 the landslide itself. We smoothed the elevation data within the terrain sample by averaging the
242 elevation values within a moving window 10×10 pixels in size and applied this procedure three times.
243 We differenced this smoothed sample with the original to obtain a DEM with only the meter-scale
244 roughness (because of the DEM vertical resolution), which was then added to the landslide zone.

245 **2.3.2. Model evaluation and analysis criteria**

246 The first criterion used to compare our simulations to observations is the runout distance. We also
247 considered two secondary criteria: the final position of the centre of mass and the deposit thickness
248 map (*see section 2.1.3*), to assess the simulation results. To calculate the centre of mass, we first
249 extracted the positive values of the thickness map corresponding to the deposits of each landslide. A
250 grid of deposit thicknesses is a direct output of the model so no extraction is required. Then we convert
251 the deposition thickness map into points where each point corresponds to a pixel and contains its

252 thickness value. The x and y coordinates of the centre of mass is calculated according to the average x
 253 and y coordinates of the centroids of each point weighted by the thickness value.

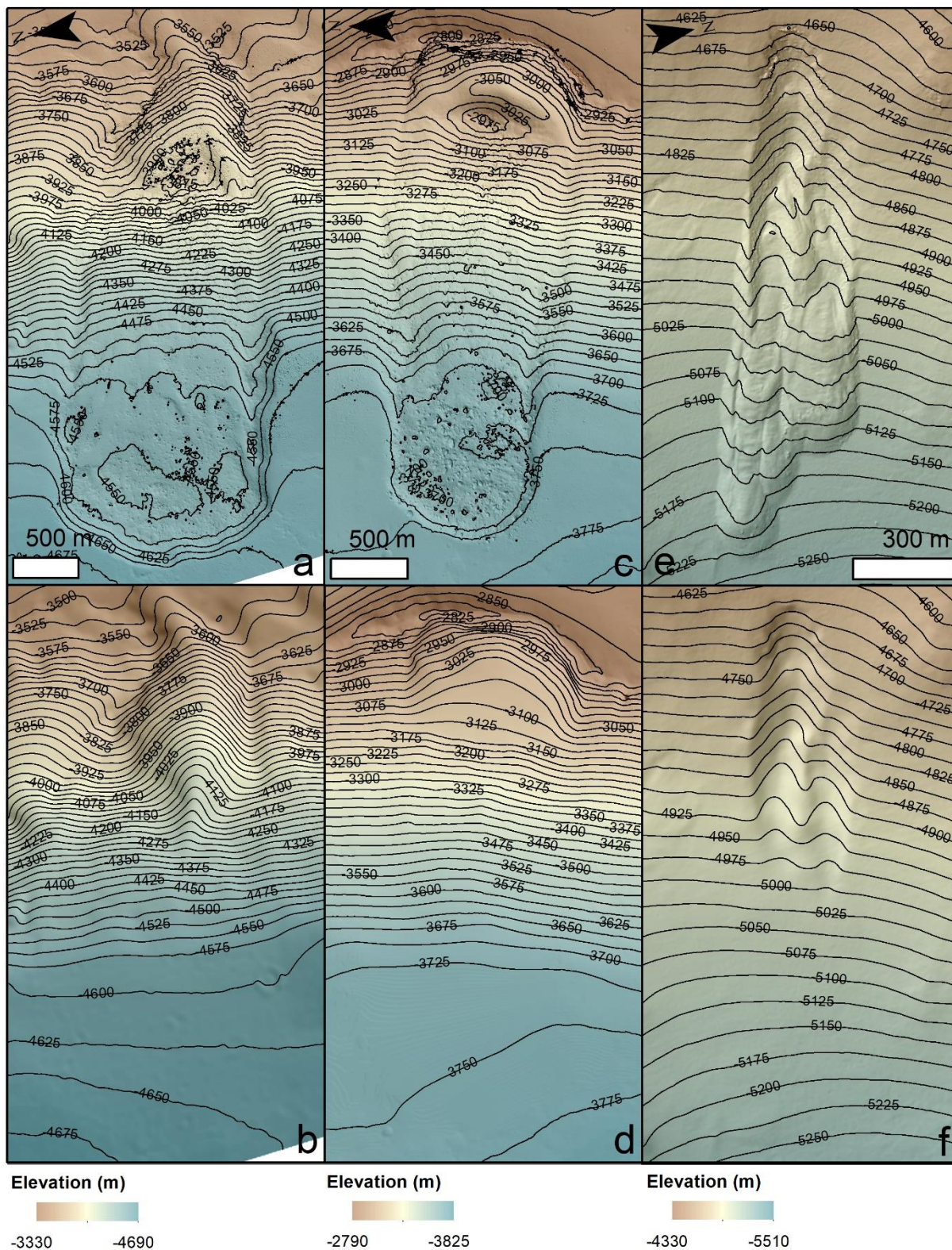


Figure 2 - A 3D shaded-relief rendering of the Capri Chasma (CCh, a & b); Chryse Chaos (ChrC, c & d) and Nilosyrtis Mensae (NM, e & f) with their respective topography. (a, c, e) Is the present-day topography. (b, d, f) Is the topographic input for SHALTOP, where the estimated pre-landslide surface is combined with the present-day erosion zone.

258

2.3.3. SHALTOP description

259 SHALTOP is a numerical model that simulates homogeneous flows propagating on complex topography
 260 using the thin-layer approximation (that is, simplifications to the governing equations that can be used
 261 when the flow thickness is small in comparison to its lateral extent) (Bouchut et al., 2003; Bouchut and
 262 Westdickenberg, 2004; Mangeney-Castelnau et al., 2005; Mangeney et al., 2007). In the thin-layer
 263 approximation, the flow is described by its thickness h in the direction normal to the topography and
 264 by its depth-averaged velocity u . In SHALTOP, energy is dissipated through basal friction. In contrast to
 265 most of the depth-averaged landslide models, SHALTOP accounts for the curvature tensor of the
 266 topography with all its components. The resulting topography effects can significantly change the
 267 runout and/or flow velocity, in particular for rapid granular flows over complex topographies
 268 (Peruzzetto et al., 2020). SHALTOP has already been used to successfully model terrestrial landslides
 269 (Lucas et al., 2007; Favreau et al., 2010; Moretti et al., 2012, 2015, 2020; Brunet et al., 2017; Yamada
 270 et al., 2016, 2018; Peruzzetto et al., 2019), as well as martian landslides and recent gullies (Mangold et
 271 al., 2010; Lucas and Mangeney, 2007; Lucas et al., 2011, 2014). We recognise that thin-layer models
 272 lack some of the features of real flows such as the presence of water, erosion/deposition processes or
 273 polydispersity (particles with very different sizes) (Delannay et al., 2017). However, our knowledge of
 274 the model limitations, in particular when compared to granular experiments (Mangeney-Castelnau et
 275 al., 2005; Mangeney et al., 2007; Gray, 2014; Rocha et al., 2019) makes it possible to better interpret
 276 the comparison between simulated and observed deposits.

277 In the simplest Coulomb friction law implemented in SHALTOP, the friction coefficient μ , is constant
 278 during the simulation. Using SHALTOP to simulate about 15 landslides on Earth and Mars, Lucas et al.
 279 (2014) found that the coefficient of friction decreases with increased volume of material released
 280 during these landslides due to the increase of the flow velocity. As a result, there is a relationship
 281 between the volume of the landslide and the friction coefficient associated with its movement. The
 282 friction coefficient therefore also varies as a function of the landslide runout distance.

283 In our investigation, we will vary this friction parameter to best-fit runout distance, and then compare
 284 the results to the empirical law of Lucas et al. (2014) $\mu = V^{-0.0774}$ where V is the landslide volume.

285 Laboratory experiments show that μ may actually depend on the flow velocity and thickness.

286 Hence, we will also use the Pouliquen and Forterre friction law (Pouliquen and Forterre, 2002) to
 287 simulate our landslides in the SHALTOP model. This law involves six empirical parameters: three
 288 friction angles, $\delta_{1,2,3}$, the particle size, L , an empirical dimensionless parameter, β (deduced from
 289 laboratory experiments and taken as constant here ($\beta = 0.136$)), and an exponent γ ($\gamma = 10^{-3}$).

290 Several regimes, low-velocity and high-velocity, are described depending on the Froude number (Fr)
 291 defined as $\frac{u}{\sqrt{gh}}$.

292 If $Fr \geq \beta$

$$293 \mu(h, u) = \tan \delta_1 + (\tan \delta_2 - \tan \delta_1) \frac{1}{1 + \frac{\beta h \sqrt{gh}}{L u}} \quad (1),$$

294 If $Fr = 0$

$$295 \mu(h) = \mu_{start}(h) = \tan \delta_3 + (\tan \delta_2 - \tan \delta_1) \frac{1}{1 + \frac{h}{L}} \quad (2),$$

296 If $0 \leq F_r \leq \beta$

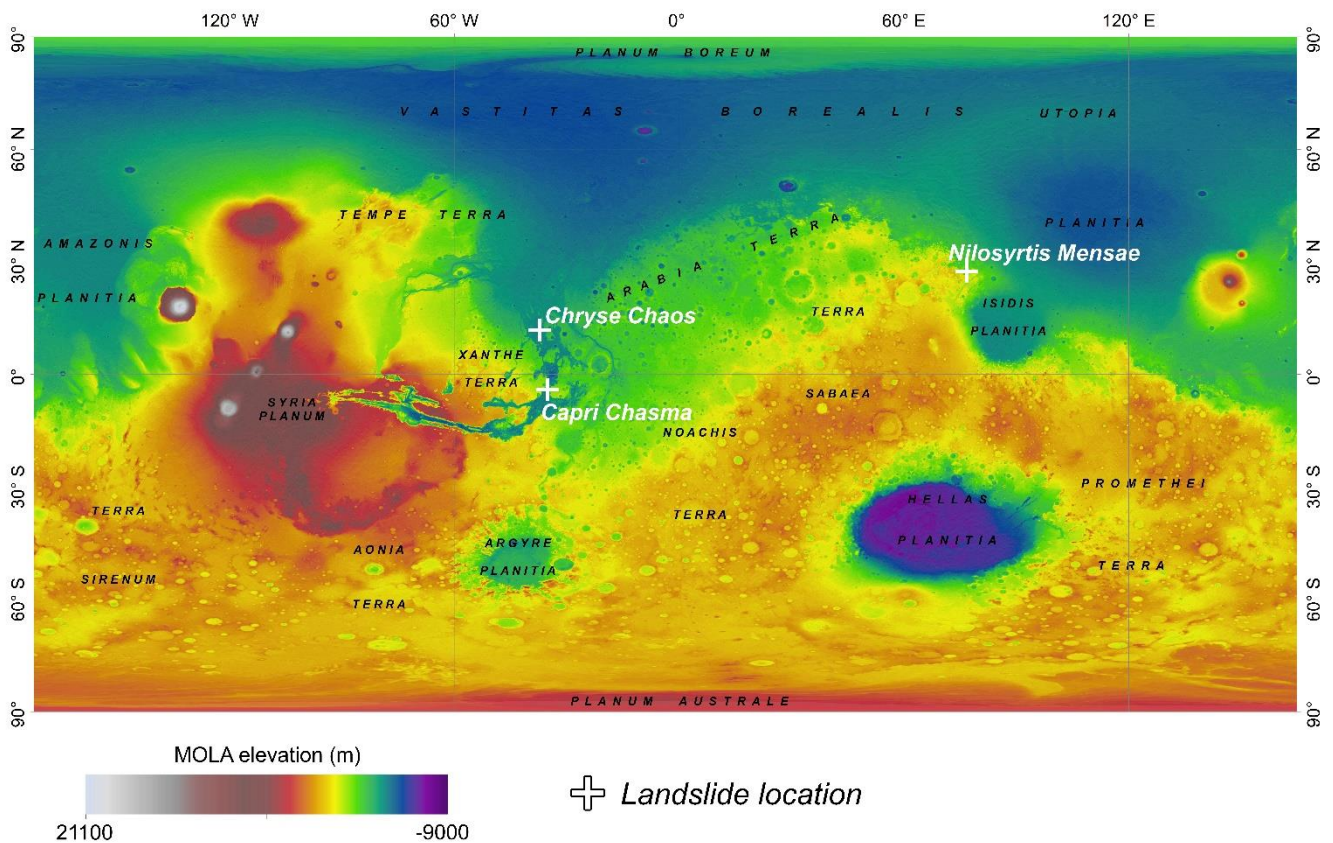
297
$$\mu(h, u) = \mu_{start}(h) + \left(\frac{F_r}{\beta}\right)^\gamma (\mu_{stop}(h) - \mu_{start}(h)) \quad (3),$$

298 Where

299
$$\mu_{stop}(h) = \tan \delta_1 + (\tan \delta_2 - \tan \delta_1) \frac{1}{1+\frac{h}{L}} \quad (4).$$

300 With the Pouliquen and Forterre law, when h decreases and u increases, μ increases and vice versa.
301 $\mu_{stop}(h)$ and $\mu_{start}(h)$ represent the tangent of the slope angle required for a certain material
302 thickness h to stop or start to flowing, respectively. In order to reduce the number of fitting parameters
303 in the Pouliquen and Forterre friction law, we fix the dimensionless parameters β and γ as well as the
304 difference $\delta_2 - \delta_1$ and $\delta_3 - \delta_1$. Thus the only parameters that will be fitted to match observed travel
305 distances are δ_1 (with $\delta_2 = \delta_1 + 10^\circ$ and $\delta_3 = \delta_1 + 2^\circ$) and L (Brunet et al., 2017).

306 In the model, we used the acceleration due to gravity on Mars $g = 3.73 \text{ m}\cdot\text{s}^{-2}$. We used a maximum
307 simulation duration of 1200 s, as after this time there was no further variation in both the velocity and
308 thickness of the simulated flows. Fourteen tests were carried out (seven for each landslide for the two
309 friction laws) in order to obtain the coefficients of friction that best reproduce the observed runout
310 distance.



311 Figure 3 - Location of the three studied landslides on a coloured MOLA topographic map of Mars with semi-transparent
312 shaded relief (Smith, et al., 2001).

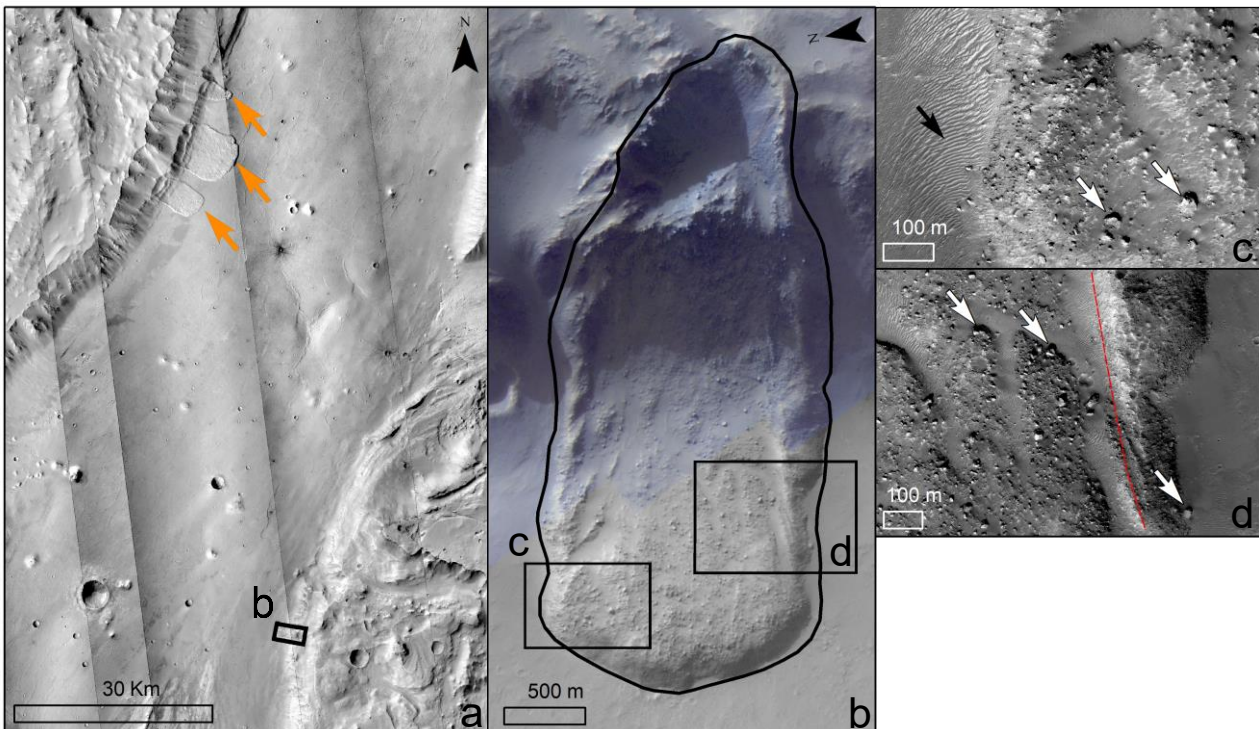
313

314 **3. Results**

315 In this section we present the geomorphological results (Section 3.1), the age estimates (Section 3.2)
316 and numerical modelling (Section 3.3). Section 3.1 is divided into two parts, the first concerning the
317 description of the results for the martian landslides followed by a comparison to the terrestrial
318 landslides.

319 **3.1. Geomorphological results**

320 **3.1.1. Martian landslides**



321 *Figure 4 - The Capri Chasma (CCh) landslide. (a) Regional view of the area around the CCh landslide, which is outlined by the*
322 *black rectangle indicating the location of panel (b), the orange arrows indicate three larger landslides located on the other*
323 *side of the outflow channel. The background is a mosaic of CTX images P20_008707_1757, P07_003578_1757,*
324 *G21_026271_1756, G03_019546_1753, and D06_029515_1755. (b) CaSSIS image of CCh landslide (MY35_008462_188, PAN-*
325 *BLU filters). (c) & (d) Detailed view of the landslide deposit, black arrows indicate windblown deposits, white arrows indicate*
326 *boulders more than 30 m in diameter, red dotted line indicates the inner edge of the southern lateral levee (HiRISE:*
327 *ESP_035831_1760). Credits: NASA/JPL/UofA/MSSS/ESA/Roscosmos/UniBe.*

328 **Capri Chasma landslide.** The Capri Chasma landslide (CCh) is located between Xanthe Terra and
329 Margaritifer Terra, at 4°4' S, 35°2' W (Fig.3). The region around this landslide is dominated by a series
330 of north-south trending canyons at the eastern extent of Capri Chasma, a canyon leading to the outflow
331 channels named Tiu Valles downstream (Coleman et al., 2007) (Fig.4a). The landslide is located on the
332 eastern flank of Capri Chasma on an escarpment which rises 2800 m above the canyon floor on the
333 middle Noachian highland unit 'mNh', (see Tanaka et al., 2014). Tanaka et al. (2014) describe this unit
334 as featuring degraded to severely degraded undifferentiated materials resulting from meteorite
335 impacts, volcanic flows, and possibly sedimentary and fluvial deposits and is dated to between 4.5 and
336 3.7 billion years old.

