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The influence of slope gradient and gully channel on the run-out behavior of rockslide-debris flow: an analysis on the Verghereto landslide in Italy

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4 **The influence of slope gradient and gully channel on the run-out behavior of**

5 **rockslide-debris flow: an analysis on the Verghereto landslide in Italy**

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9 Wei Shen¹, Matteo Berti¹, Tonglu Li^{2,3}, Andrea Benini⁴, Zhitian Qiao^{1*}

10 *Correspondence author

11

12

13

14 1 Department of Biological, Geological and Environmental Sciences, University of Bologna, Italy, 40126

15 2 Department of Geological Engineering, Chang'an University, Xi'an, China, 710064

16 3 Water Cycle and Geological Environment Observation and Research Station for the Chinese Loess Plateau, Ministry
17 of Education, Gansu China, 745399

18 4 Servizio Tecnico Bacino Romagna Regione Emilia-Romagna

19

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21 Paper submitted to "Landslides" for review and possible publication

22 **Abstract:**

23

24 Rockslide-debris flow is a hybrid type of mass movement occurring when a rockslide transforms
25 into a debris flow. This type of mass movement may cause catastrophic damages because of its high
26 speed and long run-out distance. To achieve a better understanding toward the run-out behavior of
27 this type of landslide, a recent rockslide-debris flow occurred in Verghereto (Northern Apennines of
28 Italy) is studied through field investigation and numerical simulation. The run-out process of this
29 landslide is simulated by an improved depth-averaged model, paying special attention to analyzing
30 the influence of slope gradient and gully channel. The results show that the depth-averaged model
31 can correctly simulate the entrainment and deposition characteristic of this landslide by adopting
32 different basal friction strengths for rockslide region and debris flow region. Entrainment occurs in
33 both high and low slope gradient zones. However, entrainment can only be observed in the high
34 slope gradient zones, while in the low gradient zones the post-failure topography shows
35 accumulation and deposition. The simulation results also demonstrate that the presence of a gully
36 channel is a key factor in determining landslide mobility and run-out distance. In comparison to a
37 landslide with similar size and geological settings but without a gully channel, the run-out distance
38 is much less and the landslide does not develop into a flow.

39

40 **Keywords:** Rockslide-debris flow, Numerical simulation, Solid-fluid transformation, Run-out
41 analysis, Bed entrainment

1. Introduction

A rockslide may transform into a debris flow when it disintegrates and propagates along a confined channel, and this hybrid mass movement is named as rockslide-debris flow. The term “debris flow” indicates partially or fully saturated flow-like movement propagating in gully channel (Hungr et al. 2014) and is distinguished from “rock avalanche” which describes the flow-like movement of essentially dry debris on unconfined slope. A rockslide-debris flow is typically characterized by the presence of a gully channel on the run-out path and it is renowned for the solid-fluid transformation (SFT) occurring during the run-out process. The SFT contributes to the high mobility of these types of landslides. Several factors, such as the disintegration of rock mass (Bowman et al. 2012, Crosta et al. 2007, Davies and McSaveney 2009), entrainment (Aaron and McDougall 2019, Dufresne and Geertsema 2020, Hungr and Evans 2004), and excess pore pressure (Collins and Reid 2019, Sassa and Wang 2005, Wang et al. 2002), have been identified as the possible reasons for the SFT, but the mechanism is still largely elusive because of the complexity of the geo-materials.

Numerous rockslide-debris flows have been reported around the world. Some typical events, such as the Ponti Peak landslide in India (Shugar et al. 2021), the Dujiangyan landslide (Yin et al. 2016) and Jiweishan landslide in China (Xu et al. 2010), and the Mount Meager landslide in Canada (Guthrie et al. 2012), have caused serious economic losses or death tolls to the local communities. Rockslide-debris flows tend to have catastrophic consequences because they are commonly characterized by extremely high speed (a few to tens of meters per second) and long run-out distance (several to tens of kilometers). These catastrophic events remind us the significance of making accurate risk assessment for the potential rockslide-debris flows, and this goal can be achieved only if we have a correct understanding of and can make accurate predictions for the run-out process of these landslides.

Numerical simulation is an efficient tool for the run-out analysis and prediction of rockslide-debris flow, and a variety of physically-based models have been adopted to simulate the run-out process of real landslide events. The models generally in use belong to three categories: 1) depth-averaged

