

Dynamics of recent landslides (

A. Guimpier, Susan J. Conway, A. Mangeney, A. Lucas, N. Mangold, M. Peruzzetto, M. Pajola, A. Lucchetti, G. Munaretto, T. Saemundsson, et al.

▶ To cite this version:

A. Guimpier, Susan J. Conway, A. Mangeney, A. Lucas, N. Mangold, et al.. Dynamics of recent landslides (

HAL Id: hal-03287039 https://hal.science/hal-03287039

Submitted on 15 Jul 2021

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers. L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1	Dynamics of recent landslides (<20 My) on Mars: Insights from high-resolution
2 3 4 5	topography on Earth and Mars and numerical modelling A. Guimpier ^{1,2} *, S. J. Conway ¹ , A. Mangeney ³ , A. Lucas ³ , N. Mangold ¹ , M. Peruzzetto ^{3,4} , M. Pajola ⁵ , A. Lucchetti ⁵ , G. Munaretto ^{5,6} , T. Sæmundsson ⁷ , A. Johnsson ⁸ , L. Le Deit ¹ , P. Grindrod ⁹ , J. Davis ⁹ , N. Thomas ¹⁰ , G. Cremonese ⁵
6	
7 8	¹ Laboratoire de Planétologie et Géodynamique, UMR6112 CNRS, Université de Nantes, France ² Université Paris-Saclay, France
9	³ Université de Paris, Institut de physique du globe de Paris, CNRS, F-75005, Paris France
10	⁴ Bureau de recherches géologiques et minières, Orléans, France
11	⁵ INAF - Astronomical Observatory of Padova, Padova, Italy
12	⁶ University of Padova, Padova, Italy
13	⁷ University of Iceland, Reykjavík, Iceland
14	⁸ University of Gothenburg, Gothenburg, Sweden
15	⁹ Natural History Museum, United Kingdom
16	¹⁰ University of Bern, Bern, Switzerland
17	
18	*corresponding author <u>anthony.guimpier@univ-nantes.fr</u>
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 preocupation 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 landslides with volumes greater than 10¹⁰ m³, which is larger than landslides most commonly found on 54 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.

To our knowledge, no studies have specifically focussed on understanding 'small' martian landslides with a volume less than 10¹⁰ m³, which have a similar scale to landslides that can be found on Earth. These common terrestrial landslides are well-studied and their formation mechanisms are better understood than that of their larger counterparts. This provides an opportunity to perform a 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.

In addition to this comparative morphological study, we use the thin-layer numerical code SHALTOP to simulate the landslide dynamics, assuming it is a dry granular flow. In spite of their simplifying assumptions and the uncertainty on initial and boundary conditions (see Section 2.3, and Delannay et al., 2017), thin-layer numerical models have previously been successful in reproducing the runout and approximate deposit morphology for a wide range of landslides on Earth and Mars. Using seismic data to reconstruct the dynamics of some terrestrial landslides, it was shown that thin-layer models can

- also reproduce these dynamics (Moretti et al., 2012, 2020; Levy et al., 2015; Yamada et al., 2016, 2018).
- We therefore employ a double approach, using both morphological and numerical methods, to better
 constrain the mechanism of formation of these 'small' martian landslides and to understand their
 dynamics and hence, the potential role of liquid water and/or active tectonics.

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

86 **2. Methodology**

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

91

2.1. Geomorphological analysis

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

95 2.1.1. Datasets for martian landslides

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

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

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

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

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

114 **2.1.2.** Datasets for terrestrial landslides

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

122 For the landslide located near Abisko in Sweden at 68°12' N, 19°2' E, a 2 m/pixel DEM derived from airborne laser altimetry was provided by the Swedish Land Survey, Geographical Sweden Data (GSD)-123 Elevation Data. For the landslide located near Seattle and Mount Rainier in Washington State in the 124 125 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 Hólmavík 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 the commercial software AgiSoft Photoscan multiview photogrammetry resulting in a final DEM with 129 a ground sampling of 60 cm/pixel. 130

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.



Figure 1 – Contour lines in red are derived from the original HiRISE DEM of each martian landslides of this study and contour
lines in yellow are estimated pre-landslide contours. All these contour lines have an interval of 25 meters. (a) Capri Chasma
landslide, CCh, HiRISE: ESP_035831_1760; (b) Chryse Chaos landslide, ChrC, HiRISE: PSP_005701_1920; (c) Nilosyrtis Mensae
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 placement of the reconstructed contours and connected them manually to the two intersections of 160 161 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 at 2 m intervals, attributed with the contour elevation value. These point features and associated 164 elevations were then gridded into a 'reconstructed DEM' using the ArcGIS Natural Neighbour 165 interpolation algorithm. The difference between this reconstructed DEM and the original DEM results 166 167 in a thickness map, with positive values indicating areas of deposition and negative values indicating erosion. This reconstruction was also used to estimate the volume of deposition zone of the landslide. 168 169 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 values across the whole width of the landslide, the "transport zone" is where both positive and 173 negative values are found across the width and where predominantly positive values are found across 174 175 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 landside 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

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.

- the height and width of each levee, as well as the slope between the base and the crest of the levee.
 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
- 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.
- We extracted one elevation profile ~300 m outside the boundary landslide (*Figure S2*), and perpendicular to the contour lines, in order to estimate the slope angles of the terrain before the landslide occurred and the slope angle of the deposit zone. This profile extended from the top to the bottom of three zones defined above. Here we take only one profile, because the further profiles are placed from the landslide, the less likely they are to be representative of the pre-landslide surface. For each zone we calculated the slopes, with the methodology as described above.
- For each zone of the landslide, we report the means and ranges of the slope values calculated fromthe 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

- Craterstats2 (Michael and Neukum, 2010; Michael et al., 2012). This method exploits models which
 describe how the bolide production function varies over time, enabling the size-frequency distribution
 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).
- In the deposition zone (described in section 2.1), we digitized craters to compute their distribution in size. The deposit zone area and the size-frequency distribution of the superposed craters are then used as inputs for Craterstats2 (Michael and Neukum, 2010; Michael et al. 2012). Due to the small surface area of the landslides studied here and the low number of impact craters on their surface, there are large margins of error in these estimations (e.g., Warner et al., 2015).

221 **2.3.** Numerical modelling

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

227 **2.3.1. Model input preparation**

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

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

- thickness value. The x and y coordinates of the centre of mass is calculated according to the average x
- 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

257 SHALTOP, where the estimated pre-landslide surface is combined with the present-day erosion zone.

254

258 **2.3.3.** SHALTOP description

SHALTOP is a numerical model that simulates homogeneous flows propagating on complex topography 259 260 using the thin-layer approximation (that is, simplifications to the governing equations that can be used when the flow thickness is small in comparison to its lateral extent) (Bouchut et al., 2003; Bouchut and 261 262 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 most of the depth-averaged landslide models, SHALTOP accounts for the curvature tensor of the 265 266 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 (Lucas et al., 2007; Favreau et al., 2010; Moretti et al., 2012, 2015, 2020; Brunet et al., 2017; Yamada 269 270 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.

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

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

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

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

- Several regimes, low-velocity and high-velocity, are described depending on the Froude number (*Fr*) defined as $\frac{u}{\sqrt{ab}}$.
- 292 If $F_r \ge \beta$

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

- 294 If $F_r = 0$
- 295 $\mu(h) = \mu_{start}(h) = \tan \delta_3 + (\tan \delta_2 \tan \delta_1) \frac{1}{1 + \frac{h}{L}}$ (2),

296 If $0 \le F_r \le \beta$

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

298 Where

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

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

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



311 Figure 3 - Location of the three studied landslides on a colourised MOLA topographic map of Mars with semi-transparent

312 shaded relief (Smith, et al., 2001).



314 **3. Results**

In this section we present the geomorphological results (Section 3.1), the age estimates (Section 3.2)
and numerical modelling (Section 3.3). Section 3.1 is divided into two parts, the first concerning the

description of the results for the martian landslides followed by a comparison to the terrestrial landslides.

319 **3.1. Geomorphological results**

320 **3.1.1. Martian landslides**



Figure 4 - The Capri Chasma (CCh) landslide. (a) Regional view of the area around the CCh landslide, which is outlined by the
black rectangle indicating the location of panel (b), the orange arrows indicate three larger landslides located on the other
side of the outflow channel. The background is a mosaic of CTX images P20_008707_1757, P07_003578_1757,
G21_026271_1756, G03_019546_1753, and D06_029515_1755. (b) CaSSIS image of CCh landslide (MY35_008462_188, PANBLU filters). (c) & (d) Detailed view of the landslide deposit, black arrows indicate windblown deposits, white arrows indicate
boulders more than 30 m in diameter, red dotted line indicates the inner edge of the southern lateral levee (HiRISE:
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 of north-south trending canyons at the eastern extent of Capri Chasma, a canyon leading to the outflow 330 331 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 middle Noachian highland unit 'mNh', (see Tanaka et al., 2014). Tanaka et al. (2014) describe this unit 333 334 as featuring degraded to severely degraded undifferentiated materials resulting from meteorite impacts, volcanic flows, and possibly sedimentary and fluvial deposits and is dated to between 4.5 and 335 336 3.7 billion years old.