337 The landslide faces west, measures ~4 km long and ~1.7 km-wide at its widest point. Its key attributes
338 are summarised in Table 1. The arcuate landslide scar is characterised by a continuous well-defined

339 slope-break demarking the upper limit of the erosion zone (*Fig.5a, red dashed line*). The slopes in the
 340 erosion zone reach 52° at the scar itself, and gradually decrease in the downslope direction away from
 341 the scar. Below the scar, the erosion zone is dominated by a talus slope with metre-scale blocks.
 342 Downslope these blocks become mantled by aeolian deposits.

Martian Landslide		Capri Chasma (CCh)	Chryse Chaos (ChrC)	Nilosyrtis Mensae North Lobe (NM)	Nilosyrtis Mensae South lobe (NM)
Latitude		4°4' S	11° 43' N	27° 24' N	
Longitude		35°2' W	37° 6' W	76° 42' E	
Erosion zone	Maximum length (m)	1000	500	500	750
	Maximum width (m)	1080	1200	200	220
	Aspect ratio (length/width)	0.9	0.4	2.9	3.4
	Area (m ²)	1.4x10 ⁶	1.0x10 ⁶	6.4x10 ⁴	1.2x10 ⁵
	Steepest slope (°)	52	70	40	
	Mean slope (°)	30	35	30	
Transport zone	Maximum length (m)	1350	1700	-	
	Maximum width (m)	1685	1350	-	
	Mean slope (°)	26	23	-	
	Mean adjacent slope (°)	25	23	25	
Deposit zone	Maximum length (m)	1620	1050	590	830
	Maximum width (m)	1700	1000	210	190
	Area (m ²)	3.2x10 ⁶	2.0x10 ⁶	1.2x10 ⁵	1.7x10 ⁵
	Volume (m ³)	1.4x10 ⁸	4.0x10 ⁷	2.1x10 ⁶	2.4x10 ⁶
	Maximum thickness (m)	112	64	34	31
	Range (and mean) of front scarp angle (°)	21 – 27 (24)	19 – 24 (21)	27 – 32 (29)	27 – 28 (27.5)
	Mean adjacent slope (°)	1	3	20	
Levee	Mean height (m)	45	20	20	
	Range (and mean) of lateral angle (°)	9 - 15 (12)	8 - 10 (9)	22 - 30 (26)	23 - 25 (24)
Maximum boulder size (m)		40	60	10	

343 *Table 1 – Summary of the morphological attributes of the martian landslides. Note that the value given in volume of the*
 344 *deposit zone, takes into account the volume found in the deposit zone plus the volume of the lateral levees located in the*
 345 *transport zone for ChC and ChrC.*

346 At the base of the erosion zone, there is a slope reversal of ~18° where some of the failed material has
 347 remained (*Fig.5a, red dashed line*). This material has an angular texture similar to the material
 348 constituting the deposit further downslope.

349 Immediately below the main erosion scar, there is a transport zone with a slope of ~26° decreasing to
 350 1° near the deposit zone (*Fig.5a, between blue and red dashed lines*). In this zone low lateral levees are

351 present (Fig.5b, black arrows, Table 1). This transport zone contains fewer blocks than the deposit
 352 zone.

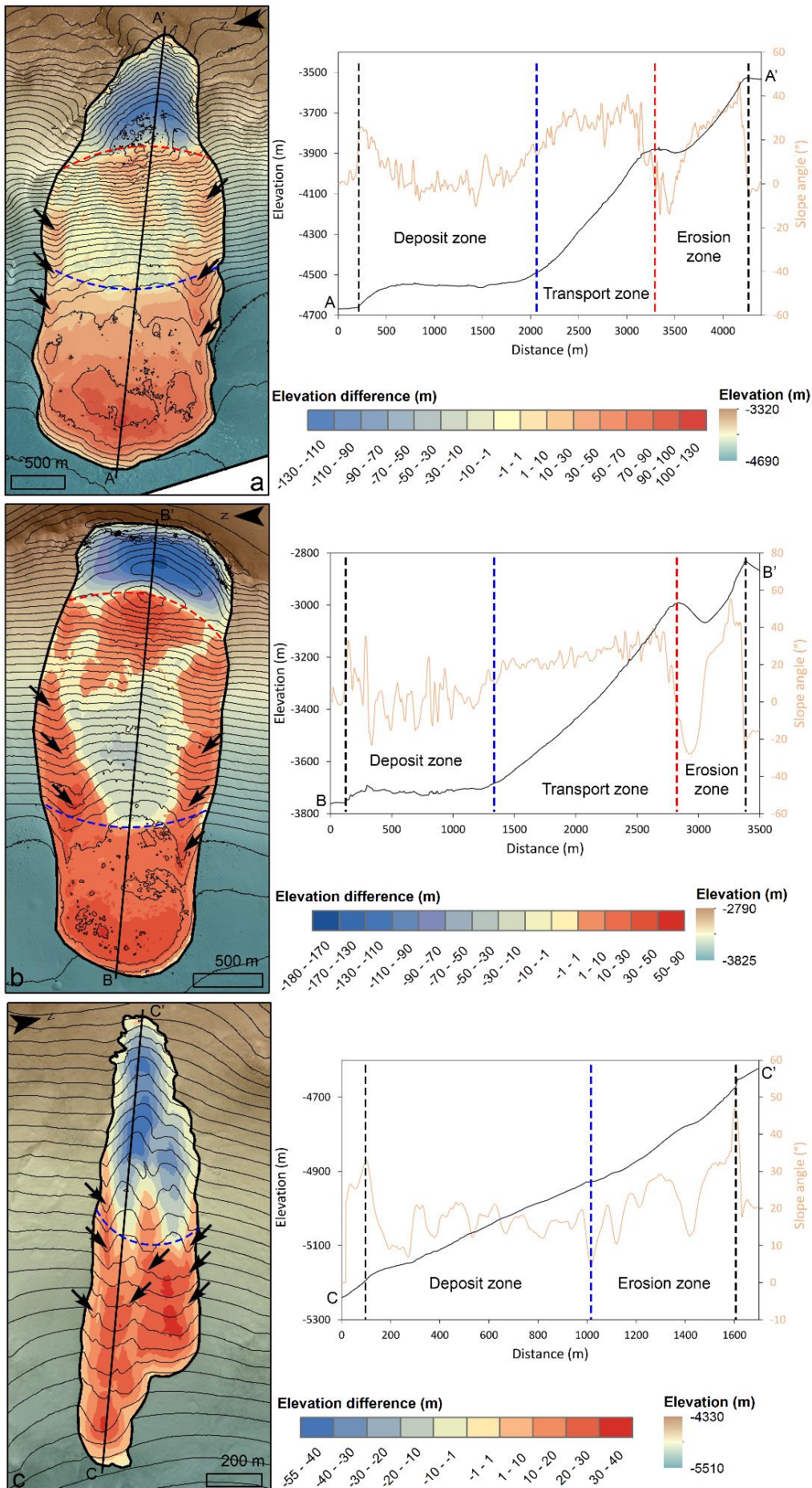
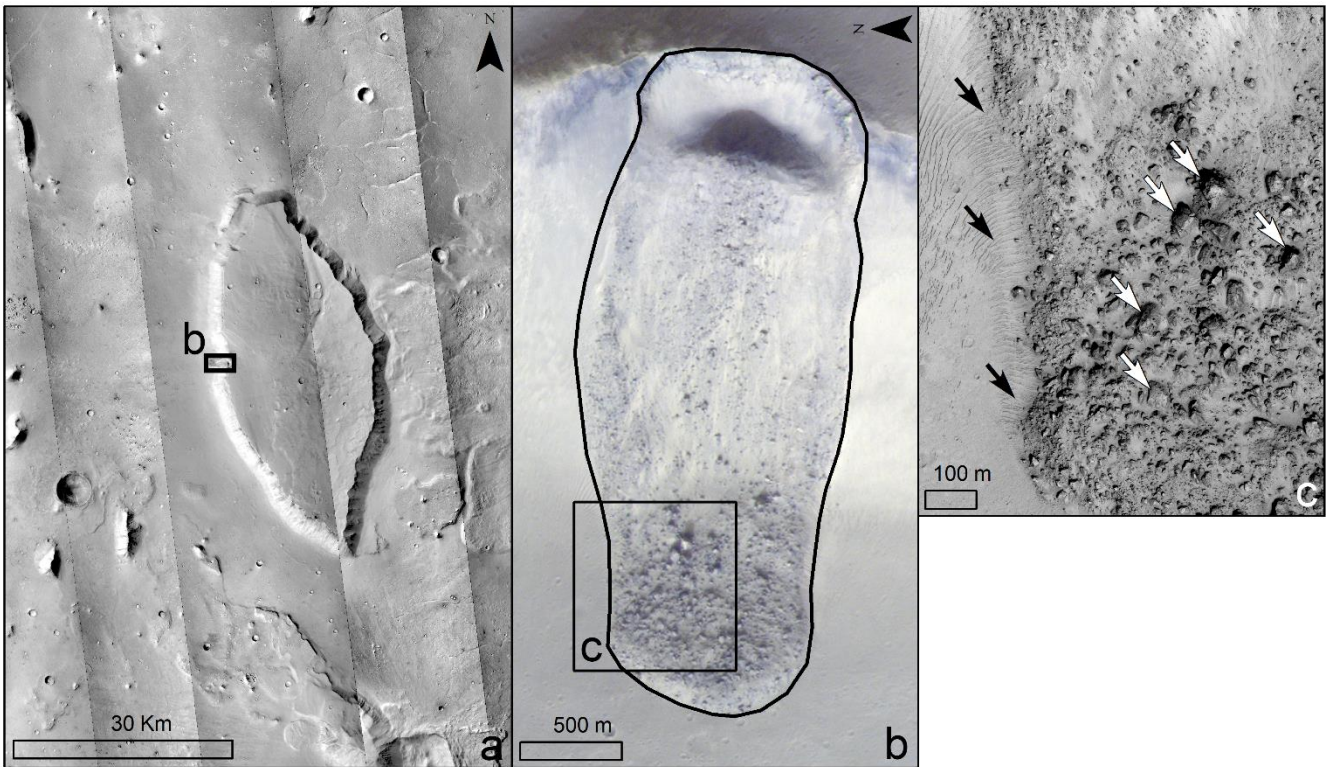


Figure 5 – Coloured digital elevation model with semi-transparent ortho-image overlay by elevation difference map, long profile and topographic contours with 25 m interval marked in black for (a) Capri Chasma CCh, A-A'; (b) Chryse Chaos ChrC, B-B' and (c) Nilosyrtris Mensae NM, C-C'. For each long profile plot the slope angle variation is indicated by the orange line. The dashed lines delimit the erosion, transport and deposition zones of the landslides. The black arrows indicate lateral levees. HiRISE images: (a) ESP_035831_1760; (b) PSP_005701_1920; (c) ESP_027480_2075. Credit: NASA/JPL/UofA.

377 The transport zone and the deposit zone that follows have similar width (*Fig.5a*, blue dashed line,
 378 *Table 1*). The deposit reaches a maximum thickness of 112 metres (*Fig.5a*). Where the deposit lies on
 379 flat ground, it forms a single steep-sided and flat-topped lobe with a thickness that increases gradually
 380 towards the toe. The texture of the deposition surface is rough and irregular. Several hundred blocks
 381 >10 m in diameter are distributed between the front of the deposit and at the base of the erosion zone
 382 (*Fig.4c & d*). Another concentration of blocks is found on the deposit that remains in the erosion zone,
 383 containing a block of 40 m in diameter. Some metre-scale aeolian bedforms are also observed at the
 384 toe of the deposit zone (*Fig.4c*).



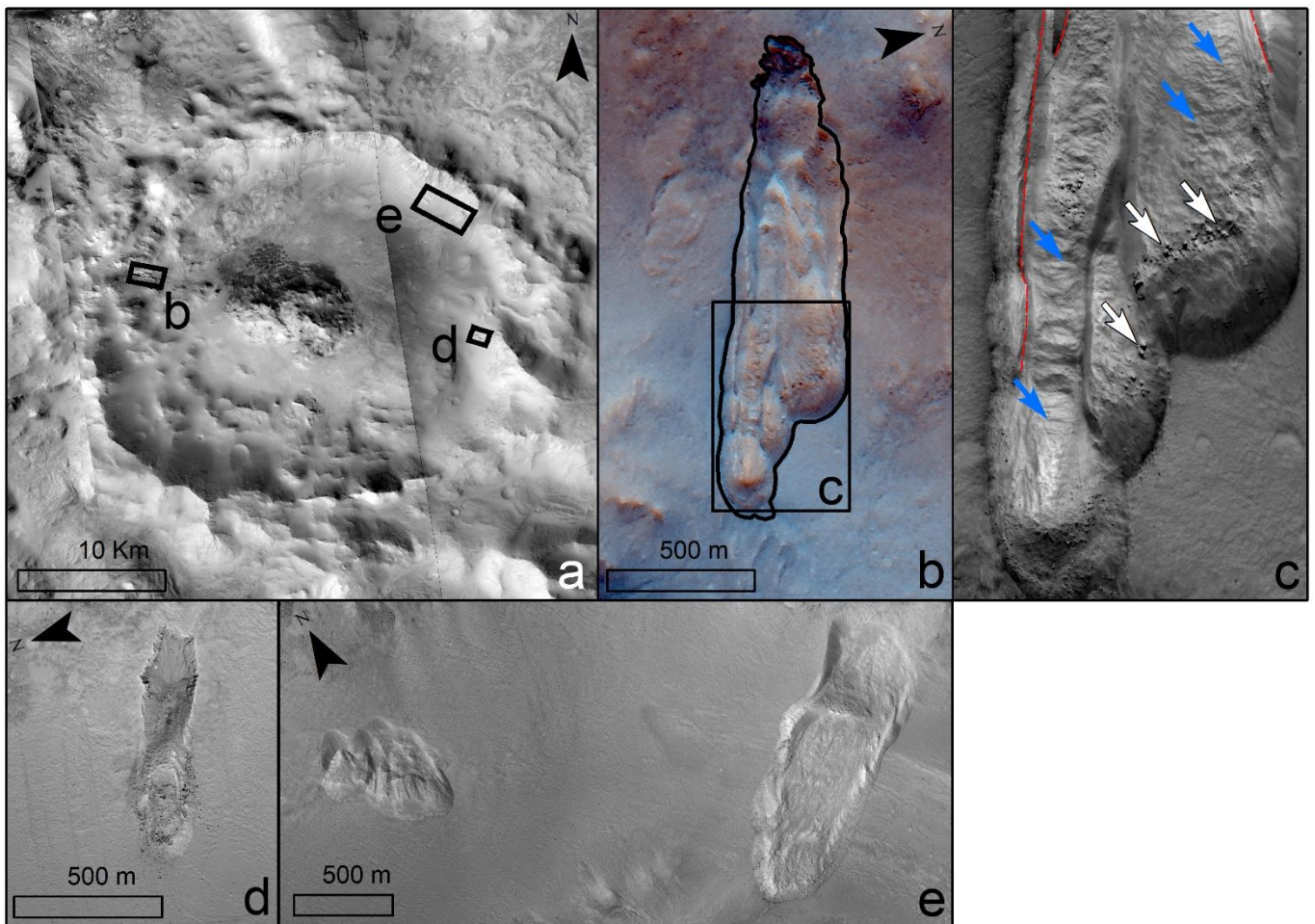
385 *Figure 6 - The Chryse Chaos (ChrC) landslide. (a) Regional view of the area surrounding the ChrC landslide where the black*
 386 *rectangle indicates the position of the landslide and panel b. CTX images: D15_033207_1901, G20_026060_1913,*
 387 *P08_004145_1902, P17_007626_1902, and F03_036965_1906. (b) CaSSIS image of the ChrC landslide (MY35_010023_012,*
 388 *PAN-BLU filters). (c) Detailed view of the landslide deposit, black arrows indicate windblown deposits, white arrows indicate*
 389 *boulders more than 50 m in diameter (HiRISE image PSP_005701_1920). Credits:*
 390 *NASA/JPL/MSSS/UofA/ESA/Roscosmos/UniBe.*

391 **Chryse Chaos landslide.** The Chryse Chaos landslide (ChrC) is located at 11°43' N, 37°6' W (*Fig.3*), in
 392 Simud Vallis, an outflow channel (Pajola et al., 2016) that together with Tiu Vallis is believed to have
 393 carried water flowing from Valles Marineris into the putative ocean of Chryse Planitia (Tanaka et al.,
 394 2003) (*Fig.6a*). Its key attributes are summarised in *Table 1*. Tanaka et al. (2014) report that this region
 395 is underlain by the 'Hto' unit, a transition valley unit dated to the Hesperian (3.56 – 3.24 Ga), composed
 396 of fluvial deposits from Tiu Vallis. Pajola et al. (2016) indicate four different evolutionary stages
 397 occurred in the area, including possible flow inversions and ponding. The landslide is located on the
 398 west-flank of a flat-topped mesa in the middle of the valley's floor rising up to 950 m above it. The
 399 mesa is composed of basement materials, with a modelled age that is Middle Noachian and consists
 400 of friable sediments, impact debris and volcanic material (Tanaka et al., 2014; Pajola et al., 2016).

401 The erosion scar of the ChrC landslide is well defined and, in some places, bedrock outcrops are
 402 apparent and several blocks seem to have detached from these outcrops. Below the scar, the erosion
 403 zone is characterised by a talus slope at ~35° with blocks visible at the base of the talus (*Fig.5b*) and is

404 relatively short compared to the total length of the landslide (*Fig.5b; Table 1*). As for the CCh landslide,
 405 the base of the erosion zone is characterised by a slope reversal of $\sim 20^\circ$ where deposits have remained
 406 (*Fig.5b, red dashed line*).