71 models based on the finite difference method (FDM) (O'Brien et al. 1993, Ouyang et al. 2013, Sassa
72 et al. 2010, Shen et al. 2019, Shen et al. 2018), finite volume method (FVM) (Christen et al. 2010,
73 Mangeney et al. 2003, Xia and Liang 2018), or smoothed particle hydrodynamics (SPH) (Hungr
74 and McDougall 2009, Pastor et al. 2009); 2) discrete models originated from the discrete element
75 method (DEM) (Gao et al. 2021, Wu et al. 2018); and 3) three-dimensional models formulated
76 according to the SPH (Dai et al. 2017, Ghaitanellis et al. 2021), particle finite element method
77 (PFEM) (Zhang et al. 2015, Zhang et al. 2020) or material point method (MPM) (Li et al. 2021,
78 Soga et al. 2016, Xu et al. 2018). Among these models, the depth-averaged models are probably the
79 most sophisticated and frequently-used in the run-out analysis of real rockslide-debris flow events,
80 mainly because they are more time efficient. Specifically, the depth-averaged model can easily
81 consider entrainment (Cuomo et al. 2016, Iverson and Ouyang 2015, McDougall and Hungr 2005)
82 which is an important phenomenon in rockslide-debris flow modeling. The main difficulty in
83 modeling rockslide-debris flow is how to account for the SFT process. As mentioned above, the
84 mechanism of SFT is still quite elusive, so nearly no existing models can reflect the real physical
85 process of this phenomenon as far as we concerned. However, ignoring the influence of SFT may
86 lead to a wrong prediction of landslide mobility. A simple approach has been adopted in some
87 depth-averaged models to account for the influence of SFT by adopting different rheological models
88 for rockslide and debris flow (Gao et al. 2017, McDougall et al. 2006), and this strategy performed
89 well in improving the simulation results. Due to the above reasons, the depth-averaged models
90 should be more suitable choices for the run-out analysis of the rockslide-debris flow in this study.

91
92 Although many studies have analyzed the run-out processes of real rockslide-debris flow events
93 around the world (Gao, et al. 2017, Liang et al. 2020, McDougall, et al. 2006, Xing et al. 2014), few
94 of them have investigated the influence of the geomorphological factors such as slope gradient and
95 gully channel. Moreover, most of these studies did not provide in-depth analysis on the performance
96 of the models according to field measurements of entrainment and deposition.

97
98 In this paper we investigate a rockslide-debris flow event recently occurred in the Northern
99 Apennines of Italy (the Verghereto landslide). The landslide was surveyed soon after the failure and

100 a detailed map of entrainment and deposition was obtained from drone surveys. These data allow
101 validating an improved depth-average model capable of reproducing the complex behavior of the
102 landslide. The model considers entrainment and the influence of SFT is taken into account by
103 changing the basal frictional strength. The influences of slope gradient and the presence of a gully
104 channel on the run-out behavior are discussed, and some insightful conclusions are obtained.

105

106

107 **2. The Verghereto landslide**

108

109 **2.1 Geological settings**

110

111 The Verghereto landslide is located in the Northern Apennines of Italy, approximately 40 km to the
112 south of Cesena City. The area is characterized by steep slopes and deeply incised valleys carved by
113 rivers, with altitudes ranging from 600 m to 900 m above the sea level (Fig. 1).

114

115 The bedrock consists of deep marine flysch deposits belonging to the Marnoso-Arenacea Formation.
116 The Marnoso-Arenacea Formation is a turbidite succession representing the filling of the Miocene
117 Apennine foredeep complex, which deposited between the Langhian and the Tortonian (Ricci
118 Lucchi and Valmori 1980). It consists of alternating sandstones and marls layers in variable
119 proportion (Fig. 2a). In the study area, the ratio between coarse and fine strata is about 1/3 and the
120 average bed thickness varies from 0.5 m to 2 m. Both sandstones and marls are strong rocks
121 characterized by high resistance to compression (the uniaxial compressive strength of intact rock
122 specimens typically ranges from 40 MPa to 60 MPa) and high resistance to weathering. When the
123 bedding planes are horizontal or dip into the slope, the high strength of the rock mass ensures the
124 stability of the slopes and supports subvertical cliffs (Fig. 2b). Instead, large failures may occur
125 when the strata dip out of the slope. In this case the rock mass can slide along one controlling
126 bedding plane generating massive rockslides as in the study area.