The landslide faces west, measures ~4 km long and ~1.7 km-wide at its widest point. Its key attributes 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
- 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.

Ma	rtian 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	
	Maximum length (m)	1000	500	500	750
	Maximum width (m)	1080	1200	200	220
Erosion zone	Aspect ratio (length/width)	0.9	0.4	2.9	3.4
	Area (m²)	1.4x10 ⁶	1.0x10 ⁶	6.4x10 ⁴	1.2x10 ⁵
	Steepest slope (°)	52	70	40	
	Mean slope (°)	30	35	30	
	Maximum length (m)	1350	1700		-
Transport zono	Maximum width (m)	1685	1350		-
Transport zone	Mean slope (°)	26	23	-	
	Mean adjacent slope (°)	25	23	25	
	Maximum length (m)	1620	1050	590	830
	Maximum width (m)	1700	1000	210	190
	Area (m²)	3.2x10 ⁶	2.0x10 ⁶	1.2x10⁵	1.7x10 ⁵
Denosit zone	Volume (m ³)	1.4x10 ⁸	4.0x10 ⁷	2.1x10 ⁶	2.4x10 ⁶
Deposit zone	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	2	0
	Mean height (m)	45	20	20	
Levee	Range (and mean) of lateral angle (°)	9 - 15 (12)	8 - 10 (9)	22 - 30 (26)	23 - 25 (24)
Maximu	ım boulder size (m)	40	60	10	

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

At the base of the erosion zone, there is a slope reversal of ~18° where some of the failed material has 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

1° near the deposit zone (*Fig.5a, between blue and red dashed lines*). In this zone low lateral levees are

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





Figure 5 – Colourised digital elevation model with semitransparent ortho-image overlain 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) Nilosyrtis 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
- 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
- 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 is underlain by the 'Hto' unit, a transition valley unit dated to the Hesperian (3.56 – 3.24 Ga), composed 395 396 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).

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

- relatively short compared to the total length of the landslide (*Fig.5b*; *Table 1*). As for the CCh landslide,
 the base of the erosion zone is characterised by a slope reversal of ~20° 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
- thickness gradually increases toward the toe to reach a maximum thickness of about 64 m (*Fig.5b*). We
- observe fluctuations in slope angle on the deposit zone that indicates a very high surface roughness,
 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
- 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*).



Figure 7 - The Nilosyrtis Mensae (NM) landslide. (a) Regional view of the area surrounding the NM landslide. Black rectangles
indicate the positions of panels b, d and e. CTX images: D04_028680_2064, B16_015942_2088, B11_014109_2058,
G01_018777_2091, G18_025225_2058, P06_003310_2082, and B03_010707_2080. (b) CaSSIS colour image of the Nilosyrtis
Mensae landslide (MY35_008751_028, PAN-BLU filters). (c) Detailed view of the NM landslide in (b), where white arrows
indicate boulders more than 5 m in diameter, and blue arrows indicate perpendicular ridges on the deposit (HiRISE:
ESP_026781_2075). (d) & (e) Other landslides in same impact crater on HiRISE images ESP_053457_2075 &
ESP_057110_2075, respectively. Credits: NASA/JPL/U0fA/MSSS/ESA/Roscosmos/UniBe.

424

425 Nilosyrtis Mensae landslide. The Nilosyrtis 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 Nilosyrtis 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 presence of fluvial erosion and deltaic deposits which date back to the Late Noachian, between 3.85 433 434 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).

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

maximum value of 40°. There are tracks left by rolling blocks (e.g., Tesson et al., 2019), particularly in
 the southern erosion zone. No slope breaks are observed between the erosion zone and deposit zone

443 (Fig.5c).

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

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

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

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

Terre	strial 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
	Maximum length (m)	340	340	550
	Maximum width (m)	440	150	350
Erosion zone	Aspect ratio (length/width)	0.7	2.2	1.5
	Steepest slope (°)	85	65	70
	Mean slope (°)	39	33	43
	Maximum length (m)	-	100	-
Transport zono	Maximum width (m)	-	140	-
Transport zone	Mean slope (°)	-	11	-
	Mean adjacent slope (°)	27	15	24
	Maximum length (m)	519	295 - 320	435 - 470
	Maximum width (m)	510	150	350
	Area (m²)	1.2x10 ⁶	5.7x10 ⁴	1.6x10 ⁶
	Volume (m ³)	3.5x10 ⁶	3.3x10 ⁵	2.2x10 ⁶
Deposit zone	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
	Mean height	-	10	50
Levee	Mean (and range) of lateral angle (°)	-	13 - 21 (17)	16 - 19 (17)
Maxim	um block size (m)	30	6	20

495

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

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

The erosion scar is sharp and well defined (*Fig.8c*), with the steepest values near the scar (85°
in Abisko, compared to 70° in ChrC, *Fig.5b* and 52° for the CCh landslide, *Fig.5a*, *Table 1 & 2*)
and rapidly descending to ~39° on the talus slope below.

5012. The length/width ratio of the erosion zones are similar, 0.4 for the ChrC landslide, and 0.7 for502the Abisko rockslide. However, in the CCh landslide the length/width ratio is 0.9. In Abisko503there 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 the505lower slope.

- The deposit areas for the Abisko rockslide, and ChrC and CCh landslides are covered by blocks of tens of metres in diameter. These blocks are highlighted in *Figures 8b* by the white arrows and reach 30 m in diameter. These blocks were also noted by Rapp (1960). The deposit zone is located only on gently inclined topography in all three cases.
- The two main differences between the Abisko rockslide and the ChrC and CCh landslides are their scale and a difference in the pre-existing topography. Rapp (1960) estimated that the Abisko landslide has a 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 less than the volume of CCh and ChrC (*Table 1*). At Abisko, the adjacent hillslope has a slope of 27° (*Table 2* and *Fig.8c*), compared to 25° and 23° on Mars, for CCh and ChrC. There is a lack of a transport zone and its associated levees in the Abisko rockslide. The planview shape and mass distribution of the 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 - Hólmavík mudslide (Iceland) and Mount Rainier mudslide (US). (a) Elevation difference map overlain on colourised 519 DEM with semi-transparent hillshaded relief for the Hólmavík mudslide. Topographic contours with 25 m interval are marked 520 in black. (b) Hillshaded relief of the Hólmavík mudslide. (c) Elevation profile and slope angle variation of the Hólmavík 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 **Nilosyrtis 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:

- In both the terrestrial mudslides and in the NM landslide, the erosion scar is sharp and has an irregular outline (*Fig.9b, 9e*). The erosion zone for Hólmavík and NM has an elongate shape.
 The length/width ratio of the erosion zone for Hólmavík is 2.2 and for NM landslide it is 2.9 and 3.4 for northern and southern parts, respectively. For comparison the Mt Rainier mudslide 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.
- 5423. The terrestrial mudslides and the NM landslide have ridges perpendicular to the flow direction.543These ridges are particularly well developed in the Hólmavík mudslide and are weakly present544in the Mt Rainer mudslide. In Hólmavík they have a height of 11 m (*Fig.9b, blue arrow*) and in545Mt 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,5472003) and submarine landslides (Hildenbrand et al., 2006; Masson et al., 2002). Neither the548terrestrial mudslides nor the NM landslide have abundant blocks at their surface.
- Lobate margins. The terrestrial mudslides and the NM landslide all have multi-lobed terminal
 margins to their deposits. The relief of the margins is 12 m at Hólmavík and more than 33 m
 at Mt Rainier, compared to 34 m at NM (*Table 1, 2*).
- 5. Lateral levees. The terrestrial mudslides and the NM landslide possess lateral levees found in
 the transport zones of the landslides. The levees lateral slopes are 13° 21° (*Table 2*)
 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

For the CCh landslide, we identified 12 impact craters on the deposit, and the largest of these craters has a diameter of 35 m. We used the crater-size frequency distribution of this landslide to estimate its age to be 13±5 Ma (*Fig.10a*). To obtain this estimate, we used only the 7 largest impact craters with a 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



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

to identify in the images (10 m), a method suggested by (Hartmann, 2005). The estimated maximum
age of formation of ChrC is ~4.5±4 Ma (*Fig.10b*), whereas for the NM landslide the maximum age is
~4.7±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**

A summary of the friction coefficients used in the simulations along with the best-fit results are presented in Table 3. We used *L*= 5 m as the mean grain size for CCh and ChrC and *L*= 1cm for NM, based on the observed maximum block size (*see section 3.1.1*) and chosen after some sensitivity tests (*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 landsides, 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 $\delta = tan^{-1}(0.27)$

for ChC and $\delta = tan^{-1}(0.30)$ for ChrC. The centre of mass of the simulated deposits is near the base

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

the model, *Fig.11f* and 112 m observed, *Fig.11g*) and in ChrC the model overestimates the maximum

		Model parameters					
Landslide	Initial mass (m³)	Tort	Pouliquen and			Coulomb's	Empirical Value
Landshac		rup	Forterre's law			law	μ_{eff} =V ^{-0.0774}
		Tun	$tan(\delta_1)$	$tan(\delta_2)$	tan(δ₃)	tan(δ)	
		1	0.12	0.30	0.16	0.23	
		2	0.14	0.32	0.18	0.25	
Count		3	0.16	0.33	0.19	0.26	
Capri	9.7x10 ⁷	4	0.18	0.35	0.21	0.27	0.24
Chasina		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	
	6.2x10 ⁷	1	0.16	0.33	0.19	0.27	
		2	0.18	0.35	0.21	0.28	
		3	0.19	0.36	0.22	0.29	
Chryse		4	0.19	0.37	0.23	0.30	0.24
Chaos		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	
	tis ae 4.2x10 ⁶	1	0.32	0.50	0.36	0.34	
		2	0.34	0.52	0.38	0.35	
		3	0.35	0.53	0.39	0.36	
Nilosyrtis		4	0.36	0.54	0.40	0.37	0.34
WIEIISae		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	

598Table 3 - Summary of friction coefficient values tested for each of our three landslides using the Pouliquen and Forterre and599Coulomb laws and the effective friction coefficient using the empirical law determined by Lucas et al. (2014) for landslides

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

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.