407 Downslope of the slope reversal there is the transport zone, characterised by slopes of approximately
 408 23° and lateral levees (*Fig.5b, black arrows, Table 1*). The transport zone width decreases from 1350 m
 409 to 920 m when the slope reaches 3° at the deposit zone. Then as for CCh landslide, the transport zone
 410 is followed by the deposit zone where the deposit forms a single steep-sided and flat-topped lobe. Its
 411 thickness gradually increases toward the toe to reach a maximum thickness of about 64 m (*Fig.5b*). We
 412 observe fluctuations in slope angle on the deposit zone that indicates a very high surface roughness,
 413 which can be explained by the presence of several dozen boulders of 60 m in diameter (*Fig.6c, white*
 414 *arrows*), also as for CCh, several hundred blocks >10 m in diameter are distributed between the front
 415 of the deposit and at the base of the erosion zone. We also observe some aeolian bedforms at the toe
 416 of the deposit (*Fig.6c, black arrows*).



417 *Figure 7 - The Nilosyrtris Mensae (NM) landslide. (a) Regional view of the area surrounding the NM landslide. Black rectangles*
 418 *indicate the positions of panels b, d and e. CTX images: D04_028680_2064, B16_015942_2088, B11_014109_2058,*
 419 *G01_018777_2091, G18_025225_2058, P06_003310_2082, and B03_010707_2080. (b) CaSSIS colour image of the Nilosyrtris*
 420 *Mensae landslide (MY35_008751_028, PAN-BLU filters). (c) Detailed view of the NM landslide in (b), where white arrows*
 421 *indicate boulders more than 5 m in diameter, and blue arrows indicate perpendicular ridges on the deposit (HiRISE:*
 422 *ESP_026781_2075). (d) & (e) Other landslides in same impact crater on HiRISE images ESP_053457_2075 &*
 423 *ESP_057110_2075, respectively. Credits: NASA/JPL/UofA/MSSS/ESA/Roscosmos/UniBe.*

424

425 ***Nilosyrtris Mensae landslide.*** The Nilosyrtris Mensae (NM) landslide (*Fig.7b*) is located at 27°24' N,
426 76°42' E, 150 km to the north of the Nili Fossae, at the southwest margin of Utopia Planitia and
427 southeast of Nilosyrtris Mensae on the western wall of a 25 km diameter and 1.7 km deep impact crater
428 (*Fig.7a*). The crater is located within a transitional unit dated between the Noachian and the Hesperian
429 (Hnt), which is composed of Noachian impacts, sedimentary and volcanic deposits, and intervening
430 aprons dated to the Hesperian (Tanaka et al., 2014). This region is characterised by the presence of
431 tectonic grabens (the Nili Fossae) generated by a major fault system, probably related to the Isidis
432 impact basin (Wichman and Schultz, 1989; Kraal et al., 1998). The region is also characterised by the
433 presence of fluvial erosion and deltaic deposits which date back to the Late Noachian, between 3.85
434 to 3.7 Ga (Fassett and Head, 2005). In addition, phyllosilicates have also been reported in this region
435 (Bibring et al., 2005; Poulet et al., 2005; Mustard et al., 2007; Mangold et al., 2007; Carter et al., 2013).

436 The NM landslide is located on a continuous slope inclined eastward at ~25° (*Fig.5c*), and has two main
437 lobate structures (*Fig.7b*). A summary of the characteristics of both of these lobes can be found in
438 *Table 1*. The landslide has an erosion and deposition zone, but does not show a distinct transport zone
439 (*Fig.5c*).

440 Two erosion zones are present upslope of each lobate deposit with their interior slopes reaching a
441 maximum value of 40°. There are tracks left by rolling blocks (e.g., Tesson et al., 2019), particularly in
442 the southern erosion zone. No slope breaks are observed between the erosion zone and deposit zone
443 (*Fig.5c*).

444 Lateral levees are present on the flanks of the landslide depositional lobes (*black arrows on Fig.5c*,
445 *Table 1*). Levees on the southernmost lobe are more pronounced than those on the northern lobe
446 (*Fig.5c*). The deposits are also characterised by ridges perpendicular to the direction of flow, mainly
447 present at the distal end of the landslide (*Fig.7c*, *blue arrows*). The average deposit thickness is ~30 m
448 spread evenly over the whole deposit surface with a maximum thickness of ~34 m. Also, we were
449 able to distinguish blocks of about 10 m in diameter at the front of the landslide deposition surface.

450 A similar lobate morphology is shown by three other smaller landslides within the same crater (*Figs.7d*
451 *& e*). They are 1.6 km, 900 m and 500 m long, respectively and are present on a similar substrate, at a
452 similar altitude as the landslide studied here.

453 **3.1.2. Terrestrial landslides**

454 To better understand the formation mechanism of martian landslides, we compare them to terrestrial
455 analogues. The morphological description of the three martian landslides shows a clear morphological
456 difference between the CCh and ChrC landslides and the NM landslide. For this reason, the description
457 of terrestrial analogues has been subdivided into two different sections.

458 ***Capri Chasma and Chryse Chaos analogue.*** We compare here the CCh and ChrC landslides to a similar-
459 looking landslide located in the Abisko region in Sweden at 68°12' N, 19°2' E. The landslide (*Fig.8a &*
460 *b*) is classified as a rockslide by Rapp (1960). Rapp (1960) indicates that it is probably a post-glacial
461 landslide because the rockslide deposit partly covers glaciofluvial deposits. Its formation has therefore
462 been linked to the release of overburden pressure induced by the disappearance of the valley glacier.
463 The measurements concerning the Abisko landslide are summarised in *Table 2*.

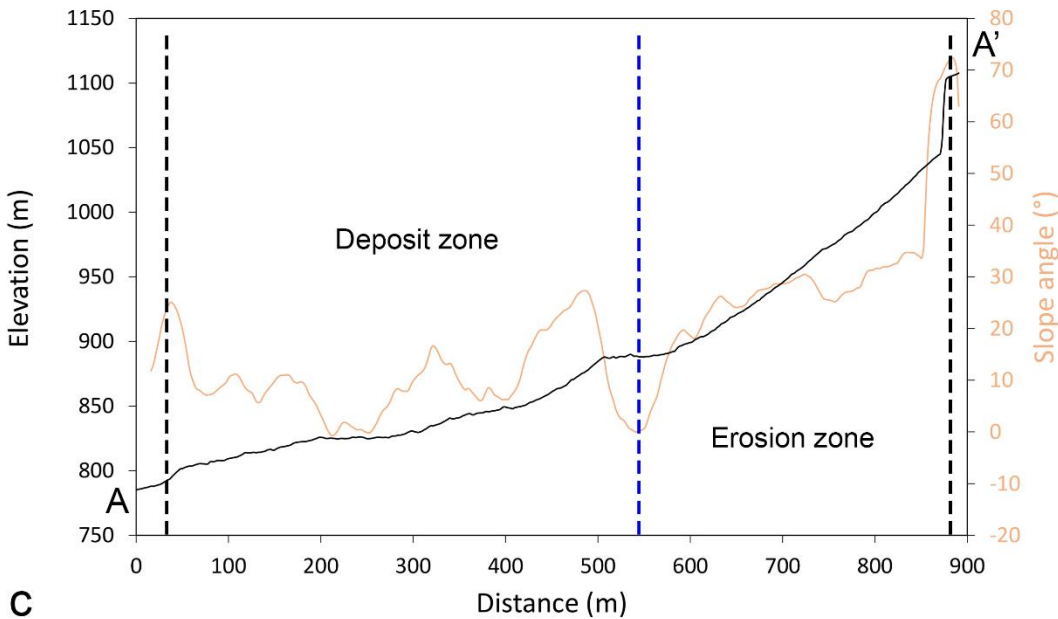
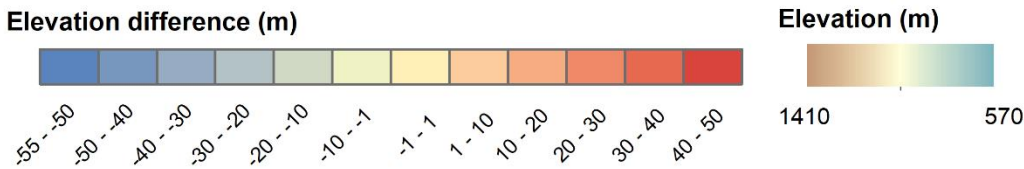
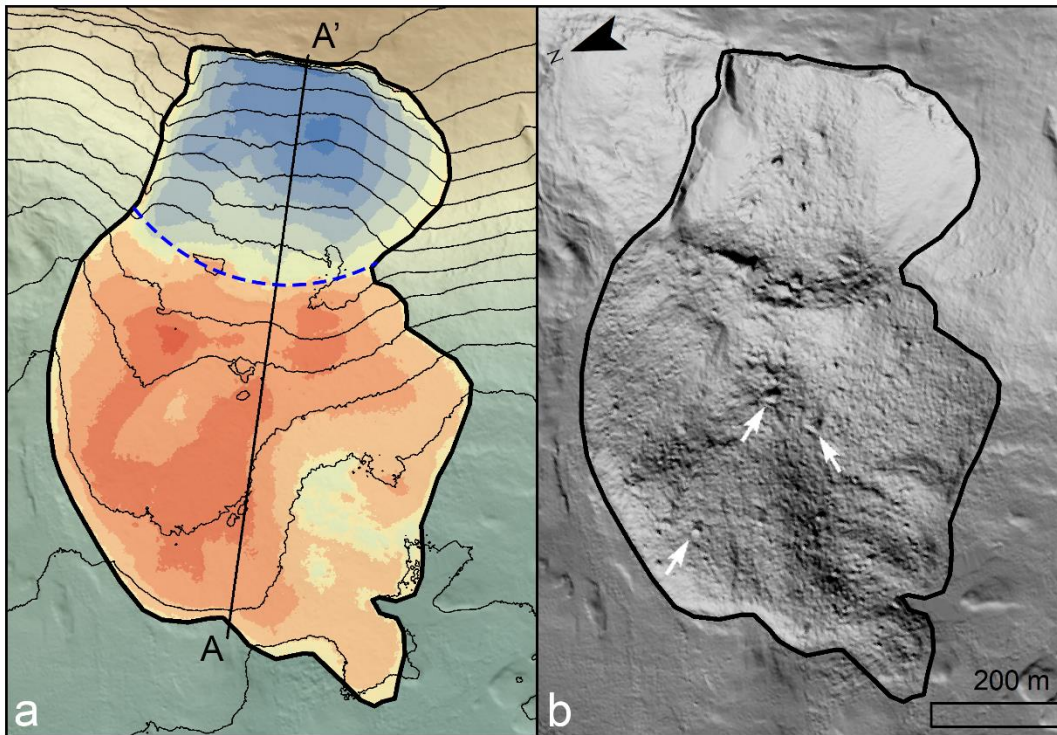


Figure 8 – The Abisko rockslide, Sweden. (a) Colourised digital elevation model with semi-transparent hillshaded relief overlain by elevation difference map and long profile derived from LiDAR topography for the Abisko landslide marked with the location of the long profile in panel c. Topographic contours with 25 m interval are marked in black. (b) LIDAR hillshaded relief image for the Abisko landslide, white arrows indicate boulders on the deposit. (c) Elevation profile and slope angle variation of the Abisko rockslide from A to A' in panel a, dashed lines indicate the erosion and deposit zones. Image credits: Geographical Sweden Data (GDS).

Terrestrial Landslides		Abisko	Hólmavík	Mount Rainier
Latitude		67°12' N	65°42' N	46°59' N
Longitude		19°2' E	21°42' W	121°40' W
Erosion zone	Maximum length (m)	340	340	550
	Maximum width (m)	440	150	350
	Aspect ratio (length/width)	0.7	2.2	1.5
	Steepest slope (°)	85	65	70
	Mean slope (°)	39	33	43
Transport zone	Maximum length (m)	-	100	-
	Maximum width (m)	-	140	-
	Mean slope (°)	-	11	-
	Mean adjacent slope (°)	27	15	24
Deposit zone	Maximum length (m)	519	295 - 320	435 - 470
	Maximum width (m)	510	150	350
	Area (m ²)	1.2x10 ⁶	5.7x10 ⁴	1.6x10 ⁶
	Volume (m ³)	3.5x10 ⁶	3.3x10 ⁵	2.2x10 ⁶
	Maximum thickness (m)	34	12	33
	Range (and mean) of front scarp angle (°)	14- 21 (17)	12 - 20 (17)	29 - 36 (32)
	Mean adjacent slope (°)	6	10	15
Levee	Mean height	-	10	50
	Mean (and range) of lateral angle (°)	-	13 - 21 (17)	16 - 19 (17)
Maximum block size (m)		30	6	20

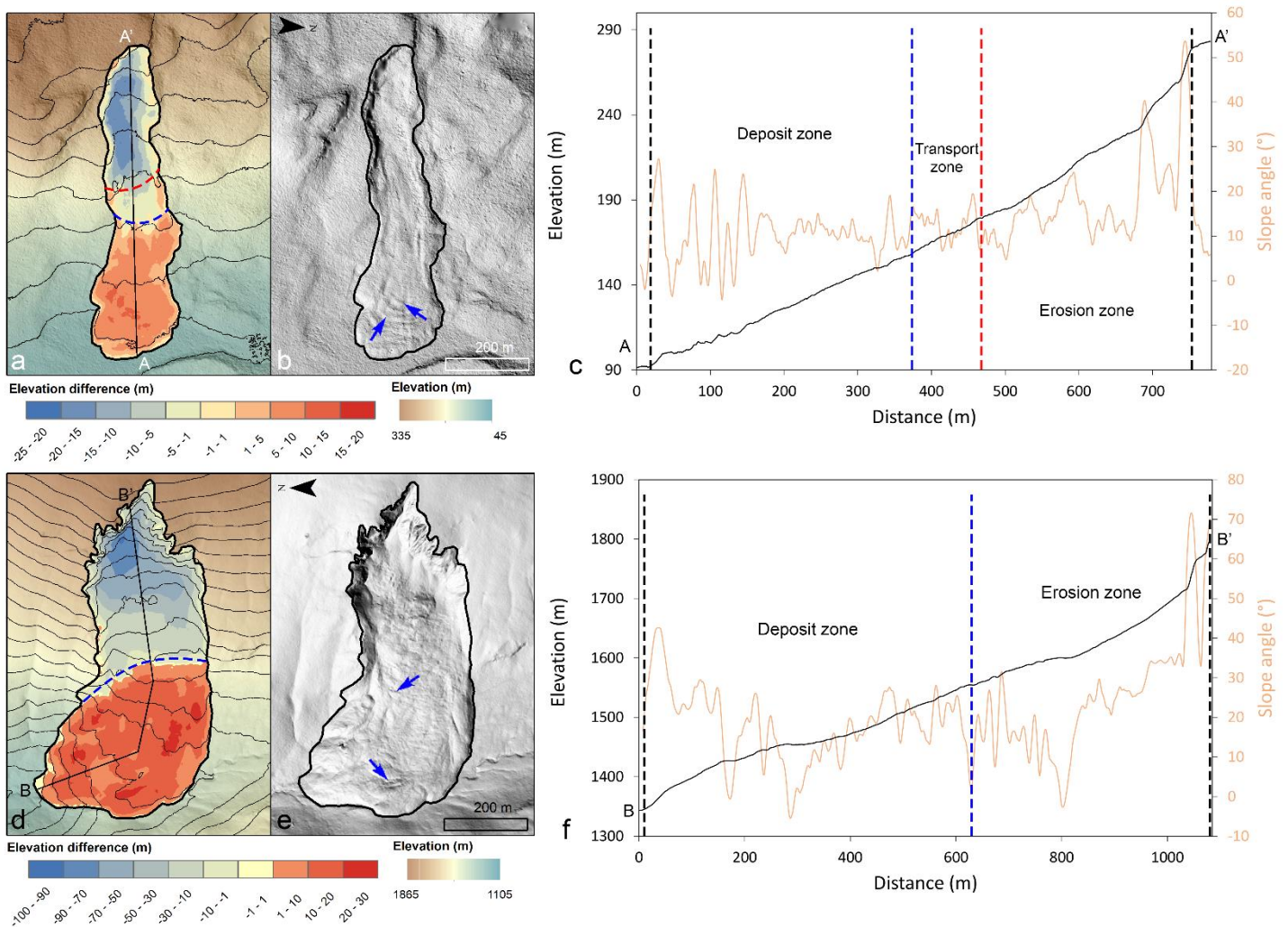
495 *Table 2 - Summary of the morphological attributes of the terrestrial analogue landslides.*

496 The morphology, topography and texture of the Abisko landslide have three notable similarities to the
497 CCh and ChrC landslides:

- 498 1. The erosion scar is sharp and well defined (*Fig.8c*), with the steepest values near the scar (85°
499 in Abisko, compared to 70° in ChrC, *Fig.5b* and 52° for the CCh landslide, *Fig.5a, Table 1 & 2*)
500 and rapidly descending to ~39° on the talus slope below.
- 501 2. The length/width ratio of the erosion zones are similar, 0.4 for the ChrC landslide, and 0.7 for
502 the Abisko rockslide. However, in the CCh landslide the length/width ratio is 0.9. In Abisko
503 there is no slope inversion at the end of the erosion zone as there is in ChrC and CCh (*Figs.5a*
504 & *b*). Yet, the slope does lower almost to zero where the deposited material has stalled on the
505 lower slope.

506 3. The deposit areas for the Abisko rockslide, and ChrC and CCh landslides are covered by blocks
 507 of tens of metres in diameter. These blocks are highlighted in *Figures 8b* by the white arrows
 508 and reach 30 m in diameter. These blocks were also noted by Rapp (1960). The deposit zone
 509 is located only on gently inclined topography in all three cases.

510 The two main differences between the Abisko rockslide and the ChrC and CCh landslides are their scale
 511 and a difference in the pre-existing topography. Rapp (1960) estimated that the Abisko landslide has a
 512 volume between 1×10^6 to 2×10^6 m³, and we have calculated its volume to be 3.5×10^6 m³ which is
 513 less than the volume of CCh and ChrC (*Table 1*). At Abisko, the adjacent hillslope has a slope of 27°
 514 (*Table 2* and *Fig.8c*), compared to 25° and 23° on Mars, for CCh and ChrC. There is a lack of a transport
 515 zone and its associated levees in the Abisko rockslide. The planview shape and mass distribution of the
 516 Abisko rockslide is more irregular than the ChrC and CCh landslides. Also, Abisko deposits is located on
 517 a non-zero surface slope, unlike the martian landslides.