127

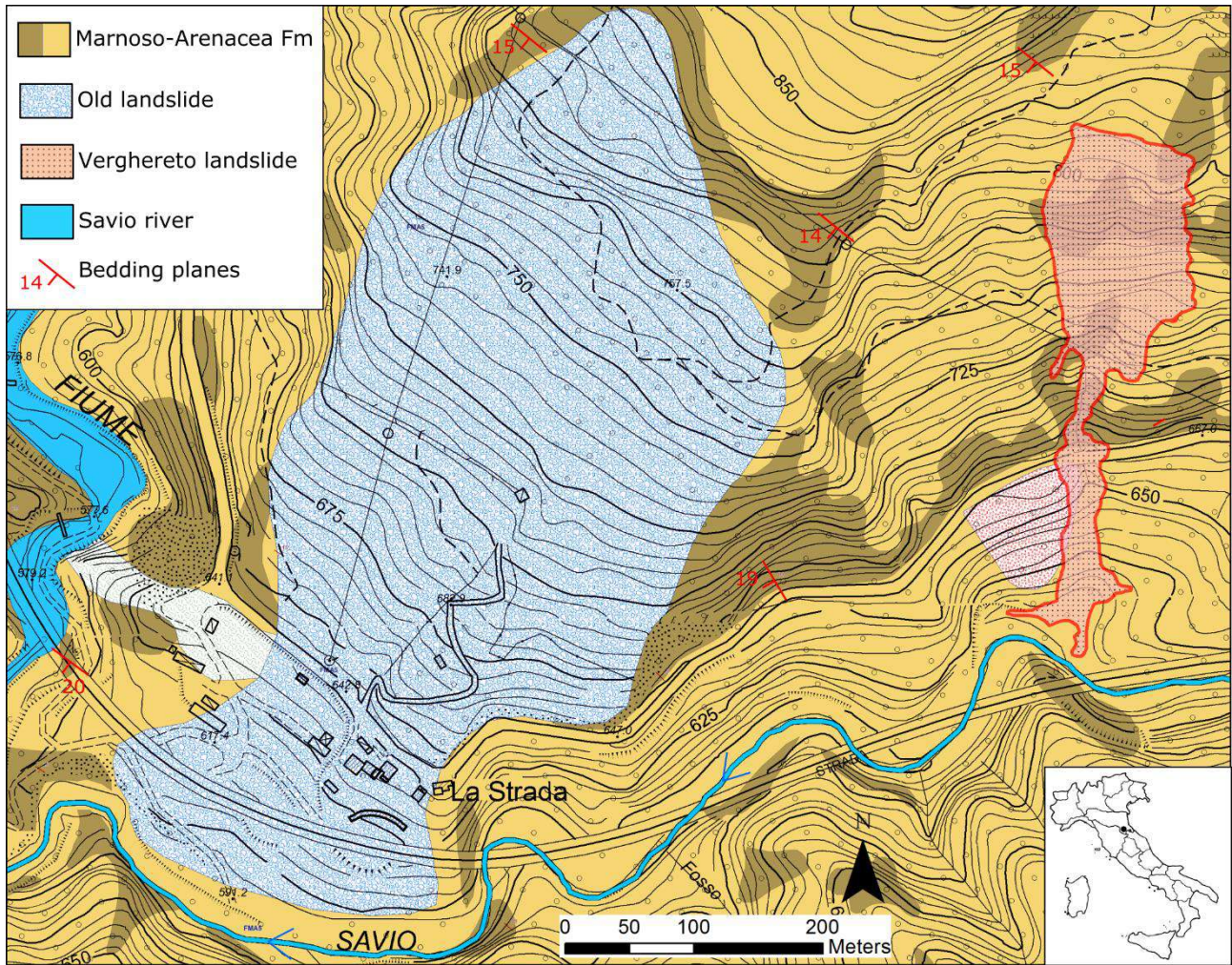
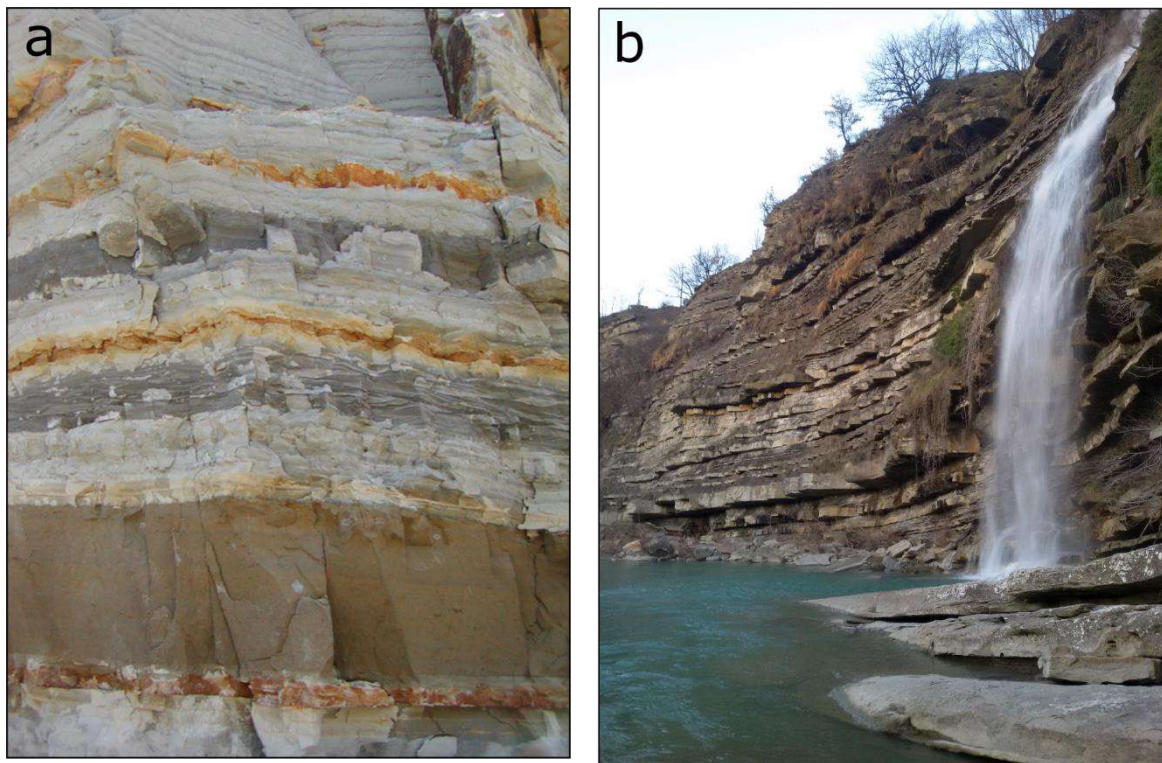