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

604Table 4 - Comparison between the front deposition angle measured in simulations using Coulomb's and Pouliquen and605Forterre's laws and the real measured deposition angle. The mean angle was determined from three different measurements606taken 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

observed, despite the fact that the Pouliquen and Forterre friction law is capable of producing levees

610 under certain conditions (Mangeney et al., 2007).

- For the NM landslide we observed a poor fit between the two-simulation law (Coulomb, *Fig.13e* and
- Pouliquen and Forterre, *Fig.13f*) and the observed deposit. We used friction angle $\delta = tan^{-1}(0.36)$
- 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
- of mass of the observed deposit. A large proportion of the eroded mass remains within the erosion
- 616 zone for each simulation.
- 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*).



Figure 11 – Modelled and observed deposit thicknesses for Capri Chasma landslide. (a-d) Dynamic evolution of the deposits
for grain size L=5m using Coulomb's law at T=80s (a), T=110s (c) and T=1200s (e) and using Pouliquen and Forterre law (2002)
at T=80s (b), T=118s (d) and T=1200s (f). The times at 1200s represent fully stabilised deposits. (g) The observed deposit. The
simulations focused on reproducing the final morphology of the deposits, so the timesteps during the simulation are provided
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.



632

633 (c) and T=1200s (e) and using Pouliquen and Forterre law (2002) at T=90s (b), T=200s (d) and T=1200s (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**

642

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





Figure 14 - Heim's ratio (H/ΔL') plotted against volume for the martian landslides (coloured dots) and terrestrial landslides
(coloured squares) of this study compared to terrestrial (blue dots, data from Legros et al., 2002, Lucas et al. 2014) and martian
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, 2006; De Blasio, 2011; Brunetti et al., 2014; Airo, 2015). These large landslides have a very large 649 650 deposition area with an average of 10⁹ m² (Quantin et al., 2004a) while our landslides have deposition areas ranging from 10⁵ to 10⁶ m². The large landslides in Valles Marineris often have overlapping layers 651 652 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 the martian and terrestrial landslides from the literature (Legros, 2002; Quantin et al., 2004a; Lucas et 657 658 al., 2014; Brunetti et al., 2014; Crosta et al., 2018) we observe that CCh and ChrC are towards the smaller end of the typical volumes or martian landslides and NM falls outside the martian population 659 660 (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 (Mangeney-664 Castelnau et al., 2005; Mangeney et al., 2010). As described in Mangeney et al. (2010), the constant 665 acceleration resulting from the sum of forces due to gravity (c_0) and friction (m), are defined as $c_0 = \sqrt{kgh_0 \cos\theta}$ and $m = g\cos\theta(tan\theta - tan\delta)$, with k being a constant, θ is the slope 666 inclination and h_0 the initial thickness. The front of the landslide stops when its velocity 667 $u = 2c_0 + mt_f = 0$, with the stopping time (t_f) defined as: 668

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

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

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

Our martian landslides are located along the trend line predicted by Lucas et al. (2014) as calculated 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 landslide. This trend line (grey dashed line, *Fig.14*) was calculated for landslides with mainly dry granular behaviour. As our landslides roughly follow this trend line, this is consistent with them having a dry granular behaviour, but does not exclude other mechanisms. This trend line does not take into account the morphology of the deposits which is another key indicator of the physical processes at 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 corroborated by the lower frequency of large blocks at the surface of the deposits of ChC compared to 685 686 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 presumably older than both CCh and ChrC) located near Montevallo crater (15°;54 °W) (Fig.15a). On 688 this landslide the blocks are less numerous and less visible (Fig.15b), possibly covered by aeolian 689 690 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).



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

703

704

4.2. Emplacement mechanisms

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

from the other two landslides studied. The first difference lies in the shape of the erosion zones. In the case of CCh and ChrC landslide, the landslide scar forms a well-marked single arc, while for NM the

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



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

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

- NM landslide, there are ridges perpendicular to the direction of flow in the deposition zone, which are
- 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.
- The morphological differences indicate different flow dynamics between the CCh/ChrC landslides andthe 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 Abikso (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).
- As described in Section 3.1.2 we observed several key similarities between the NM landslide and the mudslides of Mt. Rainer and Hólmavík: i) the scar and the erosion zone share the same irregular characteristics, ii) the landslides occur on a continuous low slope and iii) in the deposit zone there are levees, ridges perpendicular to the direction of the flow, few large blocks and a multi-lobate front. The difference in thickness between the martian landslide and the terrestrial landslides can be accounted 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

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

The morphology of mudslides on Earth is controlled by the viscous deformation of the substrate, which is in turn controlled by the degree of water saturation of the interstitial environment and the substrate mechanical properties (Comegna et al., 2007). To form a mudslide, clay size grains and liquid water are needed (Vallf-Jo, 1979) and involve variable proportions of water and clay minerals in the case of 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.

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

The absence of levees means that the deposits spread out more than observed and that the conditions for the formation of such levees have not been satisfied. These differences between the model and the observed deposits do not invalidate the granular flow (rockslide) hypothesis, but highlight that this model makes assumptions which prevent some important complexities being considered. As this model has been successfully used for larger landslides in Valles Marineris, it is likely that these complexities become dominant at smaller scales. For example, at smaller spatial scales the timescale of the mass release becomes more important with respect to the timescale of the sliding.

In SHALTOP the eroded mass is released instantaneously, but levees are observed in experiments only when supply is continuous and not from instantaneous collapse experiments (e.g., Félix and Thomas, 2004). Besides, the height of the levees simulated with the Pouliquen and Forterre's law has been shown to be too small compared to granular flow experiments (Mangeney et al., 2007; Rocha et al., 2019). This type of landslide seems to follow a sliding plane, so a multi-layer model may be more appropriate as it is able to simulate the heterogeneity in the vertical direction (Fernandez-Nieto et al., 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.
- Nilosyrtis Mensae simulation. Despite the fitting of the centre of mass and runout, the model provides
 a poor fit for the observed morphology for both of the applied friction laws. We observed that a large
 proportion of mass remains in the erosion zone for both simulations that is not observed on the real
 deposit. These observations suggest that NM landslide is less mobile than the model predicts.

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

4.3. Formation scenario

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

60 Given the evidence, we conclude that seismic shaking from a nearby meteorite impact, or from a 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 ~4.7±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.

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

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

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

933 **5.** Conclusions

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

We estimate that these landslides are all very recent, possibly formed less than 20 Ma and their
 morphological attributes suggest two distinct behaviours. The Capri Chasma and Chryse Chaos
 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 of943 bedrock failure induced by seismic shaking brought about by nearby impacts or crustal Marsquakes.

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

951 Acknowledgments

952 We are grateful to G. Stucky de Quay and another anonymous reviewer for improving the quality of 953 the manuscript with their helpful feedback. S .J. Conway, N. Mangold and A. Guimpier are grateful for 954 the support of the Programme National de Planétologie (PNP), the Centre National d'Etudes Spatiales 955 (CNES), the Groupement de Recherche Ecoulements Gravitaires et RIsques Naturels (EGRIN) and 956 A. Mangeney for the ERC contract , ERC-CG-2013-PE10-617472 SLIDEQUAKES. M. Pajola, A. Lucchetti 957 and G. Munaretto have been supported for this study by the Italian Space Agency (ASI-INAF agreement 958 no. 2017-03-17). The authors thank the spacecraft and instrument engineering teams for the 959 successful completion and operation of CaSSIS. CaSSIS is a project of the University of Bern funded 960 through the Swiss Space Office via ESA's PRODEX programme. The instrument hardware development 961 was also supported by the Italian Space Agency (ASI) (ASI-INAF agreement no. I/018/12/0), INAF/ 962 Astronomical Observatory of Padova, and the Space Research Center (CBK) in Warsaw. Support from SGF (Budapest), the University of Arizona (LPL) and NASA are also gratefully acknowledged. 963