518 *Figure 9 - Hólmafvík mudslide (Iceland) and Mount Rainier mudslide (US). (a) Elevation difference map overlain on colourised*
 519 *DEM with semi-transparent hillshaded relief for the Hólmafvík mudslide. Topographic contours with 25 m interval are marked*
 520 *in black. (b) Hillshaded relief of the Hólmafvík mudslide. (c) Elevation profile and slope angle variation of the Hólmafvík mudslide*
 521 *from A to A' in panel a. (d) Elevation difference map overlain on colourised DEM with semi-transparent hill-shaded relief*
 522 *derived from LiDAR topography for the Mount Rainier mudslide. Topographic contours with 25 m interval are marked in black.*
 523 *(e) Hillshaded relief image derived from LiDAR topography for the Mount Rainier mudslide. (f) Elevation profile and slope angle*
 524 *variation of the Mount Rainier mudslide from B to B' in panel (d). Blue arrows on panel (b) and (e) indicate ridges on the*
 525 *landslide deposit and dashed lines in black, red and blue indicate the erosion, transport and deposit zones. Credit: Land Survey*
 526 *of Iceland and Washington Lidar Portal.*

527 ***Nilosyrtris Mensae analogue.*** The morphological analysis of NM landslide shows several similarities
528 with terrestrial mudslides. Two terrestrial analogues have been identified, one in Iceland, near the
529 town of Hólmavík in Iceland (65°42' N, 21°42' W, *Fig.9b*) and one in the US, near Mount Rainier in
530 Washington State (46°59'N, 121°40'W, *Fig.9e*).

531 Both terrestrial examples are smaller in scale than the NM landslide but similar in shape. The
532 similarities can be summarised as follows:

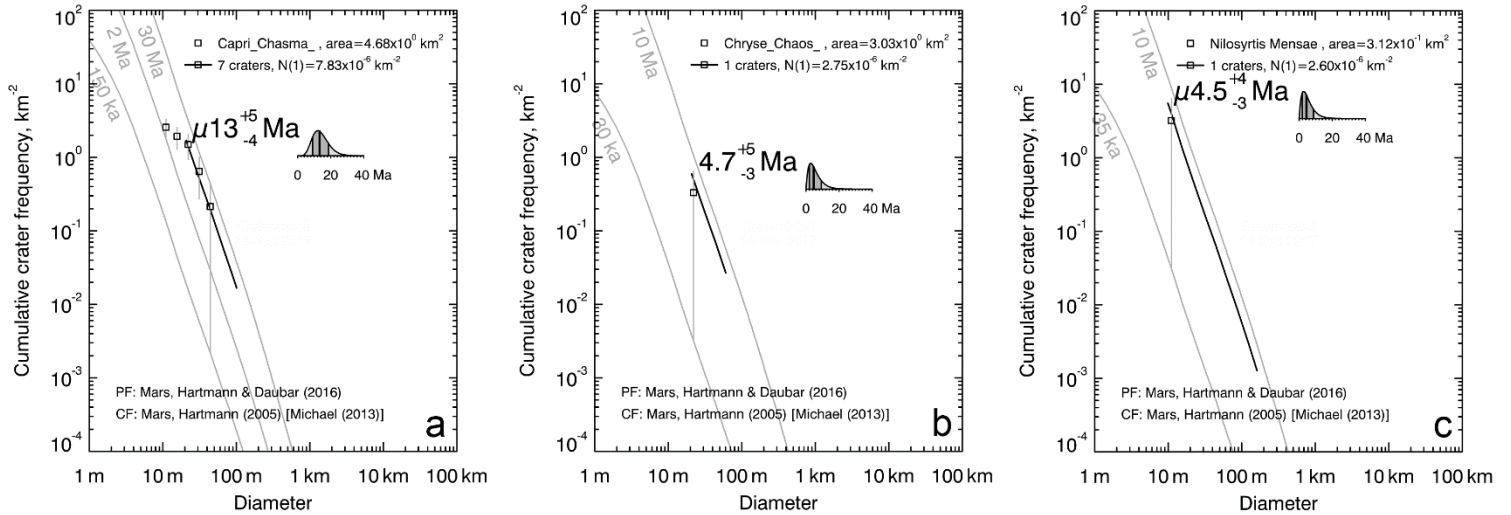
- 533 1. In both the terrestrial mudslides and in the NM landslide, the erosion scar is sharp and has an
534 irregular outline (*Fig.9b, 9e*). The erosion zone for Hólmavík and NM has an elongate shape.
535 The length/width ratio of the erosion zone for Hólmavík is 2.2 and for NM landslide it is 2.9
536 and 3.4 for northern and southern parts, respectively. For comparison the Mt Rainier mudslide
537 has an aspect ratio of 1.5.
- 538 2. The terrestrial mudslides and the NM landslide all form along a continuous hillslope, rather
539 than at an escarpment like the rockslides described above (*Fig.9c, 9f, Table 1 & 2*). None of
540 these landslides has evidence for the involvement of substantial consolidated bedrock, but
541 rather soil materials.
- 542 3. The terrestrial mudslides and the NM landslide have ridges perpendicular to the flow direction.
543 These ridges are particularly well developed in the Hólmavík mudslide and are weakly present
544 in the Mt Rainier mudslide. In Hólmavík they have a height of 11 m (*Fig.9b, blue arrow*) and in
545 Mt Rainier a height of 5 m (*Fig.9e, blue arrow*), compared to 3 m on the NM landslide (*Fig.7c,*
546 *blue arrow*). These compression ridges have already been observed in earthflows (Parise,
547 2003) and submarine landslides (Hildenbrand et al., 2006; Masson et al., 2002). Neither the
548 terrestrial mudslides nor the NM landslide have abundant blocks at their surface.
- 549 4. Lobate margins. The terrestrial mudslides and the NM landslide all have multi-lobed terminal
550 margins to their deposits. The relief of the margins is 12 m at Hólmavík and more than 33 m
551 at Mt Rainier, compared to 34 m at NM (*Table 1, 2*).
- 552 5. Lateral levees. The terrestrial mudslides and the NM landslide possess lateral levees found in
553 the transport zones of the landslides. The levees lateral slopes are 13° – 21° (*Table 2*)
554 compared to 22° – 30° on Mars.

555 There are some differences between the terrestrial and martian landslide, most notably their
556 respective sizes (*Tables 1 and 2*). A transport zone can be identified in the Hólmavík landslide, but not
557 in the Mt Rainier and martian landslides. The terrestrial analogues also have erosion zones that are
558 steeper than in the NM landslide and they also have a smaller underlying slope of 5 to 10° in the deposit
559 zone compared to the martian landslide at 20° (*Tables 1 and 2*).

560 **3.2. Age estimations of martian landslides**

561 For the CCh landslide, we identified 12 impact craters on the deposit, and the largest of these craters
562 has a diameter of 35 m. We used the crater-size frequency distribution of this landslide to estimate its
563 age to be 13±5 Ma (*Fig.10a*). To obtain this estimate, we used only the 7 largest impact craters with a
564 minimum diameter of 21 m to avoid sampling bias due to image resolution at smaller size.

565 For the ChrC and NM landslides, the number of craters is too low to provide an absolute age with one
566 23 m crater for ChrC and no craters for NM. Only an estimate of the maximum age of formation of
567 these landslides was possible. In the case of NM, due to the lack of any impact crater, we estimated a
568 maximum age by artificially adding the smallest diameter impact crater that we deemed to be possible



569 Figure 10 - Crater size-frequency plots with selected isochrons for the three dated martian landslides. (a) Capri Chasma
 570 landslide; (b) Chryse Chaos landslide; (c) Nilosyrtris Mensae landslide. PF and CF indicate the production function and
 571 chronology function, respectively. To the right of each age estimate the uncertainty is shown as a probability density function
 572 using Poisson statistics (Michael et al., 2016).

573 to identify in the images (10 m), a method suggested by (Hartmann, 2005). The estimated maximum
 574 age of formation of ChrC is $\sim 4.5 \pm 4$ Ma (Fig.10b), whereas for the NM landslide the maximum age is
 575 $\sim 4.7 \pm 4$ Ma (Fig.10c).

576 The ages obtained using this dating method do not exceed 20 Ma, hence these landslides formed
 577 recently, during the Late Amazonian.

578 3.3. Numerical modelling results

579 A summary of the friction coefficients used in the simulations along with the best-fit results are
 580 presented in Table 3. We used $L = 5$ m as the mean grain size for CCh and ChrC and $L = 1$ cm for NM,
 581 based on the observed maximum block size (see section 3.1.1) and chosen after some sensitivity tests
 582 (see supplementary material, S3- Grain size sensitivity tests).

583 On Figures 11, 12 and 13 we present the simulated dynamics of the three landslides and show the
 584 deposits from our best-fit simulations alongside the observed deposits. For CCh and ChrC landslides,
 585 none of our simulations produced entirely satisfactory results. The best-fit simulation was obtained
 586 using the Pouliquen and Forterre law (Fig.11f and 12f) using friction angle of $\delta_1 = \tan^{-1}(0.16)$, $\delta_2 =$
 587 $\tan^{-1}(0.33)$ and $\delta_3 = \tan^{-1}(0.19)$ for CCh and $\delta_1 = \tan^{-1}(0.19)$, $\delta_2 = \tan^{-1}(0.36)$ and $\delta_3 =$
 588 $\tan^{-1}(0.22)$ for ChrC, and produced a closer deposit shape to the observed deposit (Fig.11g and 12g)
 589 than the best-fit Coulomb law simulations (Fig.11e and 12e) with friction angles of $\delta = \tan^{-1}(0.27)$
 590 for ChC and $\delta = \tan^{-1}(0.30)$ for ChrC. The centre of mass of the simulated deposits is near the base
 591 of the slope (Fig.11f and 12f, red point) and compares well to the centre of mass of observed deposit
 592 (Figure 11g and 12g, black point). However, in the model results no substantial mass remains at the
 593 base of the erosion zone as is actually observed.

594 In addition, the simulated deposit is thicker in the centre and thinner towards the edges, whereas our
 595 observations show that the deposits tend to have a more constant thickness, leading to a steeper front
 596 scarp (See Table 4). For CCh the model underestimates the maximum deposit thickness (70-75 m in
 597 the model, Fig.11f and 112 m observed, Fig.11g) and in ChrC the model overestimates the maximum

Landslide	Initial mass (m ³)	Model parameters					Empirical Value $\mu_{\text{eff}}=V^{-0.0774}$
		Test run	Pouliquen and Forterre's law			Coulomb's law	
			$\tan(\delta_1)$	$\tan(\delta_2)$	$\tan(\delta_3)$	$\tan(\delta)$	
Capri Chasma	9.7x10 ⁷	1	0.12	0.30	0.16	0.23	0.24
		2	0.14	0.32	0.18	0.25	
		3	0.16	0.33	0.19	0.26	
		4	0.18	0.35	0.21	0.27	
		5	0.19	0.37	0.23	0.28	
		6	0.21	0.39	0.25	0.29	
		7	0.23	0.41	0.27	0.30	
Chryse Chaos	6.2x10 ⁷	1	0.16	0.33	0.19	0.27	0.24
		2	0.18	0.35	0.21	0.28	
		3	0.19	0.36	0.22	0.29	
		4	0.19	0.37	0.23	0.30	
		5	0.21	0.39	0.25	0.31	
		6	0.26	0.43	0.29	0.32	
		7	0.27	0.44	0.30	0.32	
Nilosyrtis Mensae	4.2x10 ⁶	1	0.32	0.50	0.36	0.34	0.34
		2	0.34	0.52	0.38	0.35	
		3	0.35	0.53	0.39	0.36	
		4	0.36	0.54	0.40	0.37	
		5	0.37	0.55	0.41	0.38	
		6	0.38	0.56	0.42	0.39	
		7	0.40	0.58	0.44	0.40	

598 Table 3 - Summary of friction coefficient values tested for each of our three landslides using the Pouliquen and Forterre and
599 Coulomb laws and the effective friction coefficient using the empirical law determined by Lucas et al. (2014) for landslides
600 volume of more than 10³ m³. In red, the best fit coefficients are indicated. For Capri Chasma and Chryse Chaos a grain size of
601 L=5m was used and for Nili Fossae, L=1cm. The initial mass is estimated from the difference between the observed topography
602 and the reconstructed scar topography and differs from the deposit volumes stated in Table 1 because it includes the stalled
603 mass in the erosion zone.

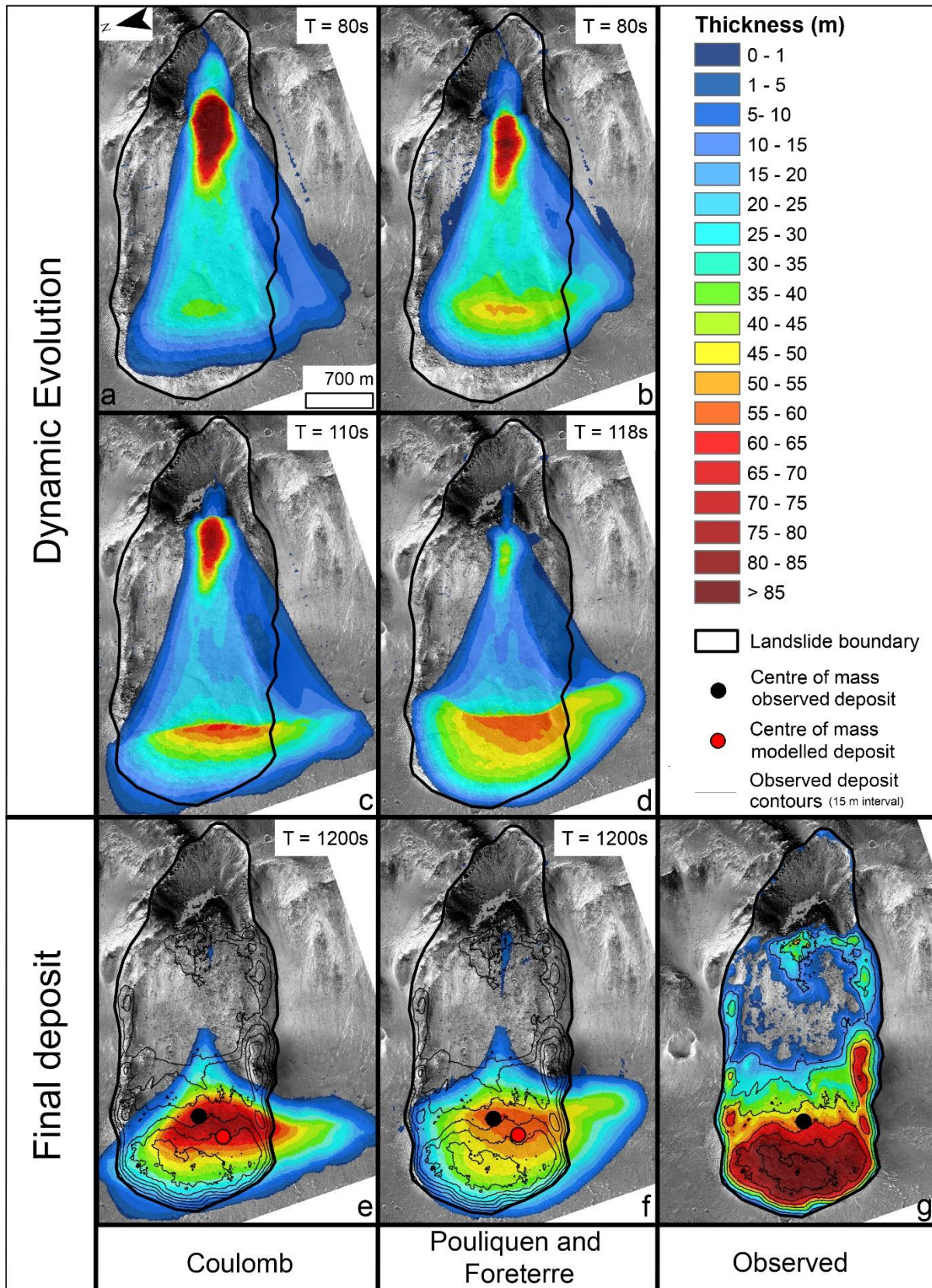
Landslide		Capri Chasma	Chryse Chaos	Nilosyrtis Mensae
Observed mean front deposit angle (°)		24	21	28
Mean simulation front deposit angle (°)	Coulomb's law	5	6	2
	Pouliquen and Forterre's law	3	4	3

604 Table 4 - Comparison between the front deposition angle measured in simulations using Coulomb's and Pouliquen and
605 Forterre's laws and the real measured deposition angle. The mean angle was determined from three different measurements
606 taken on the real and simulated deposition front.