Fig. 1 Geological map of the study area and the locations of the Verghereto rockslide-debris flow and a giant old rockslide-avalanche adjacent to this landslide

Rockslides are the predominant form of instability in the area and are very common on cataclinal slopes where bedding dip is less than slope angle. These failures can occur on bed gradients less than 10° , which is approximately equal to half of the fully-softened angle of shearing resistance of the marls (Berti et al. 1994, Berti et al. 1996). In most cases the failed mass moves as a nearly intact block for a few tens of meters, retaining the original appearance and succession. Less commonly, the failed mass collapses generating dangerous flow-like landslide. In the study area both cases are present. The large landslide shown in the map of Fig. 1 is an old rockslide that did not turn into a flow, as suggested by the rectangular shape of the deposit and by lack of a transportation channel. Instead, the Verghereto landslide (in red) mobilized into a flow that advanced down a gully to the foot of the slope.



143

144 Fig. 2 Alternating sandstone and marls layers which consist of the sliding mass of the Verghereto
 145 rockslide-debris flow

146

147 2.2 The landslide

148

149 The Verghereto landslide occurred around 5:00 a.m. in the morning of May 13, 2019. A rock mass
 150 with a volume of nearly 40,000 m³ detached from the upper part of the slope sliding along a gently
 151 dipping bedding plane. The toe of the failed mass came out the slope, disintegrated into rock debris,
 152 and transformed into a debris flow that traveled downslope for about 300 m reaching the main river
 153 (Fig. 3a). The landslide destroyed a local road, 2.4 hectares of forest, and threatened the pylons of
 154 the highway that passes on the valley floor partially closing the River Savio. The landslide was
 155 triggered by a high-intensity short-duration rainfall that caused severe flooding and other landslides
 156 in the area. The failure was preceded by a rainfall of about 80 mm in 15 hours, with a peak intensity
 157 of 8 mm/h and a return period of 10-15 years (Fig.New1).

158

159 The map in Fig. 4 shows the three geomorphological zones that were identified in the field soon

160 after the event. Zone A is the source area of the landslide. It consisted of a rockslide that moved
161 essentially as a rigid block. Apparently, the slide did not acquire enough momentum to carry all the
162 rock mass beyond the foot of the slope, and about one third of the mass stopped at 30-40 m from the
163 detachment scarp. Sliding took place at a depth of about 10 m below the ground surface, at the top
164 of a marlstone layer dipping 15° to southwest (Fig. 3b). The rock exposed on the sliding surface
165 was fresh and stiff and we did not notice any appreciable difference with the other marls layers
166 outcropping on the trench walls. The lack of previous landslides indicated that the slide was a
167 first-time failure.

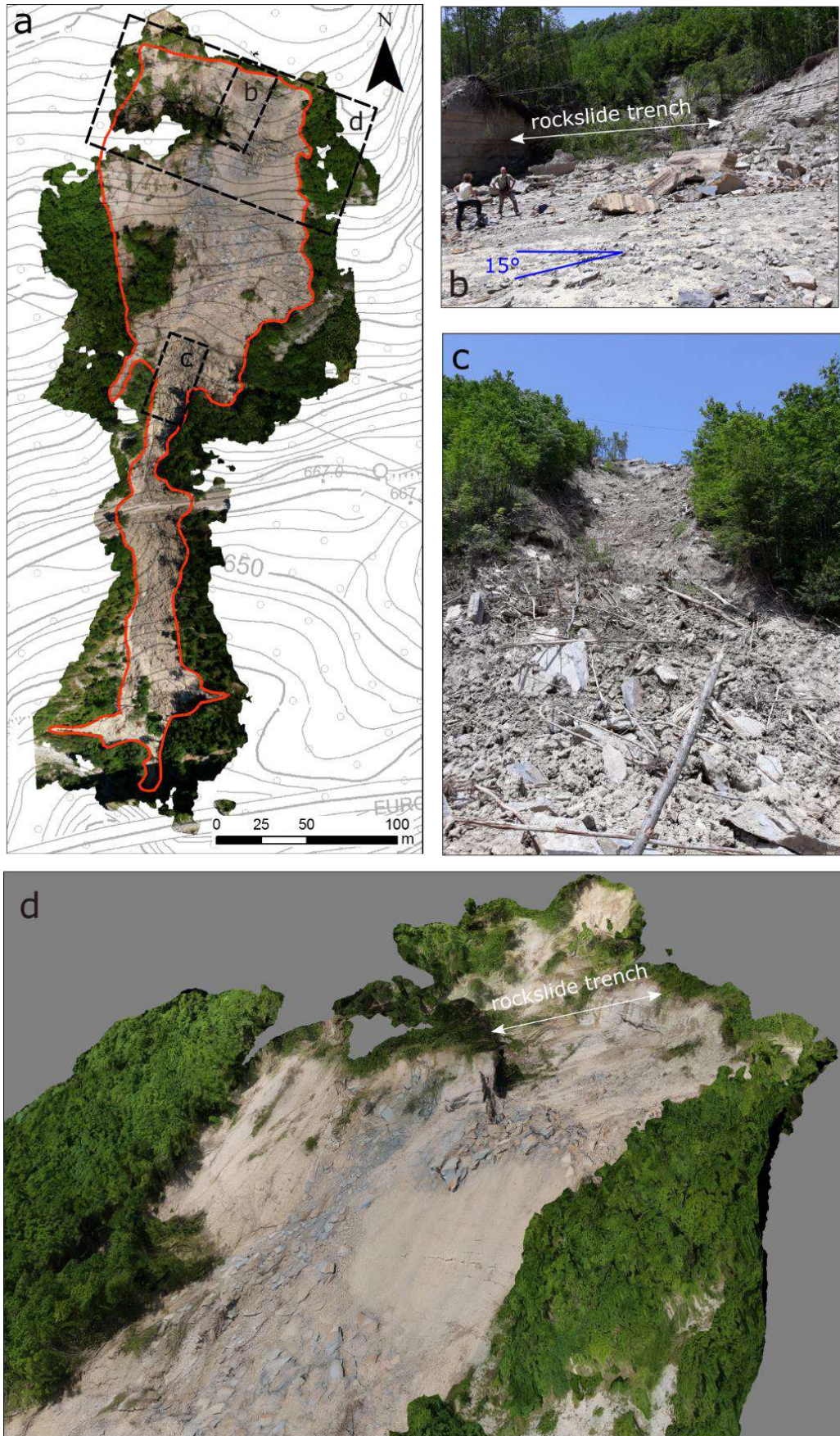
168
169 The front part of the rockslide collapsed and dropped about 28,000 m³ of fragmented rock to the
170 slope below (zone B in Fig. 4; Fig. 3d). Just below the source area, the slope is very steep (over 35°
171 degrees). Here the landslide stripped the vegetation and the soil cover over an area of about 4500 m²,
172 leaving evident scratches on the rock surface. Further downhill the slope angle decreases to less
173 than 30° allowing some crushed rocks and coarse debris to accumulate loosely in the lower part of
174 the zone.