964 **References**

- Abotalib, A.Z., Heggy, E., 2019. A deep groundwater origin for recurring slope lineae on Mars. Nat.
 Geosci. 12, 235–241. https://doi.org/10.1038/s41561-019-0327-5
- Airo, A., 2015. Landslide (Mars), in: Gargaud, M., Irvine, W.M., Amils, R., Cleaves, H.J. (Jim), Pinti, D.L.,
 Quintanilla, J.C., Rouan, D., Spohn, T., Tirard, S., Viso, M. (Eds.), Encyclopedia of Astrobiology.
 Springer Berlin Heidelberg, Berlin, Heidelberg, pp. 1358–1359. https://doi.org/10.1007/978-3662-44185-5 2473
- Baker, V.R., Strom, R.G., Gulick, V.C., Kargel, J.S., Komatsu, G., Kale, V.S., 1991. Ancient oceans, ice
 sheets and the hydrological cycle on Mars. Nature 352, 589–594.
 https://doi.org/10.1038/352589a0
- Baum, R.L., Savage, W.Z., Wasowski, J., 2014. Mechanics of earth flows 9.
- Bibring, J.-P., Langevin, Y., Gendrin, A., Gondet, B., Poulet, F., Berthé, M., Soufflot, A., Arvidson, R.,
 Mangold, N., Mustard, J., Drossart, P., Team OMEGA, 2005. Mars Surface Diversity as Revealed
 by the OMEGA/Mars Express Observations. Science 307, 1576–1581.
 https://doi.org/10.1126/science.1108806
- Binet, S., Mudry, J., Scavia, C., Campus, S., Bertrand, C., Guglielmi, Y., 2007. In situ characterization of
 flows in a fractured unstable slope. Geomorphology 86, 193–203.
 https://doi.org/10.1016/j.geomorph.2006.08.013
- Blair, T.C., McPherson, J.G., 1999. Grain-size and textural classification of coarse sedimentary particles.
 J. Sediment. Res. 69, 6–19. https://doi.org/10.2110/jsr.69.6
- Bommer, J.J., Rodríguez, C.E., 2002. Earthquake-induced landslides in Central America. Eng. Geol. 63,
 189–220. https://doi.org/10.1016/S0013-7952(01)00081-3
- Bouchut, F., Mangeney-Castelnau, A., Perthame, B., Vilotte, J.-P., 2003. A new model of Saint Venant
 and Savage–Hutter type for gravity driven shallow water flows. Comptes Rendus Math. 336,
 531–536. https://doi.org/10.1016/S1631-073X(03)00117-1
- Bouchut, F., Westdickenberg, M., 2004. Gravity driven shallow water models for arbitrary topography.
 Commun. Math. Sci. 2, 359–389.
- Brown, A.J., Hook, S.J., Baldridge, A.M., Crowley, J.K., Bridges, N.T., Thomson, B.J., Marion, G.M., de
 Souza Filho, C.R., Bishop, J.L., 2010. Hydrothermal formation of Clay-Carbonate alteration
 assemblages in the Nili Fossae region of Mars. Earth Planet. Sci. Lett. 297, 174–182.
 https://doi.org/10.1016/j.epsl.2010.06.018
- Brunet, M., Moretti, L., Le Friant, A., Mangeney, A., Fernández Nieto, E.D., Bouchut, F., 2017. Numerical
 simulation of the 30–45 ka debris avalanche flow of Montagne Pelée volcano, Martinique:
 from volcano flank collapse to submarine emplacement. Nat. Hazards 87, 1189–1222.
 https://doi.org/10.1007/s11069-017-2815-5
- 999Brunetti, M.T., Guzzetti, F., Cardinali, M., Fiorucci, F., Santangelo, M., Mancinelli, P., Komatsu, G.,1000Borselli, L., 2014. Analysis of a new geomorphological inventory of landslides in Valles1001Marineris, Mars. Earth Planet. Sci. Lett. 405, 156–168.1002https://doi.org/10.1016/j.epsl.2014.08.025
- Butcher, F.E.G., Balme, M.R., Gallagher, C., Arnold, N.S., Conway, S.J., Hagermann, A., Lewis, S.R., 2017.
 Recent Basal Melting of a Mid-Latitude Glacier on Mars. J. Geophys. Res. Planets 122, 2445–
 2468. https://doi.org/10.1002/2017JE005434
- Carter, J., Poulet, F., Bibring, J.-P., Mangold, N., Murchie, S., 2013. Hydrous minerals on Mars as seen
 by the CRISM and OMEGA imaging spectrometers: Updated global view. J. Geophys. Res.
 Planets 118, 831–858. https://doi.org/10.1029/2012JE004145
- 1009 Chandler, R.J., 1972. Periglacial mudslides in Vestspitsbergen and their bearing on the origin of fossil
 1010 'solifluction' shears in low angled clay slopes. Q. J. Eng. Geol. Hydrogeol. 5, 223–241.
 1011 https://doi.org/10.1144/GSL.QJEG.1972.005.03.02
- Chen, R.-F., Chang, K.-J., Angelier, J., Chan, Y.-C., Deffontaines, B., Lee, C.-T., Lin, M.-L., 2006.
 Topographical changes revealed by high-resolution airborne LiDAR data: The 1999 Tsaoling

- 1014landslide induced by the Chi–Chi earthquake. Eng. Geol. 88, 160–172.1015https://doi.org/10.1016/j.enggeo.2006.09.008
- 1016Chigira, M., Wu, X., Inokuchi, T., Wang, G., 2010. Landslides induced by the 2008 Wenchuan1017earthquake, Sichuan, China. Geomorphology118, 225–238.1018https://doi.org/10.1016/j.geomorph.2010.01.003
- Coleman, N.M., Dinwiddie, C.L., Baker, V.R., 2007. Evidence that floodwaters filled and overflowed
 Capri Chasma, Mars. Geophys. Res. Lett. 34. https://doi.org/10.1029/2006GL028872
- 1021 Comegna, L., Picarelli, L., Urciuoli, G., 2007. The mechanics of mudslides as a cyclic undrained–drained 1022 process. Landslides 4, 217–232. https://doi.org/10.1007/s10346-007-0083-2
- Conway, S.J., Balme, M.R., 2014. Decameter thick remnant glacial ice deposits on Mars. Geophys. Res.
 Lett. 41, 5402–5409. https://doi.org/10.1002/2014GL060314
- Coquin, J., Mercier, D., Bourgeois, O., Decaulne, A., 2019. A paraglacial rock-slope failure origin for
 cirques: a case study from Northern Iceland. Géomorphologie Relief Process. Environ. 25, 117–
 136. https://doi.org/10.4000/geomorphologie.13057
- 1028Corominas, J., 1994. Evidence of basal erosion and shearing as mechanisms contributing the1029development of lateral ridges in mudslides, flow-slides, and other flow-like gravitational1030movements 26.
- 1031 Costard, F., Forget, F., Mangold, N., Peulvast, J.P., 2002. Formation of Recent Martian Debris Flows by
 1032 Melting of Near-Surface Ground Ice at High Obliquity. Science 295, 110–113.
 1033 https://doi.org/10.1126/science.1066698
- 1034Crosta, G.B., Frattini, P., Valbuzzi, E., De Blasio, F.V., 2018. Introducing a New Inventory of Large1035Martian Landslides. Earth Space Sci. 5, 89–119. https://doi.org/10.1002/2017EA000324
- 1036De Blasio, F.V., 2011. Landslides in Valles Marineris (Mars): A possible role of basal lubrication by sub-1037surface ice. Planet. Space Sci. 59, 1384–1392. https://doi.org/10.1016/j.pss.2011.04.015
- 1038de Haas, T., Conway, S.J., Krautblatter, M., 2015. Recent (Late Amazonian) enhanced backweathering1039rates on Mars: Paracratering evidence from gully alcoves. J. Geophys. Res. Planets 120, 2169–10402189. https://doi.org/10.1002/2015JE004915
- 1041de Haas, T., Hauber, E., Kleinhans, M.G., 2013. Local late Amazonian boulder breakdown and1042denudation rate on Mars. Geophys. Res. Lett. 40, 3527–3531.1043https://doi.org/10.1002/grl.50726
- Delannay, R., Valance, A., Mangeney, A., Roche, O., Richard, P., 2017. Granular and particle-laden
 flows: from laboratory experiments to field observations. J. Phys. Appl. Phys. 50, 053001.
 https://doi.org/10.1088/1361-6463/50/5/053001
- Ehlmann, B.L., Mustard, J.F., Murchie, S.L., Bibring, J.-P., Meunier, A., Fraeman, A.A., Langevin, Y., 2011.
 Subsurface water and clay mineral formation during the early history of Mars. Nature 479, 53–
 60. https://doi.org/10.1038/nature10582
- Ehlmann, B.L., Mustard, J.F., Swayze, G.A., Clark, R.N., Bishop, J.L., Poulet, F., Des Marais, D.J., Roach,
 L.H., Milliken, R.E., Wray, J.J., Barnouin-Jha, O., Murchie, S.L., 2009. Identification of hydrated
 silicate minerals on Mars using MRO-CRISM: Geologic context near Nili Fossae and implications
 for aqueous alteration. J. Geophys. Res. 114, E00D08. https://doi.org/10.1029/2009JE003339
- 1054Fassett, C.I., Head, J.W., 2005. Fluvial sedimentary deposits on Mars: Ancient deltas in a crater lake in1055theNiliFossaeregion.Geophys.Res.Lett.32,n/a-n/a.1056https://doi.org/10.1029/2005GL023456
- Favreau, P., Mangeney, A., Lucas, A., Crosta, G., Bouchut, F., 2010. Numerical modeling of landquakes:
 landslide and seismic waves. Geophys. Res. Lett. 37, n/a-n/a.
 https://doi.org/10.1029/2010GL043512
- Félix, G., Thomas, N., 2004. Relation between dry granular flow regimes and morphology of deposits:
 formation of levées in pyroclastic deposits. Earth Planet. Sci. Lett. 221, 197–213.
 https://doi.org/10.1016/S0012-821X(04)00111-6
- Fernandez-Nieto, E.D., Garres-Diaz, J., Mangeney, A., Narbona-Reina, G., 2016. A multilayer shallow
 model for dry granular flows with the mu(I) -rheology: application to granular collapse on
 erodible beds. J. Fluid Mech. 643–681.