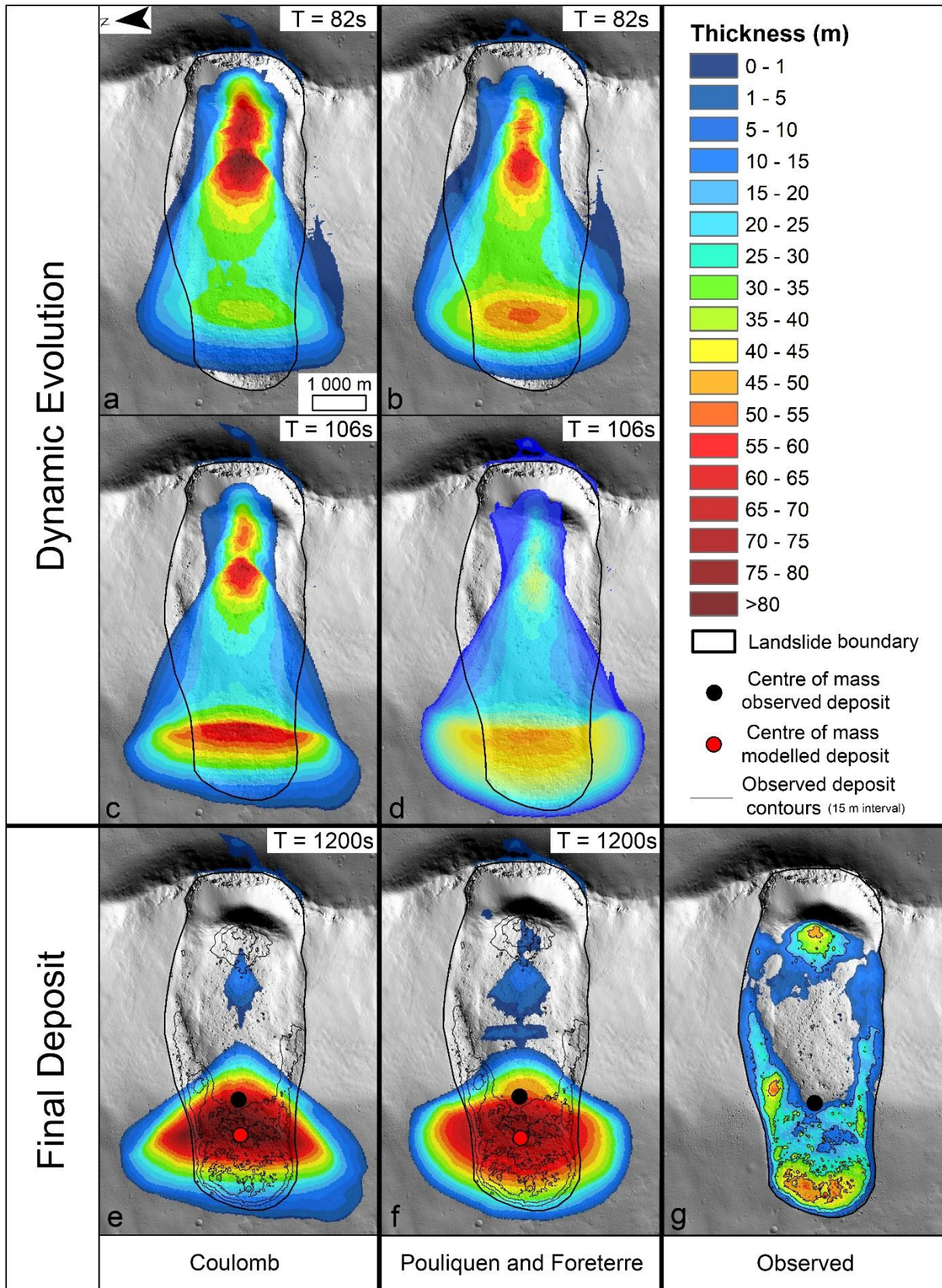
607 deposit thickness (85 m in the model, Fig.12f vs ~64 m observed, Fig.12g). Finally, the modelled
608 deposits spread over a wider area than the observed deposits in CCh and ChrC and no levees are
609 observed, despite the fact that the Pouliquen and Forterre friction law is capable of producing levees
610 under certain conditions (Mangeney et al., 2007).

611 For the NM landslide we observed a poor fit between the two-simulation law (Coulomb, *Fig.13e* and
612 Pouliquen and Forterre, *Fig.13f*) and the observed deposit. We used friction angle $\delta = \tan^{-1}(0.36)$
613 for Coulomb law and $\delta_1 = \tan^{-1}(0.36)$, $\delta_2 = \tan^{-1}(0.54)$ and $\delta_3 = \tan^{-1}(0.40)$ for Pouliquen and
614 Forterre law. We observed that for each law the simulated deposit centre of mass is close to the centre
615 of mass of the observed deposit. A large proportion of the eroded mass remains within the erosion
616 zone for each simulation.

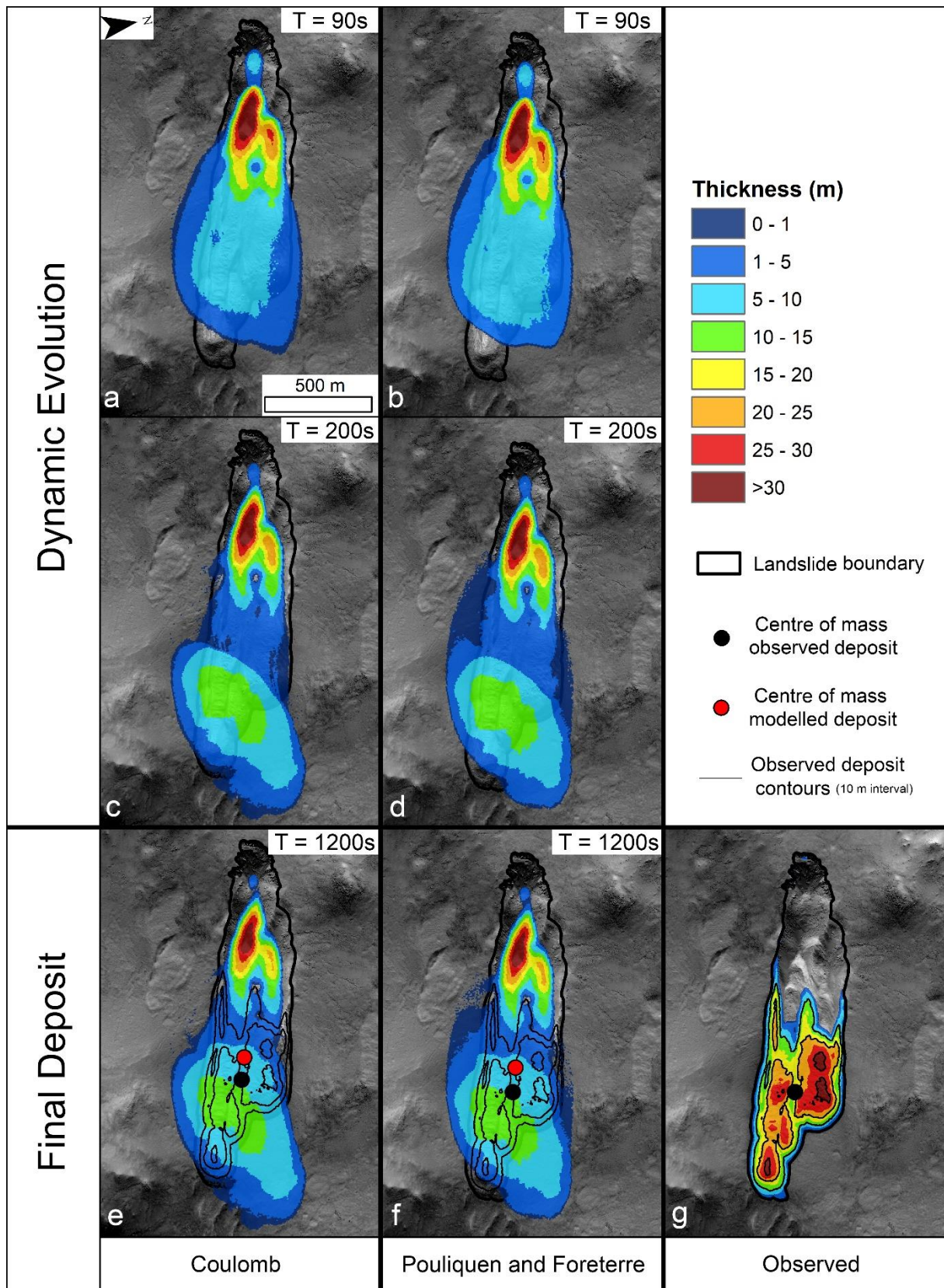
617 After tests involving the release of the masses of both erosion zones at the same time and variation of
618 the friction parameters, the best fit model produces a maximum deposit thickness of 15 m whereas it
619 is actually ~ 32 m. The overall shape of the landslide is not well matched; the furthest downslope extent
620 is located to the north of where it should be (along the steepest line of descent) (*Fig.13e and 13f*). Also,
621 the model cannot reproduce the lateral levees observed on the real deposit (*Fig.13g*).



622 Figure 11 – Modelled and observed deposit thicknesses for Capri Chasma landslide. (a-d) Dynamic evolution of the deposits
 623 for grain size $L=5\text{m}$ using Coulomb's law at $T=80\text{s}$ (a), $T=110\text{s}$ (c) and $T=1200\text{s}$ (e) and using Pouliquen and Forterre law (2002)
 624 at $T=80\text{s}$ (b), $T=118\text{s}$ (d) and $T=1200\text{s}$ (f). The times at 1200s represent fully stabilised deposits. (g) The observed deposit. The
 625 simulations focused on reproducing the final morphology of the deposits, so the timesteps during the simulation are provided
 626 here to illustrate the dynamics in the model. Background image HiRISE ESP_035831_1760. Credit: NASA/JPL/UofA.



627 Figure 12 - Dynamic evolution of Chryse Chaos landslide modelling for grain size $L=5m$ using Coulomb's law at $T=82s$ (a),
 628 $T=106s$ (c) and $T=1200s$ (e) and using Pouliquen and Forterre law, 2002 at $T=82s$ (b), $T=106s$ (d) and $T=1200s$ (f). The times at
 629 1200s represent fully stabilised deposits. The observed deposit (g). The simulations focused on reproducing the final
 630 morphology of the deposits, so the timesteps during the simulation are provided here to illustrate the dynamics in the model.
 631 Background is the hillshaded relief rendering of the DEM.

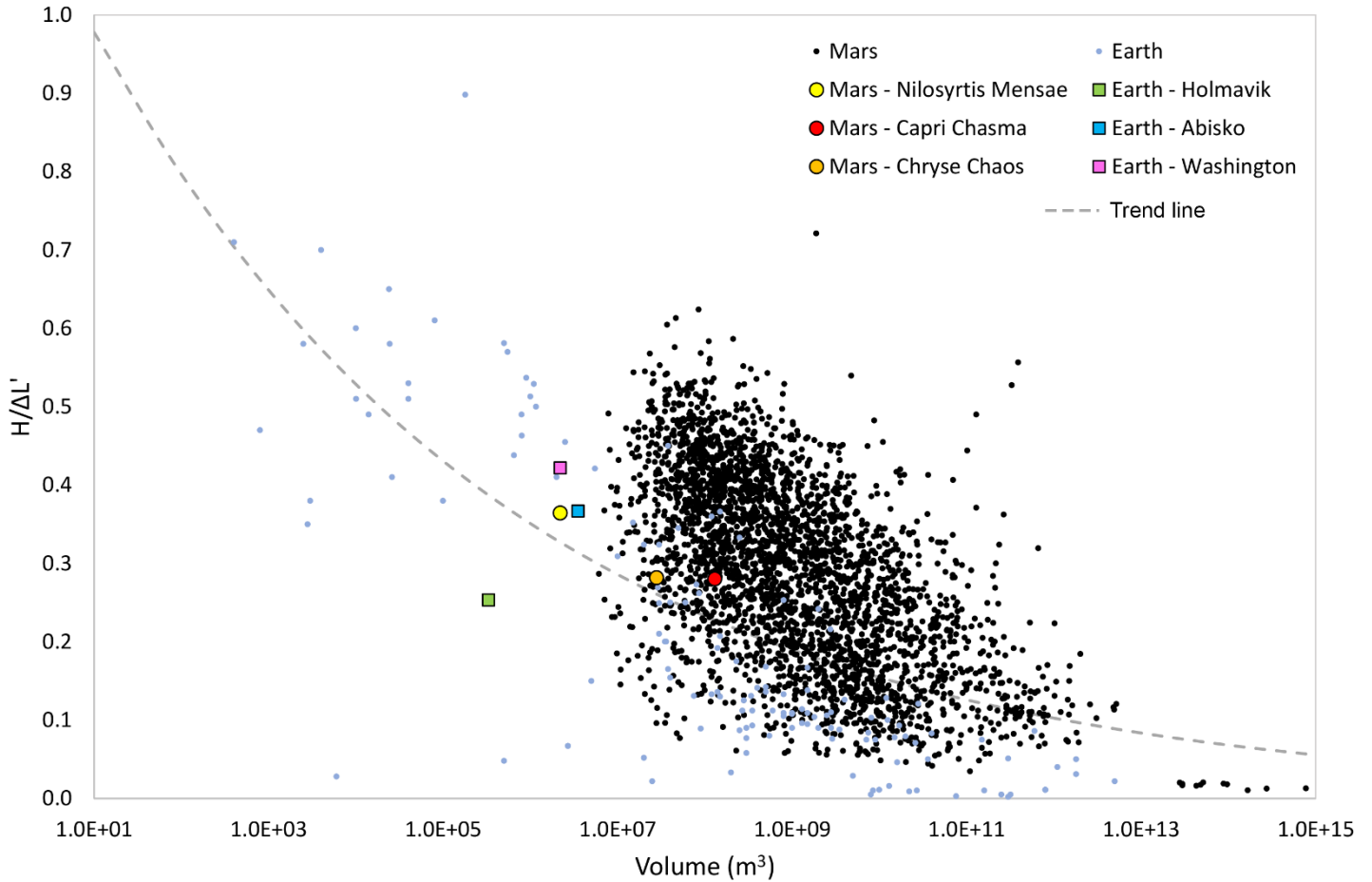


632 Figure 13 - Dynamic evolution of Nili Fossae landslide modelling for grain size $L=1\text{cm}$ using Coulomb's law at $T=90\text{s}$ (a), $T=200\text{s}$
 633 (c) and $T=1200\text{s}$ (e) and using Pouliquen and Forterre law (2002) at $T=90\text{s}$ (b), $T=200\text{s}$ (d) and $T=1200\text{s}$ (f). The times at 1200s
 634 represent fully stabilised deposits. The observed deposit (g). The simulations focused on reproducing the final morphology of
 635 the deposits, so the timesteps during the simulation are provided here to illustrate the dynamics in the model. Background is
 636 HiRISE image ESP_026781_2075. Image credits: NASA/JPL/UofA.

637 **4. Discussion**

638 In the following sections, we will first discuss how the investigated landslides compare to other martian
 639 and terrestrial landslides, and then their likely emplacement mechanisms suggested by our
 640 geomorphic observations and numerical modelling. Finally, we propose different scenarios that could
 641 have led to the formation of these landslides.

642 **4.1. Comparison with other martian landslides**



643 *Figure 14 - Heim's ratio ($H/\Delta L'$) plotted against volume for the martian landslides (coloured dots) and terrestrial landslides*
 644 *(coloured squares) of this study compared to terrestrial (blue dots, data from Legros et al., 2002, Lucas et al. 2014) and martian*
 645 *landslides (black dots, data from Quantin et al., 2004; Brunetti et al., 2014 ; Lucas et al. 2014 ; Crosta et al., 2018).*

646 **Context.** The morphology of the three martian landslides studied here differs from that found in
 647 previous studies of landslides on Mars, which mostly focused on the large landslides in Valles Marineris
 648 (Lucchitta, 1979; McEwen, 1989; Shaller, 1991; Quantin et al., 2004a; Soukhovitskaya and Manga,
 649 2006; De Blasio, 2011; Brunetti et al., 2014; Airo, 2015). These large landslides have a very large
 650 deposition area with an average of 10^9 m² (Quantin et al., 2004a) while our landslides have deposition
 651 areas ranging from 10^5 to 10^6 m². The large landslides in Valles Marineris often have overlapping layers
 652 of deposits (Grindrod and Warner, 2014) and longitudinal furrows (De Blasio, 2011; Magnarini et al.,
 653 2019), which is not the case for the three landslides presented here. Some Valles Marineris landslides
 654 do have deposit zones with a similar width to their erosion zones (Fig.4a, orange arrows) as found in
 655 our landslides.

656 When we compare the Heim's ratio (see section 2.1.3) and volume of our three martian landslides with
 657 the martian and terrestrial landslides from the literature (Legros, 2002; Quantin et al., 2004a; Lucas et
 658 al., 2014; Brunetti et al., 2014; Crosta et al., 2018) we observe that CCh and ChrC are towards the
 659 smaller end of the typical volumes or martian landslides and NM falls outside the martian population
 660 (Fig.14). Despite a gravitational difference between the Earth and Mars, the comparison between
 661 landslides is still possible because in the case of dry landslides, the flow is governed by the balance
 662 between the driving and resistance forces that are all proportional to the surface gravity (Johnson and
 663 Campbell, 2017). More specifically, only the velocity u and stopping time t_f depend on g (Mangeny-
 664 Castelnau et al., 2005; Mangeny et al., 2010). As described in Mangeny et al. (2010), the constant
 665 acceleration resulting from the sum of forces due to gravity (c_0) and friction (m), are defined as
 666 $c_0 = \sqrt{kgh_0 \cos \theta}$ and $m = g \cos \theta (\tan \theta - \tan \delta)$, with k being a constant, θ is the slope
 667 inclination and h_0 the initial thickness. The front of the landslide stops when its velocity
 668 $u = 2c_0 + mt_f = 0$, with the stopping time (t_f) defined as:

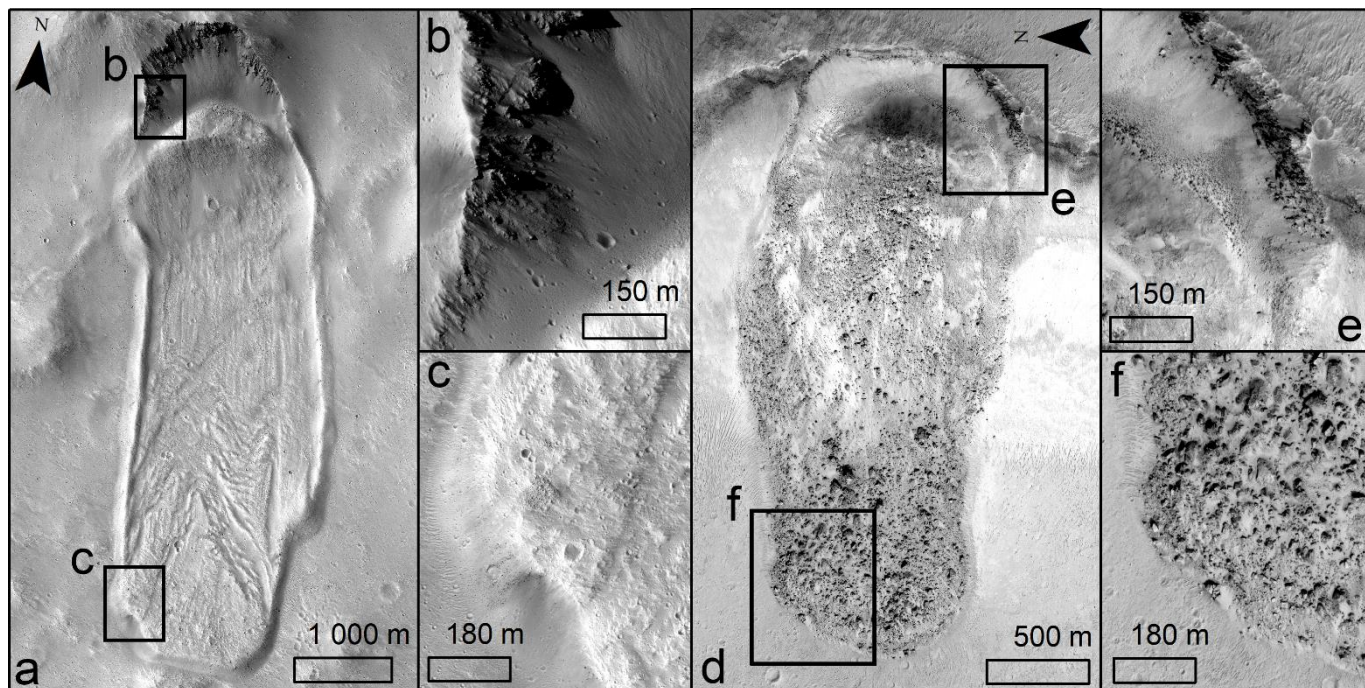
$$669 \quad t_f = \frac{2\sqrt{k}}{\tan \delta - \tan \theta} \tau_c \quad (5)$$

670 where $\tau_c = \sqrt{h_0 / (g \cos \theta)}$, the characteristic free fall time.