175
176 Part of the landslide material then entered a small, ephemeral gully incised in colluvium and
177 mobilized into a debris flow (zone C in Fig. 4). Along the steep reach of the gully the debris flow
178 showed significant bulking by scouring and erosion and created a channel 15 m wide and 2-3 m
179 deep (Fig. 3c). As the gradient decreased to 20°-25°, the flow started to deposit within the channel
180 and came to rest at the foot of the slope. In the accumulation lobe the debris was on average 1-3 m
181 thick with an overall volume of approximately 15,000-20,000 m³. The presence of scouring, lateral
182 levees, and trees damaged or debarked by the impact with debris indicate that the flow was
183 extremely rapid.

184
185 One week after the failure we conducted a drone survey of the landslide. Five flights were done
186 using a DJI Spark UAV to cover an area of about 30.000 m². DJI Spark is a mini-drone equipped
187 with GPS/GLONASS positioning system and a 12 MP CMOS camera. Images were taken from a
188 flight elevation of about 30 m retaining an overlap of 80% along the flight path. Nine ground

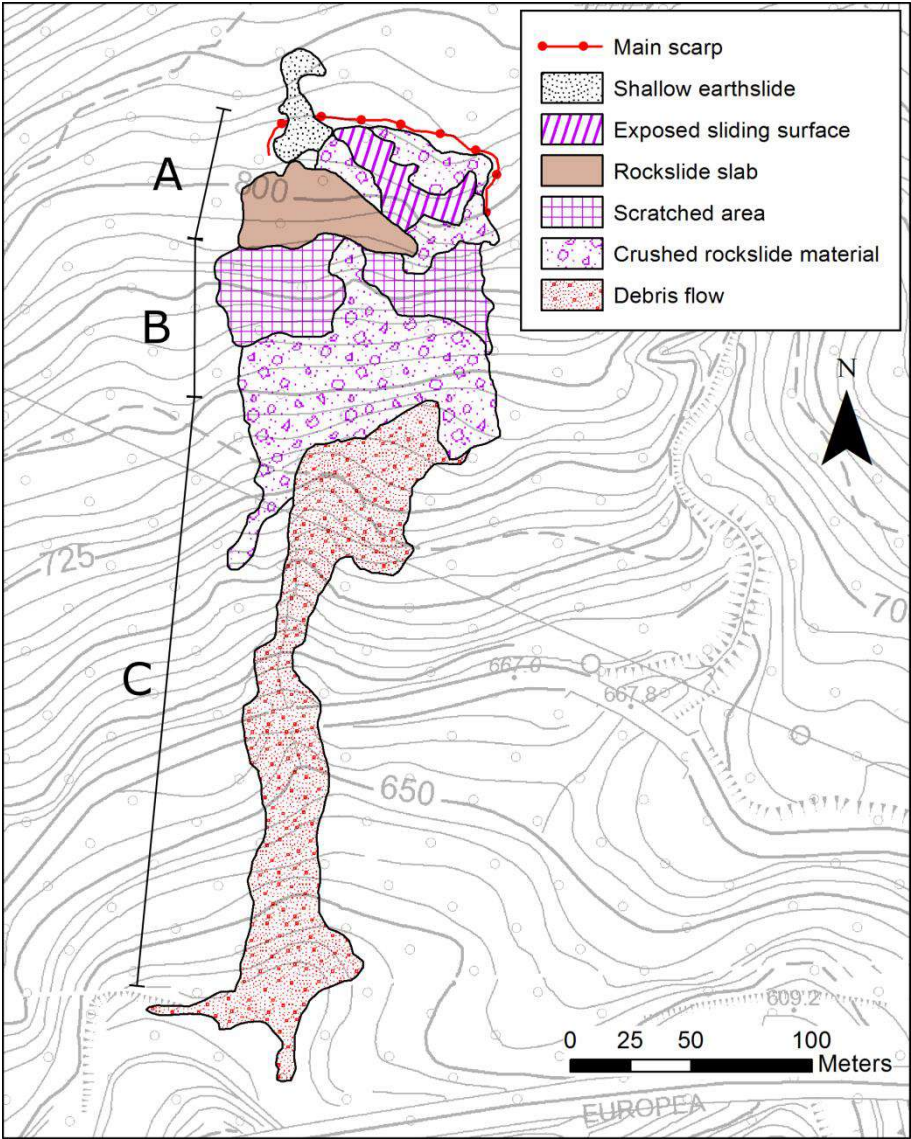
189 controls points were surveyed with a differential GPS receiver at the time of the flight. The
190 post-failure topographic model was obtained with the Structure-from-Motion photogrammetric
191 technique. Vegetated areas were masked out in the analysis in order to extract the digital terrain
192 model of the bare ground. The final model has a resolution of 5 cm/pixel and a total RMS error of
193 0.2 m. Post-failure topography was compared with the pre-failure digital terrain model (DTM)
194 available for the area with a 5 m resolution and a maximum error of 1.2 m. Estimates of change in
195 vertical elevation were finally computed by subtracting ground elevation of the two DTMs
196 (Difference of DTMs, DoD; Fig. New2). Errors from individual DTMs propagate into the DoD
197 resulting in a minimum level of change detection of about 1.5 m (rounded square root of the sum of
198 squares of individual errors).