- 1066Gallagher, C., Balme, M., 2015. Eskers in a complete, wet-based glacial system in the Phlegra Montes1067region, Mars. Earth Planet. Sci. Lett. 431, 96–109. https://doi.org/10.1016/j.epsl.2015.09.023
- 1068Gaudin, A., Dehouck, E., Mangold, N., 2011. Evidence for weathering on early Mars from a comparison1069withterrestrialweatheringprofiles.Icarus216,257–268.1070https://doi.org/10.1016/j.icarus.2011.09.004
- 1071 Giardini, D., Lognonné, P., Banerdt, W.B., Pike, W.T., Christensen, U., Ceylan, S., Clinton, J.F., van Driel, 1072 M., Stähler, S.C., Böse, M., Garcia, R.F., Khan, A., Panning, M., Perrin, C., Banfield, D., Beucler, 1073 E., Charalambous, C., Euchner, F., Horleston, A., Jacob, A., Kawamura, T., Kedar, S., Mainsant, 1074 G., Scholz, J.-R., Smrekar, S.E., Spiga, A., Agard, C., Antonangeli, D., Barkaoui, S., Barrett, E., 1075 Combes, P., Conejero, V., Daubar, I., Drilleau, M., Ferrier, C., Gabsi, T., Gudkova, T., Hurst, K., 1076 Karakostas, F., King, S., Knapmeyer, M., Knapmeyer-Endrun, B., Llorca-Cejudo, R., Lucas, A., 1077 Luno, L., Margerin, L., McClean, J.B., Mimoun, D., Murdoch, N., Nimmo, F., Nonon, M., Pardo, 1078 C., Rivoldini, A., Manfredi, J.A.R., Samuel, H., Schimmel, M., Stott, A.E., Stutzmann, E., Teanby, N., Warren, T., Weber, R.C., Wieczorek, M., Yana, C., 2020. The seismicity of Mars. Nat. Geosci. 1079 1080 13, 205–212. https://doi.org/10.1038/s41561-020-0539-8
- 1081 Gray, N., 2014. Erosion-deposition waves in shallow granular free-surface flows.
- 1082Grindrod, P.M., Warner, N.H., 2014. Erosion rate and previous extent of interior layered deposits on1083Marsrevealedbyobstructedlandslides.Geology42,795–798.1084https://doi.org/10.1130/G35790.1
- Hartmann, W.K., 2005. Martian cratering 8: Isochron refinement and the chronology of Mars. Icarus
 174, 294–320. https://doi.org/10.1016/j.icarus.2004.11.023
- Head, J.W., Mustard, J.F., Kreslavsky, M.A., Milliken, R.E., Marchant, D.R., 2003. Recent ice ages on
 Mars. Nature 426, 797–802. https://doi.org/10.1038/nature02114
- Hildenbrand, A., Gillot, P.-Y., Bonneville, A., 2006. Offshore evidence for a huge landslide of the
 northern flank of Tahiti-Nui (French Polynesia): Tahiti-Nui landslide. Geochem. Geophys.
 Geosystems 7, n/a-n/a. https://doi.org/10.1029/2005GC001003
- Jessop, D.E., Kelfoun, K., Labazuy, P., Mangeney, A., Roche, O., Tillier, J.-L., Trouillet, M., Thibault, G.,
 2012. LiDAR derived morphology of the 1993 Lascar pyroclastic flow deposits, and implication
 for flow dynamics and rheology ScienceDirect.
- 1095Johnson, B.C., Campbell, C.S., 2017. Drop Height and Volume Control the Mobility of Long-Runout1096Landslides on the Earth and Mars. Geophys. Res. Lett. 44, 12,091-12,097.1097https://doi.org/10.1002/2017GL076113
- 1098Kneissl, T., van Gasselt, S., Neukum, G., 2011. Map-projection-independent crater size-frequency1099determination in GIS environments—New software tool for ArcGIS. Planet. Space Sci. 59,11001243–1254. https://doi.org/10.1016/j.pss.2010.03.015
- Kraal, E.R., Wong, M.P., Grosfils, E.B., Gilmore, M.S., Kozak, S.J., Reinen, L.A., 1998. The Origin and
 Modification of a Trough in the Nili Fossae, Mars 3.
- Kreslavsky, M.A., Head, J.W., Marchant, D.R., 2008. Periods of active permafrost layer formation during
 the geological history of Mars: Implications for circum-polar and mid-latitude surface
 processes. Planet. Space Sci., Mars Polar Processes: Atmosphere-Surface Interactions 56, 289–
 302. https://doi.org/10.1016/j.pss.2006.02.010
- Laskar, J., Correia, A.C.M., Gastineau, M., Joutel, F., Levrard, B., Robutel, P., 2004. Long term evolution
 and chaotic diffusion of the insolation quantities of Mars. Icarus 170, 343–364.
 https://doi.org/10.1016/j.icarus.2004.04.005
- 1110
 Legros, F., 2002.
 The mobility of long-runout landslides.
 Eng.
 Geol.
 63, 301–331.

 1111
 https://doi.org/10.1016/S0013-7952(01)00090-4
- Levy, C., Mangeney, A., Bonilla, F., Hibert, C., Calder, E.S., Smith, P.J., 2015. Friction weakening in granular flows deduced from seismic records at the Soufrière Hills Volcano, Montserrat. J.
 Geophys. Res. Solid Earth 120, 7536–7557. https://doi.org/10.1002/2015JB012151
- Lognonné, P., Banerdt, W.B., Giardini, D., Pike, W.T., Christensen, U., Laudet, P., de Raucourt, S.,
 Zweifel, P., Calcutt, S., Bierwirth, M., Hurst, K.J., Ijpelaan, F., Umland, J.W., Llorca-Cejudo, R.,
 Larson, S.A., Garcia, R.F., Kedar, S., Knapmeyer-Endrun, B., Mimoun, D., Mocquet, A., Panning,