671 In case of the Coulomb friction law where $\mu = \tan \delta$ is constant, there is no dependency on gravity.
 672 But in case of the Pouliquen and Forterre friction law, as can be seen in Section 2.3.3 equation (1),
 673 gravity is involved in the friction coefficient calculation, and in this case if g decreases, the friction
 674 coefficient increases, so the velocity decreases.

675 Our martian landslides are located along the trend line predicted by Lucas et al. (2014) as calculated
 676 from the landslides used in their study where $\frac{H}{\Delta L'} = 1.2 \times V^{-0.089}$ with V being the volume of the
 677 landslide. This trend line (grey dashed line, Fig.14) was calculated for landslides with mainly dry
 678 granular behaviour. As our landslides roughly follow this trend line, this is consistent with them having
 679 a dry granular behaviour, but does not exclude other mechanisms. This trend line does not take into
 680 account the morphology of the deposits which is another key indicator of the physical processes at
 681 work during the flow, as will be discussed below.

682 **Age.** We estimate the age of these three martian landslides to be less than 20 Ma, subject to a large
 683 margin of error given the small surface area of these landslides and the size of the craters used to
 684 perform the dating (e.g., Warner et al., 2015). The older age of CCh compared to ChrC landslide is
 685 corroborated by the lower frequency of large blocks at the surface of the deposits of ChC compared to
 686 ChrC and may be related to the breakdown of rocks over time (e.g., de Haas et al., 2013). Figure 15
 687 further illustrates this point, showing a landslide with a high frequency of superposed craters (so
 688 presumably older than both CCh and ChrC) located near Montevallo crater (15°;54 °W) (Fig.15a). On
 689 this landslide the blocks are less numerous and less visible (Fig.15b), possibly covered by aeolian
 690 deposits, whereas a lot of blocks are still visible on the fresh landslide (Fig.15d). Whether or not water
 691 was involved in the formation of the ancient landslides in Valles Marineris is still under debate, because
 692 the climate could have been favourable to liquid water before the Amazonian. In the case of our
 693 studied landslides, they formed under recent martian climate conditions, which are thought to be
 694 similar to the present one, that is, dominated by cold temperatures and low atmospheric pressure,
 695 and for which the distribution of volatiles in is limited to within the ground at mid-and high-latitudes
 696 (e.g., Head et al., 2003).



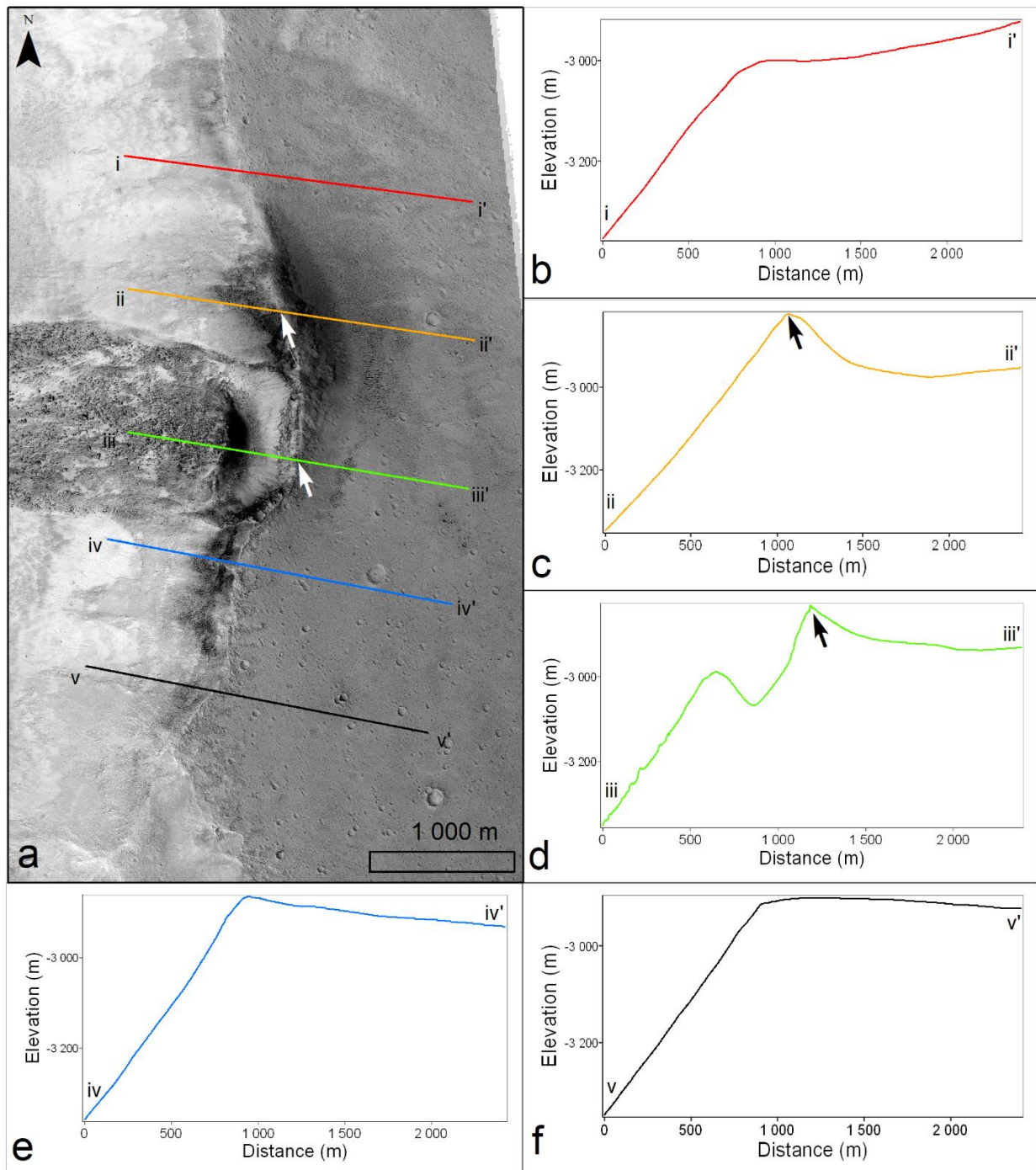
697 *Figure 15 - Morphological comparison between a degraded landslide (a) and a fresh landslide (d). (a) Landslide in Montevallo*
 698 *crater (15°N; 54°W). (b) Detailed view of the erosion zone showing multiple superposed craters and few visible blocks at the*
 699 *foot of the talus. (c) Detailed view of the deposition zone with superposed craters and a scattering of barely resolvable blocks.*
 700 *(d) Capri Chasma landslide (this study). (e) Detailed view of the erosion zone showing blocks at the foot of the talus slope and*
 701 *tracks from rolling blocks and no visible craters. (f) Detailed view of the deposition zone with densely packed large blocks and*
 702 *no visible impact craters. HiRISE images a, b, c: ESP_027643_1955; d, e, f: ESP_050033_1920. Credits: NASA/JPL/UoFA.*

703 4.2. Emplacement mechanisms

704 4.2.1. Geomorphic constraints

705 **Comparison between the three martian landslides.** Though the landslides of CCh and ChrC are similar,
 706 two main features set them apart. Firstly, for the ChrC landslide, we observed a raised rim present
 707 upslope of the erosion zone, just above the scar (*Fig.16a white arrows and Fig.16c & d, black arrows*),
 708 which is not present in the CCh landslide. We suggest that this raised topography could be the
 709 remnants of a rim of an impact crater, which formed before the Chryse outflow channel and therefore
 710 could be a location of pre-existing weakness. Secondly, the overall morphology of the CCh landslide
 711 appears more degraded, which is consistent with its older age from the crater size-frequency
 712 distribution analysis (Section 3.2). It has fewer and smaller (40 m compared to 60 m for ChrC, *Table 1*)
 713 visible blocks at the surface. The blocks in the deposition area of CCh may have been covered by
 714 windblown deposits, or broken down (de Haas et al., 2013). The surface roughness is lower for CCh
 715 than for the ChrC landslide. The older age of the CCh landslide could also explain why the slopes in the
 716 erosion zone are lower than in the ChrC landslide (30° compared to 35°, *Table 1*). Despite these
 717 differences, the similarities in topographic setting and morphological features (levees, deposit
 718 thickness, blocky surface texture) indicate these landslides had a very similar formation mechanism.

719 In contrast, the NM landslide has morphological characteristics that allow it to be easily distinguished
 720 from the other two landslides studied. The first difference lies in the shape of the erosion zones. In the
 721 case of CCh and ChrC landslide, the landslide scar forms a well-marked single arc, while for NM the
 722 scar is irregular. The slope profiles of the landslides shown in *Figure 5* show the similarity between the
 723 landslides of CCh and ChrC, where the scar is located at a sudden change in the slope of the



724 Figure 16 - (a) HiRISE image PSP_005701_1920 of Chryse Chaos landslide where the position of the profiles in panels b-f are
 725 indicated in corresponding colours. (b) to (f) Show a series of topographic profiles across the escarpment of the Chryse Chaos
 726 mesa progressing from north to south. Black arrows on profiles c and d correspond to white arrows on the HiRISE image in a
 727 and highlight the possible raised rim of a remnant impact crater. Credit: NASA/JPL/UofA.

728 topography. Furthermore, the erosion zone contains a slope inversion at its base where materials have
 729 stalled. On the contrary, the NM landslide scar is located in the middle of a continuous 25° slope, and
 730 no slope inversion is visible at the base of the erosion zone. The images and the oscillations in the slope
 731 profiles (Fig. 5) show that the deposit zones of CCh and ChrC are dominated by large blocks, whereas
 732 the NM landslide has a lower size and density of blocks on the deposit surface. These blocks seem to
 733 have been re-entrained from the middle lobe by the movement of the southern lobe (Fig. 7c, white
 734 arrow). We infer that the final morphology of the NM deposits results from at least two events. In the

735 NM landslide, there are ridges perpendicular to the direction of flow in the deposition zone, which are
736 not found on the CCh and ChrC landslides. Finally, the deposition front of the NM landslide has a multi-
737 lobe shape that is not found in the CCh and ChrC landslides, which have a gently curved flow front. The
738 levees are also less marked on the CCh and ChrC landslides than on the NM landslide.

739 The morphological differences indicate different flow dynamics between the CCh/ChrC landslides and
740 the NM landslide.

741 **Comparison between martian landslides and terrestrial analogues.** Here, we use our comparison to
742 terrestrial analogues presented in Section 3.1.2 to infer formation mechanisms for the martian
743 landslides. For CCh and ChrC, the rockslide near Abisko has similar topographic and morphological
744 features, including: the erosion scar, which is well defined, the erosion zone with similar slopes angle
745 and a deposition zone with blocks of 30 m in diameter distributed over the entire surface. However,
746 we also noted some morphological differences, which we argue can be explained by the shape and
747 size of the pre-existing slope rather than differing formation mechanisms. At Abisko, the hillslope has
748 a height and slope of 170 m and 27° (*Table 2, Fig.8c*), compared to ~1000 m and 25° on Mars (*Table 1,*
749 *Fig.5*). This difference results in a lack of a transport zone and therefore can explain the lack of levees
750 in the Abisko rockslide, as the erosion and deposit zones are directly adjacent. Where there is an
751 extended transport zone, other rockslides do show low-slope external levees (e.g., Shea and van Wyk
752 de Vries, 2008). The planview shape and mass distribution of the Abisko rockslide is more irregular and
753 the deposits have a non-zero surface slope, unlike the ChrC and CCh landslides. We attribute these
754 features to the difference in the topography underlying the deposit zone: the valley floor is not flat in
755 Abisko (6°) and is irregular unlike on Mars (slope < 3° and relatively smooth). Hence, we conclude that
756 the similarities observed between the Abisko and ChrC and CCh landslides imply similar formation
757 mechanism - catastrophic bedrock failure, whose downslope transport was driven by the action of
758 gravity. McSaveney and Davies (2007) describe the mechanism behind rockslides, as a simple gravity
759 driven movement of bedrock downslope. They usually have a single erosion surface, or thin zones of
760 intense shear strain (McSaveney and Davies, 2007).

761 As described in Section 3.1.2 we observed several key similarities between the NM landslide and the
762 mudslides of Mt. Rainer and Hólmavík: i) the scar and the erosion zone share the same irregular
763 characteristics, ii) the landslides occur on a continuous low slope and iii) in the deposit zone there are
764 levees, ridges perpendicular to the direction of the flow, few large blocks and a multi-lobate front. The
765 difference in thickness between the martian landslide and the terrestrial landslides can be accounted
766 for by the difference in volume between the landslides (Tables 1 and 2).

767 The presence of lateral levees on its own is not diagnostic. Lateral levees in landslides can be produced
768 by a variety of different mechanisms, as detailed by Corominas (1994). Indeed, the flow can cause
769 basal erosion, which results in lowering of the centre of the sliding mass, leaving lateral levees on either
770 side of the landslide. On Earth, lateral levees are a common signature of earthflows and debris flows
771 (Baum et al., 2014; Nereson and Finnegan, 2015). Earthflows are mainly composed of clays and contain
772 water which plays an important role in the landslide's mobility (Baum et al., 2014). Levees can also be
773 produced in dry granular landslides so do not necessarily indicate the presence of clay and/or water as
774 shown by Mangeney et al. (2007) and Félix and Thomas (2004). Here we observe lateral levees with
775 exterior slopes of ~25°, quite similar to those observed on pyroclastic flow deposits in Chile, between
776 20° and 25° (e.g., *Figure 9* of Jessop et al., 2012) whereas granular materials in laboratory tend to have
777 lower slopes (e.g., Félix and Thomas, 2004). The higher levee angle observed for these pyroclastic flows

778 has been suggested to be linked to the high polydispersity of the material involved. For the NM
779 landslide, it could be the presence of clays or polydisperse granular materials that cause these high-
780 standing steep levees, but given the other morphological similarities to mudslides we favour the
781 presence of clays.

782 The morphology of mudslides on Earth is controlled by the viscous deformation of the substrate, which
783 is in turn controlled by the degree of water saturation of the interstitial environment and the substrate
784 mechanical properties (Comegna et al., 2007). To form a mudslide, clay size grains and liquid water are
785 needed (Vallf-Jo, 1979) and involve variable proportions of water and clay minerals in the case of
786 earthflows. This raises the question of the importance of clays in the formation of the NM landslide.

787 Near infrared orbital spectra of the landslide region show the presence of clay minerals likely formed
788 by hydrothermal activity (Mangold et al., 2007; Ehlmann et al., 2009; Michalski et al., 2010; Viviano et
789 al., 2013), although locations have showed possible weathering through pedogenetic alteration as well
790 (Gaudin et al., 2011). Clay minerals corresponding to Fe/Mg-phyllosilicates are observed on the
791 bedrock of the impact crater rim where the NM landslide occurs (see supplementary material, *S1-
792 Spectral analysis, Figure.S1*). However, the landslide deposit itself does not exhibit any clear spectral
793 signature of clays and it seems to be associated with a light-toned unit mantling this clay-rich bedrock
794 rather than to the bedrock itself. Thus, even if the occurrence of clay minerals in the landslide deposit
795 should favour mudslide development, no clear relationship can be demonstrated between the clay-
796 rich bedrock and the material mobilised by the landslide. As mudslides on Earth necessarily involve
797 liquid water, the morphological similarities with the NM landslide suggest liquid water might also be
798 involved at this location. However, the dating of this landslide to the late Amazonian means it occurred
799 during a period when liquid water is expected to be rare at the surface of Mars, as discussed further
800 below.

801 Given the morphological comparison between the landslides of CCh, ChrC and NM and their respective
802 terrestrial analogues, we classify them into two distinct categories. CCh and ChrC are found to be most
803 similar to landslides that fall into the rockslide category and NM in the mudslide category. We used
804 these morphological constraints to inform the numerical modelling of the CCh/ChrC and NM
805 landslides.

806 **4.2.2. Numerical modelling**

807 **Capri Chasma/Chryse Chaos simulations.** The difference observed between our simulation and the
808 observed deposit (i.e., the final shape of the deposit) could be partly due to the shape of the
809 reconstructed erosion zone that, being partly covered by deposits, was hard to constrain accurately.
810 Tests using a flatter base within the erosion zone do lead to more deposited mass in this zone, as
811 illustrated in the Supplementary material (*section S4- Topographic reconstructions*), but not as much
812 as is observed.

813 The lack of deposits in the erosion zone and on the sloping terrain where the levees are deposited
814 suggests that the best-fit friction coefficient models a landslide that is too mobile (a lower friction
815 coefficient than the one considered in the simulations).

816 The absence of levees means that the deposits spread out more than observed and that the conditions
817 for the formation of such levees have not been satisfied. These differences between the model and
818 the observed deposits do not invalidate the granular flow (rockslide) hypothesis, but highlight that this

819 model makes assumptions which prevent some important complexities being considered. As this
820 model has been successfully used for larger landslides in Valles Marineris, it is likely that these
821 complexities become dominant at smaller scales. For example, at smaller spatial scales the timescale
822 of the mass release becomes more important with respect to the timescale of the sliding.