199 The DoD (Fig. New2) clearly show the deep trench behind the rockslide body and the displaced rock
200 block in zone A. The translational movement of the rockslide created negative topography in the
201 trench and positive topography in the deposit. Zone B is characterized by a slight positive
202 topography generated by the debris accumulated on the scratched ground surface. Zone C shows a
203 complex alternation of erosion and deposition caused by the debris flow. In these zones, however,
204 elevation changes are difficult to interpret because of the low accuracy of pre-failure topographic
205 data.



206

207 Fig. 3 a) Top view of the rockslide, b) the trench exposed in the source zone after the occurrence of
 208 the rockslide, c) debris deposit in the channel, and d) the steep slope below the source zone



210

211 Fig. 4 Characteristics of the deposit in different regions of the landslide influenced zone and the
212 three geomorphological zones identified after the event. A is the source area of the landslide, B is
213 the transformation zone, and C is the debris flow zone

214

215 **3. Methodology**

216

217 **3.1 Numerical model**

218

219 An improved finite difference model (Shen, et al. 2018) is adopted to simulate the run-out process
220 of the Verghereto landslide. This model is built in a global Cartesian coordinate, with the positive

221 direction of z axis parallel to the opposite direction of gravity. Similar to the typical depth-averaged
 222 models, this model consists of one mass balance equation and two momentum balance equations,
 223 which are given by:

$$224 \quad \frac{\partial h}{\partial t} + \frac{\partial Q_x}{\partial x} + \frac{\partial Q_y}{\partial y} = -\frac{\partial Z}{\partial t} = \frac{\tau_b - \tau_e}{\rho_e \sqrt{v_x^2 + v_y^2}} \quad (1)$$

$$225 \quad \frac{\partial Q_x}{\partial t} + \frac{\partial Q_x^2 / h}{\partial x} + \frac{\partial Q_x Q_y / h}{\partial y} = -\frac{\partial k_x g h^2 / 2}{\partial x} + \frac{(Ag + B)h \tan \alpha}{\tan^2 \alpha + \tan^2 \beta + 1} - \frac{\tau_b A_b h v_x}{m \sqrt{v_x^2 + v_y^2 + v_z^2}} \quad (2)$$

$$226 \quad \frac{\partial Q_y}{\partial t} + \frac{\partial Q_x Q_y / h}{\partial x} + \frac{\partial Q_y^2 / h}{\partial y} = -\frac{\partial k_y g h^2 / 2}{\partial y} + \frac{(Ag + B)h \tan \beta}{\tan^2 \alpha + \tan^2 \beta + 1} - \frac{\tau_b A_b h v_y}{m \sqrt{v_x^2 + v_y^2 + v_z^2}} \quad (3)$$