M.P., Weber, R.C., Sylvestre-Baron, A., Pont, G., Verdier, N., Kerjean, L., Facto, L.J., 1118 1119 Gharakanian, V., Feldman, J.E., Hoffman, T.L., Klein, D.B., Klein, K., Onufer, N.P., Paredes-Garcia, J., Petkov, M.P., Willis, J.R., Smrekar, S.E., Drilleau, M., Gabsi, T., Nebut, T., Robert, O., 1120 1121 Tillier, S., Moreau, C., Parise, M., Aveni, G., Ben Charef, S., Bennour, Y., Camus, T., Dandonneau, P.A., Desfoux, C., Lecomte, B., Pot, O., Revuz, P., Mance, D., tenPierick, J., Bowles, N.E., 1122 1123 Charalambous, C., Delahunty, A.K., Hurley, J., Irshad, R., Liu, H., Mukherjee, A.G., Standley, I.M., Stott, A.E., Temple, J., Warren, T., Eberhardt, M., Kramer, A., Kühne, W., Miettinen, E.-P., 1124 1125 Monecke, M., Aicardi, C., André, M., Baroukh, J., Borrien, A., Bouisset, A., Boutte, P., Brethomé, 1126 K., Brysbaert, C., Carlier, T., Deleuze, M., Desmarres, J.M., Dilhan, D., Doucet, C., Faye, D., Faye-Refalo, N., Gonzalez, R., Imbert, C., Larigauderie, C., Locatelli, E., Luno, L., Meyer, J.-R., Mialhe, 1127 1128 F., Mouret, J.M., Nonon, M., Pahn, Y., Paillet, A., Pasquier, P., Perez, G., Perez, R., Perrin, L., 1129 Pouilloux, B., Rosak, A., Savin de Larclause, I., Sicre, J., Sodki, M., Toulemont, N., Vella, B., Yana, 1130 C., Alibay, F., Avalos, O.M., Balzer, M.A., Bhandari, P., Blanco, E., Bone, B.D., Bousman, J.C., 1131 Bruneau, P., Calef, F.J., Calvet, R.J., D'Agostino, S.A., de los Santos, G., Deen, R.G., Denise, R.W., 1132 Ervin, J., Ferraro, N.W., Gengl, H.E., Grinblat, F., Hernandez, D., Hetzel, M., Johnson, M.E., 1133 Khachikyan, L., Lin, J.Y., Madzunkov, S.M., Marshall, S.L., Mikellides, I.G., Miller, E.A., Raff, W., 1134 Singer, J.E., Sunday, C.M., Villalvazo, J.F., Wallace, M.C., Banfield, D., Rodriguez-Manfredi, J.A., 1135 Russell, C.T., Trebi-Ollennu, A., Maki, J.N., Beucler, E., Böse, M., Bonjour, C., Berenguer, J.L., Ceylan, S., Clinton, J., Conejero, V., Daubar, I., Dehant, V., Delage, P., Euchner, F., Estève, I., 1136 Fayon, L., Ferraioli, L., Johnson, C.L., Gagnepain-Beyneix, J., Golombek, M., Khan, A., 1137 1138 Kawamura, T., Kenda, B., Labrot, P., Murdoch, N., Pardo, C., Perrin, C., Pou, L., Sauron, A., 1139 Savoie, D., Stähler, S., Stutzmann, E., Teanby, N.A., Tromp, J., van Driel, M., Wieczorek, M., 1140 Widmer-Schnidrig, R., Wookey, J., 2019. SEIS: Insight's Seismic Experiment for Internal 1141 Structure of Mars. Space Sci. Rev. 215, 12. https://doi.org/10.1007/s11214-018-0574-6 Lucas, A., Mangeney, A., 2007. Mobility and topographic effects for large Valles Marineris landslides 1142 1143 on Mars. Geophys. Res. Lett. 34. https://doi.org/10.1029/2007GL029835 1144 Lucas, A., Mangeney, A., Ampuero, J.P., 2014. Frictional velocity-weakening in landslides on Earth and on other planetary bodies. Nat. Commun. 5. https://doi.org/10.1038/ncomms4417 1145 1146 Lucas, A., Mangeney, A., Bouchut, F., Bristeau, M.O., Mège, D., 2007. Benchmarking exercises for 1147 granular flows, in: Unpublished. https://doi.org/10.13140/rg.2.1.3945.0088 1148 Lucas, A., Mangeney, A., Mège, D., Bouchut, F., 2011. Influence of the scar geometry on landslide 1149 dynamics and deposits: Application to Martian landslides. J. Geophys. Res. 116. https://doi.org/10.1029/2011JE003803 1150 1151 Lucchitta, B.K., 1979. Landslides in Valles Marineris, Mars. J. Geophys. Res. 84, 8097. 1152 https://doi.org/10.1029/JB084iB14p08097 Magnarini, G., Mitchell, T.M., Grindrod, P.M., Goren, L., Schmitt, H.H., 2019. Longitudinal ridges 1153 1154 imparted by high-speed granular flow mechanisms in martian landslides. Nat. Commun. 10, 1– 1155 7. https://doi.org/10.1038/s41467-019-12734-0 1156 Malin, M.C., Bell, J.F., Cantor, B.A., Caplinger, M.A., Calvin, W.M., Clancy, R.T., Edgett, K.S., Edwards, 1157 L., Haberle, R.M., James, P.B., Lee, S.W., Ravine, M.A., Thomas, P.C., Wolff, M.J., 2007. Context 1158 Camera Investigation on board the Mars Reconnaissance Orbiter. J. Geophys. Res. 112, 1159 E05S04. https://doi.org/10.1029/2006JE002808 1160 Mangeney, A., Bouchut, F., Thomas, N., Vilotte, J.P., Bristeau, M.O., 2007. Numerical modeling of selfchanneling granular flows and of their levee-channel deposits. J. Geophys. Res. Earth Surf. 112. 1161 1162 https://doi.org/10.1029/2006JF000469 1163 Mangeney, A., Roche, O., Hungr, O., Mangold, N., Faccanoni, G., Lucas, A., 2010. Erosion and mobility in granular collapse over sloping beds. J. Geophys. Res. Earth Surf. 115. 1164 1165 https://doi.org/10.1029/2009JF001462 1166 Mangeney-Castelnau, A., Bouchut, F., Vilotte, J.P., Lajeunesse, E., Aubertin, A., Pirulli, M., 2005. On the 1167 use of Saint Venant equations to simulate the spreading of a granular mass. J. Geophys. Res. 1168 Solid Earth 110. https://doi.org/10.1029/2004JB003161

40

- Mangold, N., Mangeney, A., Migeon, V., Ansan, V., Lucas, A., Baratoux, D., Bouchut, F., 2010. Sinuous
 gullies on Mars: Frequency, distribution, and implications for flow properties. J. Geophys. Res.
 1171 115. https://doi.org/10.1029/2009JE003540
- Mangold, N., Poulet, F., Mustard, J.F., Bibring, J.-P., Gondet, B., Langevin, Y., Ansan, V., Masson, Ph.,
 Fassett, C., Head, J.W., Hoffmann, H., Neukum, G., 2007. Mineralogy of the Nili Fossae region
 with OMEGA/Mars Express data: 2. Aqueous alteration of the crust: MINERALOGY OF THE NILI
 FOSSAE REGION, 2. J. Geophys. Res. Planets 112. https://doi.org/10.1029/2006JE002835
- 1176 Marc, O., Hovius, N., Meunier, P., Uchida, T., Hayashi, S., 2015. Transient changes of landslide rates 1177 after earthquakes. Geology 43, 883–886. https://doi.org/10.1130/G36961.1
- Masson, D.G., Watts, A.B., Gee, M.J.R., Urgeles, R., Mitchell, N.C., Le Bas, T.P., Canals, M., 2002. Slope
 failures on the flanks of the western Canary Islands. Earth-Sci. Rev. 57, 1–35.
 https://doi.org/10.1016/S0012-8252(01)00069-1
- 1181 McEwen, A.S., 1989. Mobility of large rock avalanches: Evidence from Valles Marineris, Mars. Geology 1182 17, 1111–1114. https://doi.org/10.1130/0091-7613(1989)017<1111:MOLRAE>2.3.CO;2
- McEwen, A.S., Eliason, E.M., Bergstrom, J.W., Bridges, N.T., Hansen, C.J., Delamere, W.A., Grant, J.A.,
 Gulick, V.C., Herkenhoff, K.E., Keszthelyi, L., Kirk, R.L., Mellon, M.T., Squyres, S.W., Thomas, N.,
 Weitz, C.M., 2007. Mars Reconnaissance Orbiter's High Resolution Imaging Science
 Experiment (HiRISE). J. Geophys. Res. Planets 112. https://doi.org/10.1029/2005JE002605
- McSaveney, M., Davies, T., 2007. Rockslides and Their Motion, in: Sassa, K., Fukuoka, H., Wang, F.,
 Wang, G. (Eds.), Progress in Landslide Science. Springer Berlin Heidelberg, Berlin, Heidelberg,
 pp. 113–133. https://doi.org/10.1007/978-3-540-70965-7_8
- Mellon, M.T., Phillips, R.J., 2001. Recent gullies on Mars and the source of liquid water. J. Geophys.
 Res. Planets 106, 23165–23179. https://doi.org/10.1029/2000JE001424
- 1192Meunier, P., Hovius, N., Haines, A.J., 2007. Regional patterns of earthquake-triggered landslides and1193their relation to ground motion. Geophys. Res. Lett. 34.1194https://doi.org/10.1029/2007GL031337
- 1195Michael, G.G., 2013. Planetary surface dating from crater size-frequency distribution measurements:1196Multiple resurfacing episodes and differential isochron fitting. Icarus 226, 885–890.1197https://doi.org/10.1016/j.icarus.2013.07.004
- 1198Michael, G.G., Kneissl, T., Neesemann, A., 2016. Planetary surface dating from crater size-frequency1199distribution measurements: Poisson timing analysis. Icarus 277, 279–285.1200https://doi.org/10.1016/j.icarus.2016.05.019
- Michael, G.G., Neukum, G., 2010. Planetary surface dating from crater size–frequency distribution
 measurements: Partial resurfacing events and statistical age uncertainty. Earth Planet. Sci.
 Lett. 294, 223–229. https://doi.org/10.1016/j.epsl.2009.12.041
- Michael, G.G., Platz, T., Kneissl, T., Schmedemann, N., 2012. Planetary surface dating from crater size–
 frequency distribution measurements: Spatial randomness and clustering. Icarus 218, 169–
 177. https://doi.org/10.1016/j.icarus.2011.11.033
- Michalski, J., Poulet, F., Bibring, J.-P., Mangold, N., 2010. Analysis of phyllosilicate deposits in the Nili
 Fossae region of Mars: Comparison of TES and OMEGA data. Icarus 206, 269–289.
 https://doi.org/10.1016/j.icarus.2009.09.006
- Moratto, Z.M., Michael J. Broxton, Ross A. Beyer, Mike Lundy, Kyle Husmann, 2010. Ames Stereo
 Pipeline, NASA's Open Source Automated Stereogrammetry Software. 41st Lunar and
 Planetary Science Conference.
- Moretti, L., Allstadt, K., Mangeney, A., Capdeville, Y., Stutzmann, E., Bouchut, F., 2015. Numerical
 modeling of the Mount Meager landslide constrained by its force history derived from seismic
 data. J. Geophys. Res. Solid Earth 120, 2579–2599. https://doi.org/10.1002/2014JB011426
- Moretti, L., Mangeney, A., Capdeville, Y., Stutzmann, E., Huggel, C., Schneider, D., Bouchut, F., 2012.
 Numerical modeling of the Mount Steller landslide flow history and of the generated long
 period seismic waves: landquake modeling. Geophys. Res. Lett. 39, n/a-n/a.
 https://doi.org/10.1029/2012GL052511