823 In SHALTOP the eroded mass is released instantaneously, but levees are observed in experiments only
824 when supply is continuous and not from instantaneous collapse experiments (e.g., Félix and Thomas,
825 2004). Besides, the height of the levees simulated with the Pouliquen and Forterre's law has been
826 shown to be too small compared to granular flow experiments (Mangeney et al., 2007; Rocha et al.,
827 2019). This type of landslide seems to follow a sliding plane, so a multi-layer model may be more
828 appropriate as it is able to simulate the heterogeneity in the vertical direction (Fernandez-Nieto et al.,
829 2016).

830 Comparison of our results with laboratory experiments on granular flows (Pouliquen and Forterre,
831 2002) suggest that the dynamics could be quite different if the presence of an erodible bed were
832 included in the simulations, in this case the internal deformation may play a less important role.
833 Pouliquen and Forterre (2002) noted the difference between the spreading of a granular cap over a
834 rigid bed (*Figure 5a* of Pouliquen and Forterre, 2002) and over an erodible bed. With an erodible bed,
835 there was a steeper front and the flowing mass had a more 'croissant' shape (Pouliquen and Forterre,
836 2002, see also Mangeney et al., 2007), which are attributes expressed by our observed deposits. The
837 development of a steeper front in the presence of an erodible bed is also observed for simulations at
838 the field scale (Moretti et al., 2012) (their *Figures 2c* and *2f*). The erodible bed on Mars could be
839 provided by the talus slope over which the landslides propagate.

840 ***Nilosyrtis Mensae simulation.*** Despite the fitting of the centre of mass and runout, the model provides
841 a poor fit for the observed morphology for both of the applied friction laws. We observed that a large
842 proportion of mass remains in the erosion zone for both simulations that is not observed on the real
843 deposit. These observations suggest that NM landslide is less mobile than the model predicts.

844 The failure of the model to reproduce the morphology does not necessarily invalidate the granular
845 flow mechanism for this landslide, but the aspects where the model fails suggests that this landslide is
846 not behaving like a granular flow. Firstly, the fact that the deposits do not follow the steepest line of
847 descent suggests they are momentum dominated. Second, the mass being released from a low-slope
848 fracture zone suggest, given its morphology, some cohesion within the deposits, which is not predicted
849 by the granular flow model.

850 In general, the dry granular flow laws in SHALTOP cannot reproduce satisfactorily any of these martian
851 landslides, but produces a closer fit for the landslides interpreted as rockslides. We suggest that
852 possible improvements to the model, such as: gradual release of the mass, inclusion of an erodible bed
853 and/or adjusting the law used to model the rheology of the landslide may result in better fits and could
854 be the object of future work.

855 **4.3. Formation scenario**

856 On Earth, landslides can be triggered by various phenomena, such as earthquakes (Meunier et al.,
857 2007), heavy rainfall (Wang et al., 2002) or melting permafrost (Niu et al., 2015). We will not consider
858 liquid water as a factor in the landslides of CCh and ChrC, as their morphology is not compatible with
859 its involvement. We therefore discuss the possible role that seismic shaking could have had in

860 triggering these recent landslides. For the NM landslide, we will consider among other things the
861 potential roles of liquid water and seismic shaking as triggering factor(s).

862 **4.3.1. The recent formation of Capri Chasma and Chryse Chaos landslides**

863 On Earth, seismic activity can cause landslides (e.g., Strecker and Fauque, 1988; Bommer and
864 Rodríguez, 2002; Chen et al., 2006; Chigira et al., 2010; Sepúlveda et al., 2010). Earthquakes can
865 destabilize the hydrogeological environment and after repeated earthquakes can cause landslides
866 (e.g., Sassa et al., 2007; Walter and Joswig, 2008; Sæmundsson et al., 2018). A seismic origin has also
867 been considered for martian landslides (e.g., Schultz, 2002; Quantin et al., 2004b). The source of the
868 seismic activity could be tectonic: evidence for recent crustal seismicity in the zone of our landslides
869 comes in the form of faults cutting through the large landslides in Valles Marineris, which are <1 Ga
870 (Quantin et al., 2004a), and wrinkle ridges and/or blind faults (e.g., Schultz, 2000) which are found
871 within 30 km of the CCh and ChrC landslides. In addition, recent results from the InSight instrument
872 Seismic Experiment for Interior Structure (SEIS, (Lognonné et al., 2019) provide evidence for
873 marsquakes, probably related to upper crustal structures (Giardini et al., 2020).

874 Alternatively, seismic shaking could be caused by meteorite impacts. Teanby and Wookey (2011)
875 estimate that an earthquake with a magnitude between 3.9 and 4.5 could be generated by a meteorite
876 impact forming a crater with a diameter between 617 and 1280 m. The SEIS instrument on the InSight
877 lander has not yet been able to link seismic detection with certainty to a meteorite impact, but a study
878 by Wójcicka et al. (2020) estimates that the instrument would be capable of detecting an impact of
879 more than 10 m in diameter within a radius of 400 km. At least five impact craters with a diameter up
880 to 617 m and with preserved ejecta (indicating a young age) have been identified around the landslides
881 but it is currently not possible to directly link the formation of these impact craters to the formation
882 of one of our landslides.

883 Given the evidence, we conclude that seismic shaking from a nearby meteorite impact, or from a
884 crustal marsquake are equally likely triggers for these landslides.

885 **4.3.2. The recent formation of Nilosyrtis Mensae landslide**

886 We have identified morphological similarities between the NM landslide and terrestrial mudslides,
887 including the lobate margins, lateral levees and ridges perpendicular to the flow direction. On Earth,
888 mudslides are caused by water saturation of the surface materials. Saturation causes a loss of cohesion
889 by increasing pore pressure and increases the weight of the materials, allowing them to flow on
890 relatively low slopes (e.g., Chandler, 1972). Saturation can occur throughout, causing deformation *en*
891 *masse* of the material, or can occur in a particular layer, causing the material above it to slide (e.g.,
892 Comegna et al., 2007). On Mars, the NM landslide is dated to a maximum of $\sim 4.7 \pm 4$ Ma and thus
893 formed near the end of the Late Amazonian period (3.1 Ga to present). It is generally accepted that
894 this period is dominated by hyper-arid conditions hostile to the existence of surface liquid water
895 (e.g., Baker et al., 1991; Ehlmann et al., 2011). The question is therefore: how could water have
896 contributed to the formation of the landslide? As above, we first examine the potential role of
897 seismicity, which can play a role in bringing water to the surface. We then discuss climatic factors and
898 geothermal heat flow.

899 **Seismic activity.** On Earth, earthquakes can trigger mudslides via two main mechanisms: liquefaction
900 and aquifer perturbation with saturation of unstable formations (e.g., Binet et al., 2007; Marc et al.,

901 2015). For the NM landslide, we searched for evidence of recent tectonic activity or meteorite impacts
902 that may have triggered the landslide. The Nili Fossae region is known for its extensive fault system,
903 but we found no convincing evidence for recent faults or recent activity near the host crater. We found
904 three young impact craters (ranging in diameter from 2,500 to 3,700 m) but all appear to be somewhat
905 degraded. It therefore seems unlikely that seismic activity played a role in forming the landslide.

906 ***Influence of climate.*** The latitude of the NM landslide is just beyond the generally recognised limit of
907 30° for discontinuous ground ice (e.g., Mustard et al., 2001), but considering Mars' frequent changes
908 in obliquity we consider it likely that ice could have been present in the ground at the time of the
909 landslide. Indeed, Mars is thought to have undergone a significant downward shift in average obliquity
910 at ~5 Ma (Laskar et al., 2004), which roughly corresponds to our estimated maximum age of the NM
911 landslide at 4.7±4 Ma. This shift destabilised ice deposits at the mid-latitudes causing them to migrate
912 back towards the poles (Laskar et al., 2004). It has been hypothesised that the thermal disequilibrium
913 caused by such obliquity shifts could melt ground ice in the first metre of the subsurface (Costard et
914 al., 2002), but that melting over a thickness >10 m is extremely unlikely (e.g., Kreslavsky et al., 2008;
915 Mellon and Phillips, 2001). Hence, this mechanism seems inadequate to trigger a landslide, whose
916 erosion zone penetrates to 50 m below the surface (*Fig.5c*).

917 ***Water and ice interaction.*** The impact crater where NM landslide occurs also contain three other
918 landslides with a same morphology which may have a similar formation mechanism. As discussed
919 above, we consider that ice could have been present in the ground when the landslides formed. We
920 then need an external mechanism to melt this ice. Recent work has revealed that basal melting of
921 debris-covered glaciers may be triggered by a locally higher geothermal flow (Gallagher and Balme,
922 2015; Butcher et al., 2017). Clay minerals formed by hydrothermal activity have been previously
923 identify in the Nili Fossae region (Ehlmann et al., 2009; Brown et al., 2010; Viviano et al., 2013). They
924 reveal the presence of geothermal heat flow in the region mostly active during the early to middle
925 Noachian (Ehlmann et al., 2011). The presence of remnant and localised geothermal flux in this impact
926 crater could explain why this type of landslide is not observed elsewhere in the region but this
927 hypothesis needs to be qualified by the fact that the crater itself is at least 3 billion years old and itself
928 is unlikely to represent the source of the recent active heating. Nevertheless, the crater wall is a
929 preferential location for fluids, due to the presence of fractures potentially linking to surface deep
930 aquifers under pressure (e.g., Abotalib and Heggy, 2019).

931 In summary, no option is fully satisfying, but we consider that triggering by ground ice melt, via
932 increased geothermal heat flux, is an attractive hypothesis to further develop in future work.

933 **5. Conclusions**

934 We have studied three martian landslides using high-resolution images and digital elevation models,
935 comparison with Earth analogues and numerical simulations. The aim is to deduce hypotheses of
936 landslide formation on Mars where in situ analysis is generally not possible. Our results show the
937 importance of using morphological comparison between martian and terrestrial landslides to identify
938 key morphologies, combined with numerical modelling.

939 We estimate that these landslides are all very recent, possibly formed less than 20 Ma and their
940 morphological attributes suggest two distinct behaviours. The Capri Chasma and Chryse Chaos
941 landslides are both located in equatorial regions and share similarities in shape and morphometric

942 characteristics with rockslides on Earth. We consider it likely that these landslides were a result of
943 bedrock failure induced by seismic shaking brought about by nearby impacts or crustal Marsquakes.

944 In contrast, the morphology of the Nilosyrtis Mensae landslide is different from the CCh and ChrC
945 landslides and is more similar to terrestrial mudslides. Similarities include the presence of levees,
946 lobate fronts and ridges perpendicular to the direction of sliding. This suggests a role of clay-sized
947 grains, perhaps related to the presence of phyllosilicates in this region, and of recent, local episodes of
948 liquid water release. We hypothesise that this landslide could have been caused by melting of ground
949 ice by locally elevated geothermal heat flux, and further investigations are currently underway in the
950 Nilosyrtis Mensae area to support or refute this hypothesis.

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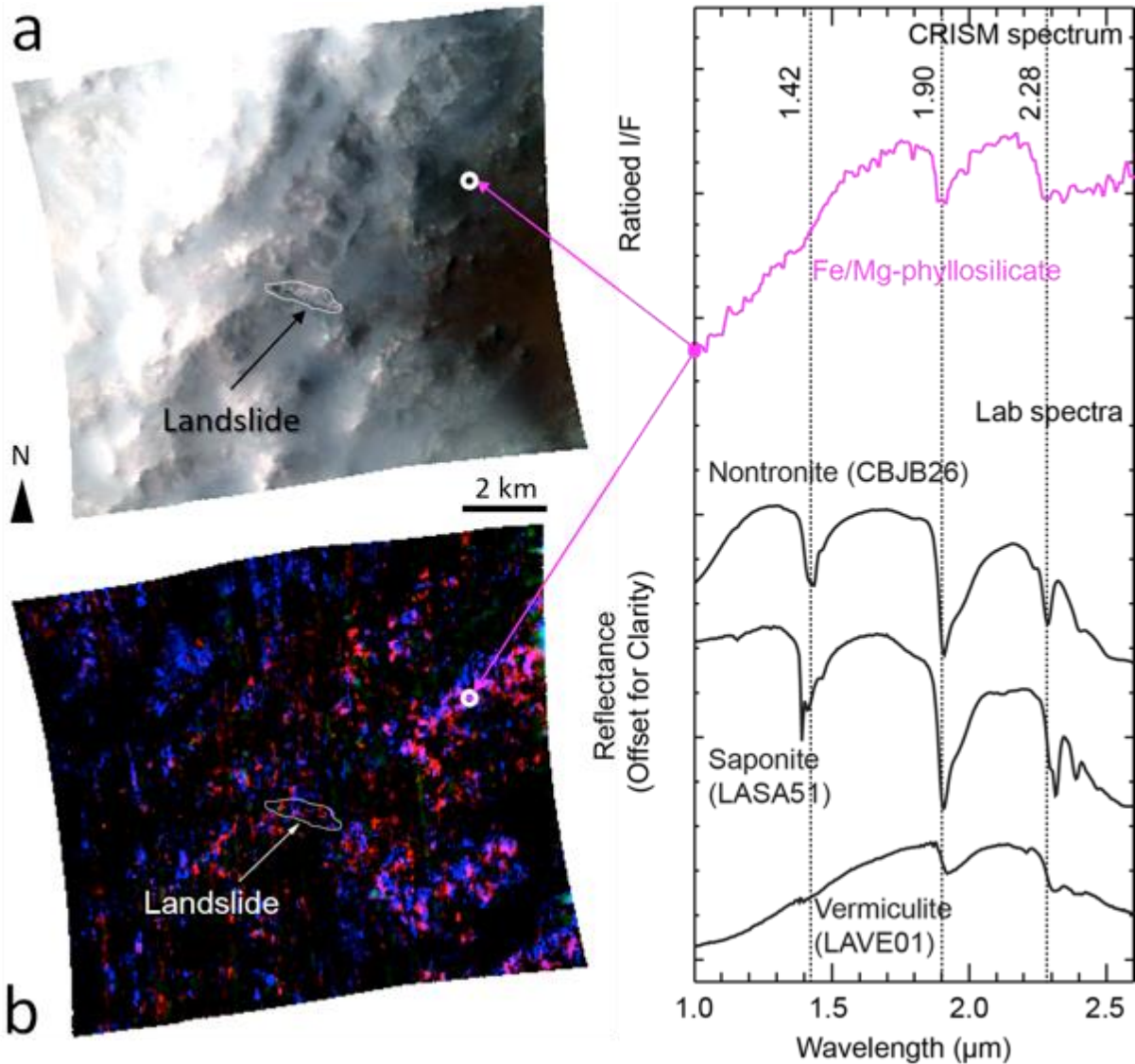
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Supplementary material

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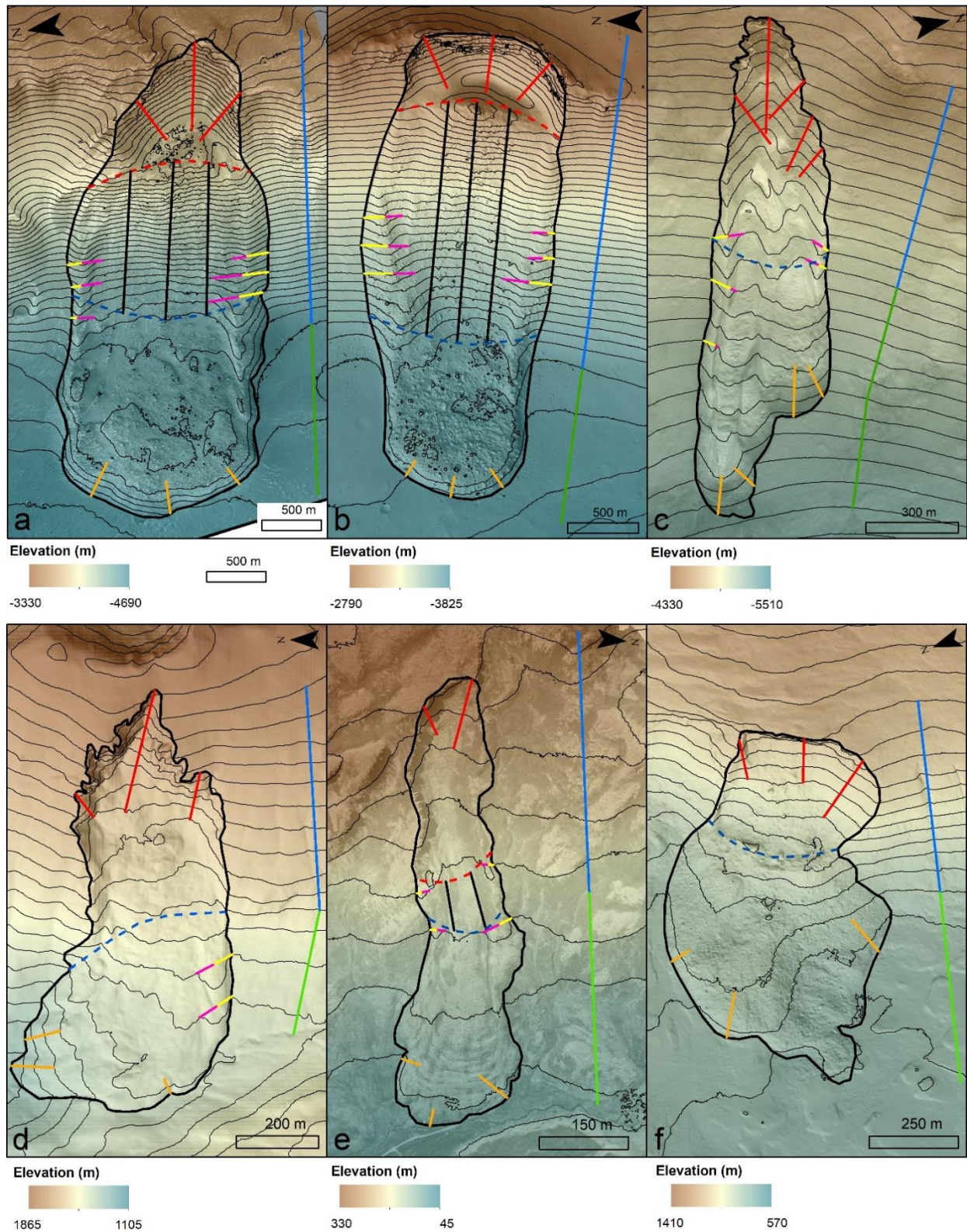
S1-Spectral analysis



1360 *Figure S1 - Fe/Mg clays exposed along the impact crater rim in Nili Fossae region. (a) CRISM enhanced IR colour, R= 2.5 µm;*
 1361 *G= 1.5 µm; B= 1.1 µm. (b) Spectral parameter map, R= 2.3 µm band depth (D2300); G= 2.2 µm band depth (BD2200);*
 1362 *B= 1.9 µm band depth (BD1900R), pink colour: Fe/Mg phyllosilicates, blue colour: hydrated phase. (c) CRISM spectrum of*
 1363 *Fe/Mg-phyllosilicate 'average' of several pixels divided by pixels of a neutral region compared to laboratory spectra (RELAB*
 1364 *library spectra). (CRISM: HRS00024571_07_IF175S_TRR3)*

1365 The hyperspectral CRISM observation presented in Figure S1 was processed using the CAT (CRISM
 1366 Analysis Toolkit; (Morgan et al., 2014)) extension for the ENVI software (Exelis VIS), which was
 1367 developed by the CRISM Science Team. The spectral parameter maps were computed using the
 1368 spectral parameter summary products produced by Viviano-Beck et al. (2014).