227 where: h is flow depth; $Q_x = v_x h$ and $Q_y = v_y h$ are mass fluxes in x and y directions; v_x , v_y and v_z are
 228 depth-averaged velocities in x, y and z directions; k_x and k_y are lateral pressure coefficients in x and
 229 y directions determined according to soil state (Ouyang et al. 2015); g is gravitational acceleration;
 230 A and B are terms related to static and centrifugal/centripetal normal forces on bed; α and β are dip
 231 angles in x and y directions; τ_b is the basal shear stress of flow; τ_e is the shear stress in erodible
 232 mass; ρ_e is the bulk density of entrained mass; A_b is the bottom area of a control volume; m is the
 233 mass of flow in the control volume. The expressions of A , B , A_b , τ_b and τ_e are given by:

$$234 \quad A = 1 + \frac{\partial k_x h^2 / 2}{\partial x} \tan \alpha + \frac{\partial k_y h^2 / 2}{\partial y} \tan \beta \quad (4)$$

$$235 \quad B = \frac{C_x}{\cos \alpha} \left(\frac{v_x}{\cos \alpha} \right)^2 + \frac{C_y}{\cos \beta} \left(\frac{v_y}{\cos \beta} \right)^2 \quad (5)$$

$$236 \quad A_b = \Delta x \Delta y \sqrt{\tan^2 \alpha + \tan^2 \beta + 1} \quad (6)$$

$$237 \quad \tau_b = \sigma(1 - r_{ub}) \tan \varphi'_b + c'_b \quad (7)$$

$$238 \quad \tau_e = \sigma(1 - r_{ue}) \tan \varphi'_e + c'_e \quad (8)$$

239 in which: C_x and C_y are bed curvatures in x and y directions; Δx , Δy are the sizes of a control
 240 volume in x and y directions; σ is the normal stress on bed; r_{ub} and r_{ue} are the pore pressure
 241 coefficients (the ratio of the pore pressure to the total normal stress) in flow bottom and erodible
 242 mass; φ' and c' are effective frictional angle and cohesion. The subscripts b and e refer to flow
 243 bottom and erodible mass, respectively.

244

245 A finite difference scheme is utilized to solve the above governing equations, and the details of the
246 numerical scheme could be found in Shen, et al. (2018).

247

248 3.2 Simulation setup

249

250 . Within the area we selected a region which covers the whole run-out zone of the landslide as the
251 computational domain. The size of this domain is 522 m in x direction (N-S) and 291 m in y
252 direction (E-W). Uniform computational grids 3 m long in both x and y directions are adopted in the
253 present study, and the maximum time step is 0.02 s.

254

255 According to the landslide characteristic described in Section 2, we divided the computational
256 domain into two regions (Fig. 5). The first region is the area above the gully head ($x < 270$ m),
257 which include the source zone of the rockslide and the steep slope below (zones A and B in Fig. 3).
258 The second region is the zone below the gully head, where the rockslide turned into a debris flow
259 (zone C in Fig. 3). According to our field observations, the landslide essentially moved like a solid
260 in the first region and like a flow in the gully.

261

262 Although the transformation of rockslide and debris flow is gradual rather than sudden, in order to
263 simulate the complex behavior of the Verghereto landslide with a single-phase model, we must
264 necessarily assume different material properties in the two regions. A simple way to do it is to
265 assign a high frictional strength in region 1 (where the landslide moved like a slide) and a low
266 frictional strength in region 2 (where the landslide moved like a flow). Different values of the
267 frictional strength were obtained by adopting different values of the pore pressure coefficient r_u in
268 the two regions. In particular, we used a pore pressure coefficient of zero to simulate the high
269 frictional strength at the base of the landslide, and a pore pressure coefficient of 0.3 to simulate low
270 frictional strength.

271

272 This assumption is basically reasonable, since the basal pore pressure is usually higher when a

landslide is in fluid state than in solid state. All the other model parameters (density, friction and cohesion) were assumed to be identical in the whole domain.

Based on these assumptions, three groups of simulation were conducted using the parameters listed in Table 1. According to field investigation, the thickness of erodible soil cover was set to be 1.5 m in the whole region except in the source zone of the rockslide where the bedrock outcrops. The erodible mass is required to have a higher pore pressure than the sliding mass in order to be entrained. Here, erodible mass is assumed to have the same effective strength parameters (c' and φ') as the sliding mass, while its pore pressure coefficient r_{ue} takes a higher value ($r_u=0.8$).

Table 1 Parameters for simulating the Verghereto landslide

Group	Pore pressure coefficient in R1 r_{u1}	Pore pressure coefficient in R2 r_{u2}	Basal effective cohesion c' (kPa)	Basal effective friction angle φ'
S1	0.0	0.0	5	30
S2	0.3	0.3	5	30
S3	0.0	0.3	5	30

Notes: R1 and R2 refer to Region 1 and Region 2, respectively.

4. Results

4.1 Depositional characteristics

In Fig. 5, we illustrate the difference between the pre-failure and post-failure topography obtained from simulations (a-b-c) and measurements (d). The analysis S1 simulates a landslide with high frictional resistance at the base ($r_u=0$ in the whole domain). In this scenario, the landslide stops in the upper part of the slope and reaches a much smaller run-out distance than that observed in the field. However, the computed depositional pattern agrees well with the survey data in Zone A and B (comparing simulation results with Fig. 4 and Fig.3d). As mentioned above, in region 1 the