- Moretti, L., Mangeney, A., Walter, F., Capdeville, Y., Bodin, T., Stutzmann, E., Le Friant, A., 2020.
 Constraining landslide characteristics with Bayesian inversion of field and seismic data.
 Geophys. J. Int. 221, 1341–1348. https://doi.org/10.1093/gji/ggaa056
- 1223 Murchie, S., Arvidson, R., Bedini, P., Beisser, K., Bibring, J.-P., Bishop, J., Boldt, J., Cavender, P., Choo, T., Clancy, R.T., Darlington, E.H., Des Marais, D., Espiritu, R., Fort, D., Green, R., Guinness, E., 1224 1225 Hayes, J., Hash, C., Heffernan, K., Hemmler, J., Heyler, G., Humm, D., Hutcheson, J., Izenberg, 1226 N., Lee, R., Lees, J., Lohr, D., Malaret, E., Martin, T., McGovern, J.A., McGuire, P., Morris, R., Mustard, J., Pelkey, S., Rhodes, E., Robinson, M., Roush, T., Schaefer, E., Seagrave, G., Seelos, 1227 1228 F., Silverglate, P., Slavney, S., Smith, M., Shyong, W.-J., Strohbehn, K., Taylor, H., Thompson, P., 1229 Tossman, B., Wirzburger, M., Wolff, M., 2007. Compact Reconnaissance Imaging Spectrometer 1230 for Mars (CRISM) on Mars Reconnaissance Orbiter (MRO). J. Geophys. Res. 112, E05S03. 1231 https://doi.org/10.1029/2006JE002682
- Mustard, J.F., Cooper, C.D., Rifkin, M.K., 2001. Evidence for recent climate change on Mars from the
 identification of youthful near-surface ground ice. Nature 412, 411–414.
 https://doi.org/10.1038/35086515
- Mustard, J.F., Poulet, F., Head, J.W., Mangold, N., Bibring, J.-P., Pelkey, S.M., Fassett, C.I., Langevin, Y.,
 Neukum, G., 2007. Mineralogy of the Nili Fossae region with OMEGA/Mars Express data: 1.
 Ancient impact melt in the Isidis Basin and implications for the transition from the Noachian
 to Hesperian. J. Geophys. Res. Planets 112. https://doi.org/10.1029/2006JE002834
- Nereson, A.L., Finnegan, N.J., 2015. Earthflow yield strength constrained by lateral levee morphology.
 AGU Fall Meet. Abstr. 2015, EP41C-0943.
- Neukum, G., Jaumann, R., 2004. HRSC: the High Resolution Stereo Camera of Mars Express 1240, 17–
 35.
- 1243Niu, F., Luo, J., Lin, Z., Fang, J., Liu, M., 2015. Thaw-induced slope failures and stability analyses in1244permafrost regions of the Qinghai-Tibet Plateau, China. Landslides 13.1245https://doi.org/10.1007/s10346-014-0545-2
- 1246Pajola, M., Rossato, S., Baratti, E., Mangili, C., Mancarella, F., McBride, K., Coradini, M., 2016. The1247Simud–Tiu Valles hydrologic system: A multidisciplinary study of a possible site for future Mars1248on-site exploration. Icarus 268, 355–381. https://doi.org/10.1016/j.icarus.2015.12.049
- 1249Parise, M., 2003. Observation of surface features on an active landslide, and implications for1250understanding its history of movement. Nat. Hazards Earth Syst. Sci. 3, 569–580.1251https://doi.org/10.5194/nhess-3-569-2003
- Peruzzetto, M., Komorowski, J.-C., Le Friant, A., Rosas-Carbajal, M., Mangeney, A., Legendre, Y., 2019.
 Modeling of partial dome collapse of La Soufrière of Guadeloupe volcano: implications for
 hazard assessment and monitoring. Sci. Rep. 9, 1–15. https://doi.org/10.1038/s41598-019 49507-0
- Porter, C., Morin, P., Howat, I., Noh, M.-J., Bates, B., Peterman, K., Keesey, S., Schlenk, M., Gardiner, J.,
 Tomko, K., Willis, M., Kelleher, C., Cloutier, M., Husby, E., Foga, S., Nakamura, H., Platson, M.,
 Wethington, M., Williamson, C., Bauer, G., Enos, J., Arnold, G., Kramer, W., Becker, P., Doshi,
 A., D'Souza, C., Cummens, P., Laurier, F., Bojesen, M., 2018. ArcticDEM.
 https://doi.org/10.7910/DVN/OHHUKH
- Poulet, F., Bibring, J.-P., Mustard, J.F., Gendrin, A., Mangold, N., Langevin, Y., Arvidson, R.E., Gondet,
 B., Gomez, C., 2005. Phyllosilicates on Mars and implications for early martian climate. Nature
 438, 623. https://doi.org/10.1038/nature04274
- 1264 Pouliquen, O., Forterre, Y., 2002. Friction law for dense granular flows: application to the motion of a 1265 mass down а rough inclined plane. J. Fluid Mech. 453, 133–151. https://doi.org/10.1017/S0022112001006796 1266
- 1267Quantin, C., Allemand, P., Delacourt, C., 2004a. Morphology and geometry of Valles Marineris1268landslides. Planet. Space Sci. 52, 1011–1022. https://doi.org/10.1016/j.pss.2004.07.016
- 1269Quantin, C., Allemand, P., Mangold, N., Delacourt, C., 2004b. Ages of Valles Marineris (Mars) landslides1270and implications for canyon history.Icarus 172, 555–572.1271https://doi.org/10.1016/j.icarus.2004.06.013

- 1272Rapp, A., 1960. Recent Development of Mountain Slopes in Kärkevagge and Surroundings, Northern1273Scandinavia. Geogr. Ann. 42, 65–200. https://doi.org/10.1080/20014422.1960.11880942
- 1274Rocha, F.M., Johnson, C.G., Gray, J.M.N.T., 2019. Self-channelisation and levee formation in1275monodisperse granular flows. J. Fluid Mech. 876, 591–641.1276https://doi.org/10.1017/jfm.2019.518
- Sæmundsson, Þ., Morino, C., Helgason, J.K., Conway, S.J., Pétursson, H.G., 2018. The triggering factors
 of the Móafellshyrna debris slide in northern Iceland: Intense precipitation, earthquake
 activity and thawing of mountain permafrost. Sci. Total Environ. 621, 1163–1175.
 https://doi.org/10.1016/j.scitotenv.2017.10.111
- Sassa, K., Fukuoka, H., Wang, F., Wang, G., 2007. Landslides Induced by a Combined Effect of
 Earthquake and Rainfall, in: Sassa, K., Fukuoka, H., Wang, F., Wang, G. (Eds.), Progress in
 Landslide Science. Springer, Berlin, Heidelberg, pp. 193–207. https://doi.org/10.1007/978-3540-70965-7_14
- Schultz, R.A., 2002. Stability of rock slopes in Valles Marineris, Mars. Geophys. Res. Lett. 29, 38-1-38–
 4. https://doi.org/10.1029/2002GL015728
- Schultz, R.A., 2000. Localization of bedding plane slip and backthrust faults above blind thrust faults:
 Keys to wrinkle ridge structure. J. Geophys. Res. Planets 105, 12035–12052.
 https://doi.org/10.1029/1999JE001212
- Sepúlveda, S.A., Serey, A., Lara, M., Pavez, A., Rebolledo, S., 2010. Landslides induced by the April 2007
 Aysén Fjord earthquake, Chilean Patagonia. Landslides 7, 483–492.
 https://doi.org/10.1007/s10346-010-0203-2
- Shaller, P.J., 1991. Analysis and implications of large Martian and terrestrial landslides. CaliforniaInstitute of Technology.
- 1295
 Sharp,
 R.P.,
 1973.
 Mars:
 Troughed
 terrain.
 J.
 Geophys.
 Res.
 1896-1977
 78,
 4063–4072.