1369 Fe-Mg-phyllosilicates (pink tones in Fig.S1b) such as nontronite, saponite, and vermiculite are observed
 1370 on the bedrock along the impact crater rim and on its floor in the region where the Nilosyrtis Mensae
 1371 landslide occurs. Other hydrated phase(s) are also associated with the bedrock. Either this or these
 1372 hydrated phases are not phyllosilicates, or if they correspond to phyllosilicates, then the amount is too
 1373 small for their diagnostic absorption band at 2.3 µm to be detected.



1374 Figure S2 – Position of topographic profiles used to extract measured parameters on Mars (a) Capri Chasma; (b) Chryse Chaos
 1375 and (c) Nilosyrtis Mensae and on Earth (d) Mount Rainier; (e) Hólmanvík and (f) Abisko. Profiles in red, black and orange are
 1376 used for the erosion zone, the transport zone and front scarp slope calculations, respectively. Profiles outside the landslide’s
 1377 boundary in blue and in green are used to estimate the slope angle before the landslide occurred and to estimate the slope
 1378 underlying the deposit zone, respectively. Profiles in yellow and pink are used for outer and inner levee angle calculation,
 1379 respectively. The dashed line in red delimits the erosion zone and the dashed line in blue delimits the deposition zone.
 1380 Background images: HiRISE images for (a) ESP_035831_1760, (b) PSP_005701_1920 and (c) ESP_026781_2075; (d)
 1381 Washington Lidar Portal hillshaded rendering, (e) Aerial image from the Land Survey of Iceland and (f) hillshaded LiDAR data
 1382 from Geographical Sweden Data.

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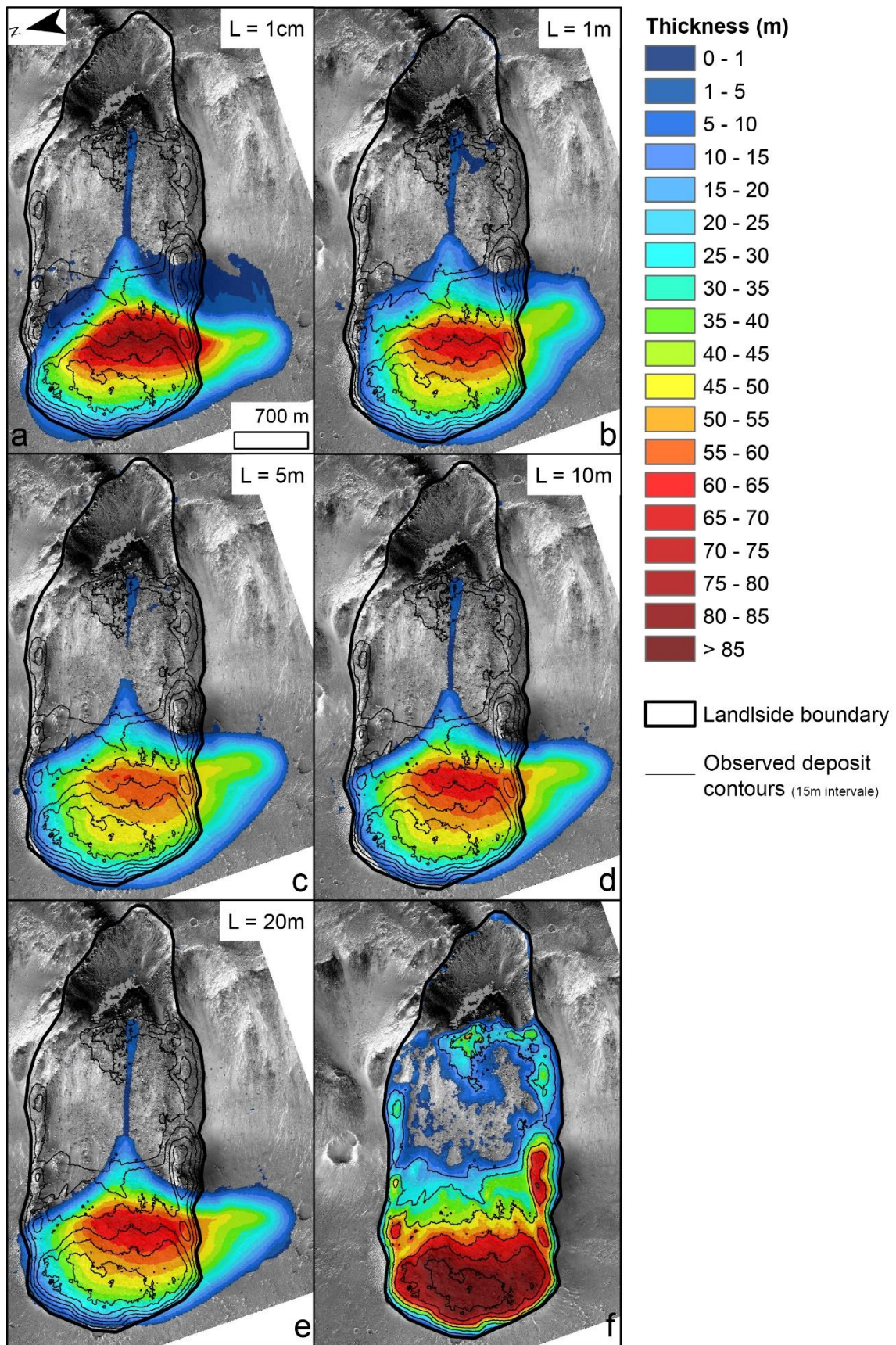
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S2-Topographic profiles

On figure S2, for the three profiles placed for each levee, each cross-section line was placed by connecting with a straight line the points where a given contour line crosses the levee boundary on each side. We isolated each levee by identifying the point at which it rose from the surrounding terrain aided by the image data and extracted the elevation values at these points. We identified and extracted the elevation of the crest of the levee. For each side we calculated the slopes by taking the arctangent of the difference between the base and crest elevations divided by the distance to the crest along the profile. We extracted the maximum height for each levee by taking the maximum difference between the base and crest elevations.

S3-Grain size sensitivity tests

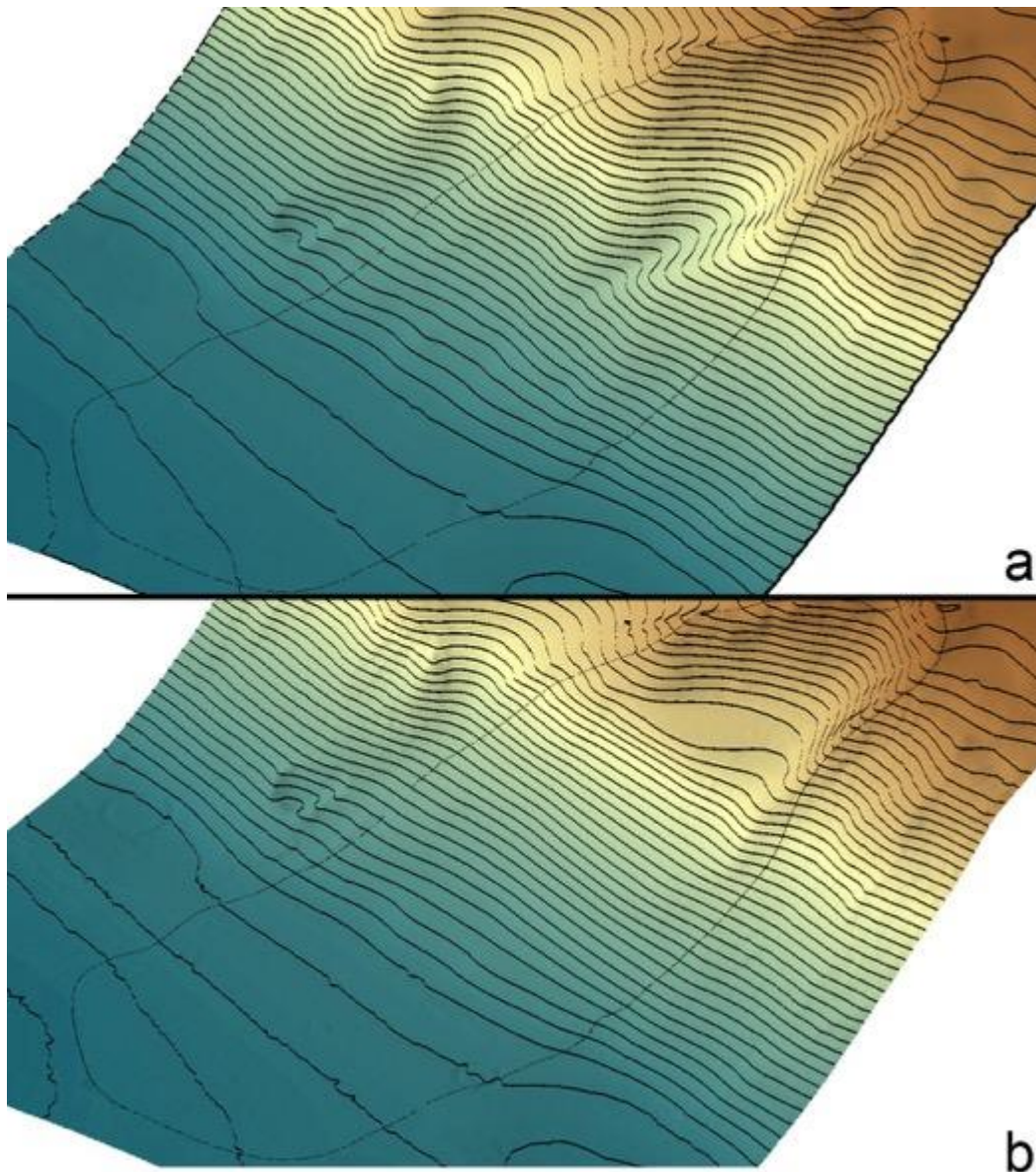
In order to investigate the influence of grain size in the model using the Pouliquen and Forterre’s law, we performed a sensitivity test on the Capri Chasma landslide by taking grain sizes of L= 1 cm, 1 m, 5 m, 10 m and 20 m (Figure S3). The grain size influences the friction coefficients required to correctly fit the runout distance. The smaller the grain size, the smaller the friction angles must be. In addition, the smaller the grain size, the less rounded the landslide deposit front is. The decision to use L= 5 m for CCh and ChrC and L= 1 cm for Nilosyrtis Mensae is based on morphological observations of the deposition surface. On the CCh and ChrC slides, the entire deposition surface is covered with blocks larger than 5 m. This size has therefore been taken as a reference for modelling. For NM landslide, there are only a few scattered blocks of maximum 10 m in size near the toe of the deposit. Also, we have not identified any blocs of one metre or more on the main deposit it’s that why we choose to take our smaller tested value of 1 cm to conduct the NM simulations.



1405 Figure S3 – Final deposit thickness maps resulting from the grain size sensitivity tests applied to Capri Chasma landslide
 1406 simulation with grain size (L) of (a) 1 cm; (b) 1 m; (c) 5 m; (d) 10 m; (e) 20 m. (f) is the observed deposit on Mars.

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S4-Topographic reconstructions

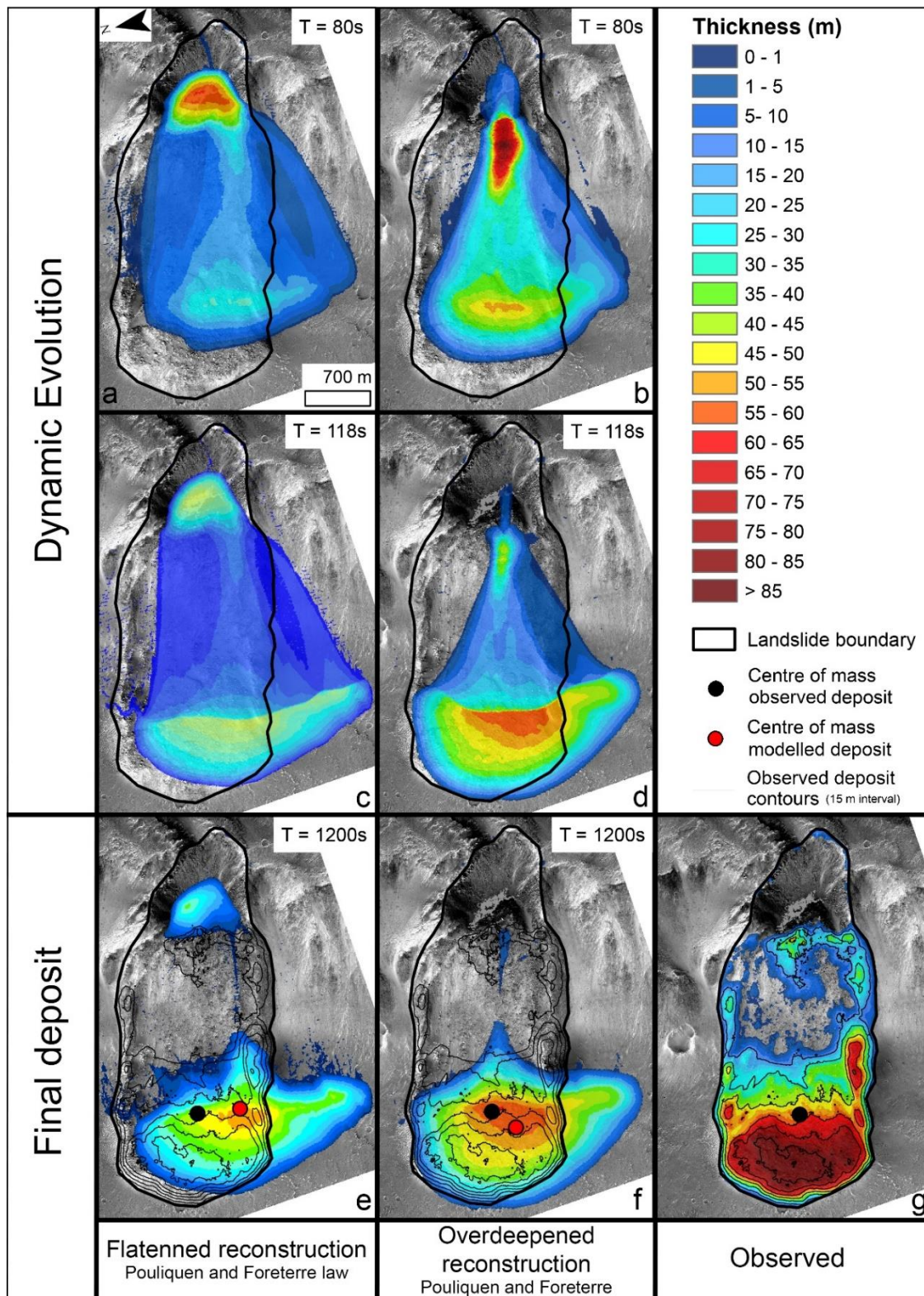


1409 *Figure S4 - 3D view of two alternate topographic reconstructions for Capri Chasma landslide, where the coloured DEM is*
 1410 *overlain by semi-transparent hillshaded relief and 25 m interval contour lines in black. (a) Reconstruction with overdeepened*
 1411 *areas in the erosion zone. (b) Reconstruction with a flattened base in the erosion zone.*

1412 We tested two alternate reconstructions of the initial topography of Capri Chasma landslide to assess
 1413 the sensitivity of the numerical modelling to this parameter. The actual topography of the erosion zone
 1414 is unknown as it is partially covered with deposits. We tested the two end members shown in Figure S4.

1415 The reconstruction in Figure S4a includes an over-deepening at the centre of the erosion zone and in
 1416 Figure S4b a flatter topographic base of the erosion zone. The results of test simulations using these
 1417 two reconstructions are shown in Figures S5e for the flattened reconstruction and S5f for the
 1418 overdeepened reconstruction. For the flatter topography (Fig. S5e) some of the mass is retained in
 1419 the erosion zone as is observed for the real deposit, but significant mass is deflected to one side,
 1420 hence we did not choose this reconstruction, as it was a poor fit for the observed deposits. For the
 1421 overdeepened reconstruction (Fig. S5f), we observe almost no deposition at the base of the erosion
 1422 zone and the deposit is located more centrally within the deposition zone compared to the flattened

1423 erosion zone reconstruction. The results are almost indistinguishable from those with no
 1424 overdeepened portion (Figure 11f), hence for simplicity we decided not to use this reconstruction as
 1425 the presence/absence of this overdeepened section was impossible to determine.



1426 Figure S5 - Dynamic evolution of the simulation using flattened topographic reconstruction of the erosion zone (figure S4b),
 1427 (a) at T=80s, (c) at T=118s and (e) at T=1200s when the flow is stabilised and using the overdeepened reconstruction (b) at
 1428 T=80s, (d) at T=118s and (f) at T=1200s when the flow is stabilised compared to (g), the actual deposits. Pouliquen and
 1429 Forterre's law used with angles of $\delta_1=8$, $\delta_2=18$ and $\delta_3=11^\circ$.

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