 1296
 https://doi.org/10.1029/JB078i020p04063
 https://doi.org/10.1029/JB078i020p04063
 https://doi.org/10.1029/JB078i020p04063
- Shea, T., van Wyk de Vries, B., 2008. Structural analysis and analogue modeling of the kinematics and
 dynamics of rockslide avalanches. Geosphere 4, 657. https://doi.org/10.1130/GES00131.1
- Smith, D.E., Zuber, M.T., Frey, H.V., Garvin, J.B., Head, J.W., Muhleman, D.O., Pettengill, G.H., Phillips,
 R.J., Solomon, S.C., Zwally, H.J., Banerdt, W.B., Duxbury, T.C., Golombek, M.P., Lemoine, F.G.,
 Neumann, G.A., Rowlands, D.D., Aharonson, O., Ford, P.G., Ivanov, A.B., Johnson, C.L.,
 McGovern, P.J., Abshire, J.B., Afzal, R.S., Sun, X., 2001. Mars Orbiter Laser Altimeter:
 Experiment summary after the first year of global mapping of Mars. J. Geophys. Res. Planets
 106, 23689–23722. https://doi.org/10.1029/2000JE001364
- Soukhovitskaya, V., Manga, M., 2006. Martian landslides in Valles Marineris: Wet or dry? Icarus 180,
 348–352. https://doi.org/10.1016/j.icarus.2005.09.008
- 1307Strecker, M.R., Fauque, L., 1988. Large rock avalanche deposits (Sturzströme, sturzstroms) at Sierra1308Aconquija, northern Sierras Pampeanas, Argentina. https://doi.org/10.5169/seals-166195
- Tanaka, K.L., Skinner, J.A., Dohm, J.M., Irwin, R.P., Kolb, E.J., Fortezzo, C.M., Platz, T., Michael, G.G.,
 Hare, T.M., 2014. Scientific Investigations Map (Scientific Investigations Map).
- Tanaka, K.L., Skinner, J.A., Hare, T.M., Joyal, T., Wenker, A., 2003. Resurfacing history of the northern
 plains of Mars based on geologic mapping of Mars Global Surveyor data. J. Geophys. Res.
 Planets 108. https://doi.org/10.1029/2002JE001908
- Teanby, N.A., Wookey, J., 2011. Seismic detection of meteorite impacts on Mars. Phys. Earth Planet.
 Inter. 186, 70–80. https://doi.org/10.1016/j.pepi.2011.03.004
- Tesson, P.-A., Conway, S.J., Mangold, N., Ciazela, J., Lewis, S.R., Mège, D., 2019. Evidence for thermal
 fatigue on Mars from rockfall patterns on impact crater slopes. Presented at the 50th Lunar
 and Planetary Science Conference, The Woodlands, Houston, Texas, USA.
- Thomas, N., Cremonese, G., Ziethe, R., Gerber, M., Brändli, M., Bruno, G., Erismann, M., Gambicorti,
 L., Gerber, T., Ghose, K., Gruber, M., Gubler, P., Mischler, H., Jost, J., Piazza, D., Pommerol, A.,
 Rieder, M., Roloff, V., Servonet, A., Trottmann, W., Uthaicharoenpong, T., Zimmermann, C.,
 Vernani, D., Johnson, M., Pelò, E., Weigel, T., Viertl, J., De Roux, N., Lochmatter, P., Sutter, G.,
 Casciello, A., Hausner, T., Ficai Veltroni, I., Da Deppo, V., Orleanski, P., Nowosielski, W.,

- Zawistowski, T., Szalai, S., Sodor, B., Tulyakov, S., Troznai, G., Banaskiewicz, M., Bridges, J.C.,
 Byrne, S., Debei, S., El-Maarry, M.R., Hauber, E., Hansen, C.J., Ivanov, A., Keszthelyi, L., Kirk, R.,
 Kuzmin, R., Mangold, N., Marinangeli, L., Markiewicz, W.J., Massironi, M., McEwen, A.S.,
 Okubo, C., Tornabene, L.L., Wajer, P., Wray, J.J., 2017. The Colour and Stereo Surface Imaging
 System (CaSSIS) for the ExoMars Trace Gas Orbiter. Space Sci. Rev. 212, 1897–1944.
 https://doi.org/10.1007/s11214-017-0421-1
- 1330 Vallf-Jo, L.E., 1979. Mechanics of mudflow mobilization in low-angled clay slopes 8.
- 1331 Viviano, C.E., Moersch, J.E., McSween, H.Y., 2013. Implications for early hydrothermal environments
 1332 on Mars through the spectral evidence for carbonation and chloritization reactions in the Nili
 1333 Fossae region. J. Geophys. Res. Planets 118, 1858–1872. https://doi.org/10.1002/jgre.20141
- Walter, M., Joswig, M., 2008. Seismic monitoring of fracture processes from a creeping landslide in the
 Vorarlberg Alps. Geophys Res Abstr 10, 09212.
- Wang, F.W., Sassa, K., Wang, G., 2002. Mechanism of a long-runout landslide triggered by the August
 1998 heavy rainfall in Fukushima Prefecture, Japan. Eng. Geol. 63, 169–185.
 https://doi.org/10.1016/S0013-7952(01)00080-1
- Warner, N.H., Gupta, S., Calef, F., Grindrod, P., Boll, N., Goddard, K., 2015. Minimum effective area for
 high resolution crater counting of martian terrains. Icarus 245, 198–240.
 https://doi.org/10.1016/j.icarus.2014.09.024
- Watkins, J.A., Ehlmann, B.L., Yin, A., 2015. Long-runout landslides and the long-lasting effects of early
 water activity on Mars. Geology 43, 107–110. https://doi.org/10.1130/G36215.1
- Wichman, R.W., Schultz, P.H., 1989. Sequence and mechanisms of deformation around the Hellas and
 Isidis Impact Basins on Mars. J. Geophys. Res. Solid Earth 94, 17333–17357.
 https://doi.org/10.1029/JB094iB12p17333
- 1347 Wójcicka, N., Collins, G.S., Bastow, I.D., Teanby, N.A., Miljković, K., Rajšić, A., Daubar, I., Lognonné, P.,
 1348 2020. The Seismic Moment and Seismic Efficiency of Small Impacts on Mars. J. Geophys. Res.
 1349 Planets 125, e2020JE006540. https://doi.org/10.1029/2020JE006540
- Yamada, M., Mangeney, A., Matsushi, Y., Matsuzawa, T., 2018. Estimation of dynamic friction and
 movement history of large landslides. Landslides 15, 1963–1974.
 https://doi.org/10.1007/s10346-018-1002-4
- Yamada, M., Mangeney, A., Matsushi, Y., Moretti, L., 2016. Estimation of dynamic friction of the
 Akatani landslide from seismic waveform inversion and numerical simulation. Geophys. J. Int.
 206, 1479–1486. https://doi.org/10.1093/gji/ggw216
- 1356

1357

1358 Supplementary material

1359 S1-Spectral analysis



Figure S1 - Fe/Mg clays exposed along the impact crater rim in Nili Fossae region. (a) CRISM enhanced IR colour, R= 2.5 μm;
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);
B= 1.9 μm band depth (BD1900R), pink colour: Fe/Mg phyllosilicates, blue colour: hydrated phase. (c) CRISM spectrum of
Fe/Mg-phyllosilicate 'average' of several pixels divided by pixels of a neutral region compared to laboratory spectra (RELAB
library spectra). (CRISM: HRS00024571_07_IF175S_TRR3)

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

- Fe-Mg-phyllosilicates (pink tones in Fig.S1b) such as nontronite, saponite, and vermiculite are observed on the bedrock along the impact crater rim and on its floor in the region where the Nilosyrtis Mensae landslide occurs. Other hydrated phase(s) are also associated with the bedrock. Either this or these 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) Hólmavík 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.

1383 S2-Topographic profiles

1384 On figure S2, for the three profiles placed for each levee, each cross-section line was placed by 1385 connecting with a straight line the points where a given contour line crosses the levee boundary on 1386 each side. We isolated each levee by identifying the point at which it rose from the surrounding terrain 1387 aided by the image data and extracted the elevation values at these points. We identified and 1388 extracted the elevation of the crest of the levee. For each side we calculated the slopes by taking the 1389 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 1390 1391 between the base and crest elevations.

1392 S3-Grain size sensitivity tests

1393 In order to investigate the influence of grain size in the model using the Pouliquen and Forterre's law, 1394 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 1395 1396 fit the runout distance. The smaller the grain size, the smaller the friction angles must be. In addition, 1397 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 1398 1399 deposition surface. On the CCh and ChrC slides, the entire deposition surface is covered with blocks 1400 larger than 5 m. This size has therefore been taken as a reference for modelling. For NM landslide, 1401 there are only a few scattered blocks of maximum 10 m in size near the toe of the deposit. Also, we 1402 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. 1403

1404



1405Figure S3 – Final deposit thickness maps resulting from the grain size sensitivity tests applied to Capri Chasma landslide1406simulation 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.

S4-Topographic reconstructions



Figure S4 - 3D view of two alternate topographic reconstructions for Capri Chasma landslide, where the colourised DEM is
overlain by semi-transparent hillshaded relief and 25 m interval contour lines in black. (a) Reconstruction with overdeepened
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
- 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.



1426Figure 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) at1428T=80s, (d) at T=118s and (f) at T=1200s when the flow is stabilised compared to (g), the actual deposits. Pouliquen and1429Forterre's law used with angles of $\delta_1 = 8$, $\delta_2 = 18$ and $\delta_3 = 11^\circ$.

1430 **References**

- Morgan, F., Seelos, F.P., Murchie, S.L., CRISM Team, 2014. CRISM Analysis Toolkit (CAT). The Johns
 Hopkins University.
- 1433 Viviano-Beck, C.E., Seelos, F.P., Murchie, S.L., Kahn, E.G., Seelos, K.D., Taylor, H.W., Taylor, K.,
- 1434 Ehlmann, B.L., Wiseman, S.M., Mustard, J.F., Morgan, M.F., 2014. Revised CRISM spectral
- 1435 parameters and summary products based on the currently detected mineral diversity on
- 1436Mars. Journal of Geophysical Research: Planets 119, 1403–1431.
- 1437 https://doi.org/10.1002/2014JE004627