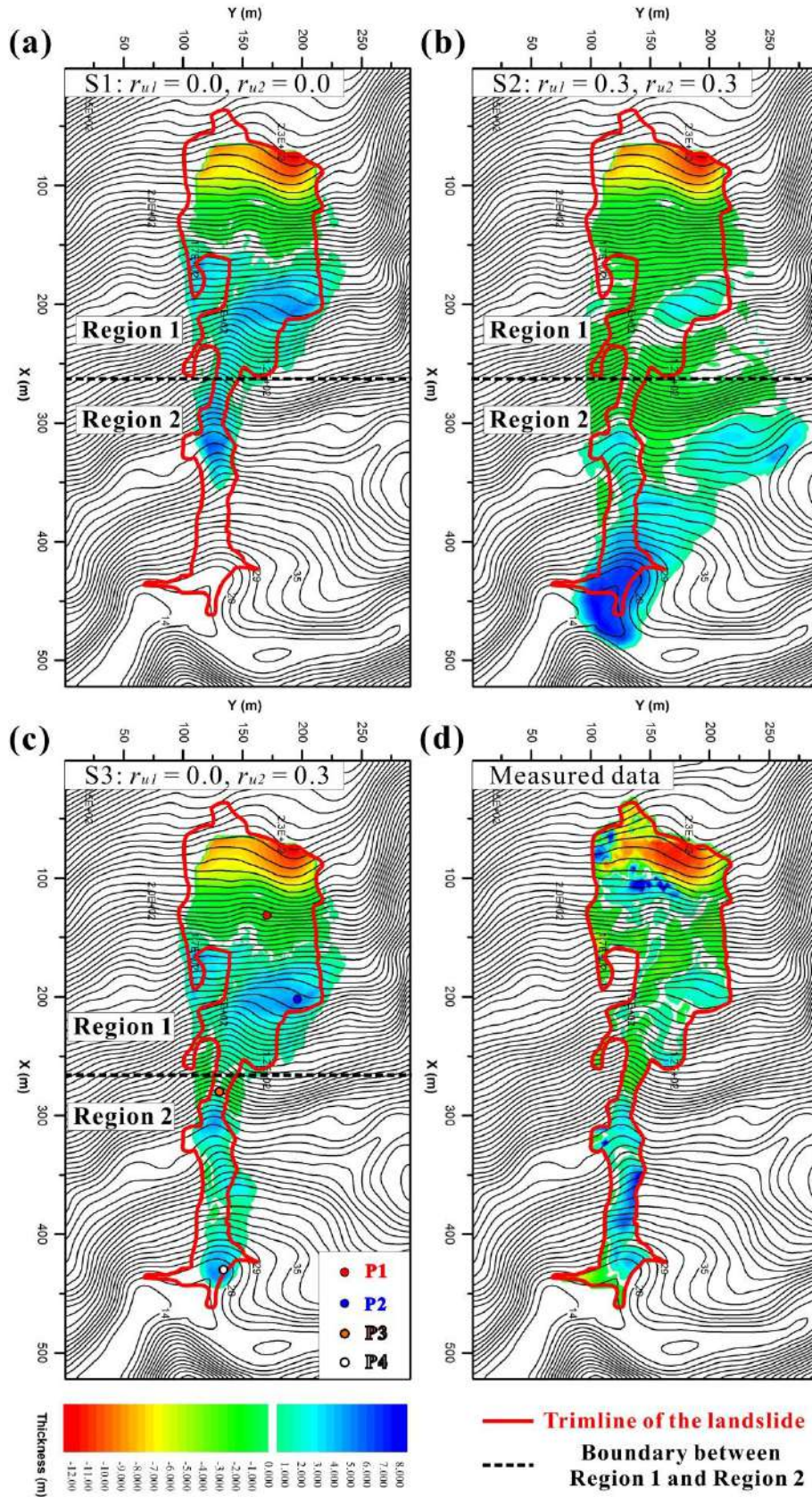
295 landslide caused erosion on steep slopes and deposition on gentle slopes. The model captures this
296 spatial variability, showing entrainment on the steep slopes right below the source zone (negative
297 DEM difference ranges from -1.0 m to -1.5 m) and deposition on the gentle slopes further downhill
298 (positive DEM difference ranges from 4.0 m to 5.0 m).

299

300 By assuming low frictional strength in the whole domain ($r_u=0.3$; simulation S2) the model predicts
301 a larger mobility of the landslide (Fig. 5b). In this case, the landslide spreads over a much broader
302 region than the measured one, leading to a significantly inaccurate prediction of the run-out. In
303 particular, the landslide runs downslope laterally rather than flowing into the gully (Fig. 5b). These
304 results indicate that in region 1 the frictional strength at the base of the landslide should be
305 relatively high, so that most of the fragmented material can come to rest in this area.

306

307 This is confirmed by the results of simulation S3, which provides the best agreement with reality.
308 By adopting a high friction in region 1 and a low friction in region 2 simultaneously, part of the
309 material stops below the source area and part continues downslope as a flow. With this combination
310 of r_u the model can simulate the debris flow (zone C in Fig. 4) and the landslide reaches a run-out
311 distance similar to the measured one (Fig. 5d). Moreover, the model correctly predicts erosion in the
312 upper reach of the channel and deposition in the lower reach, where the slope becomes gentle (Fig.
313 5c). The above analysis indicates that friction change caused by the SFT plays a significant role in
314 the run-out behavior of this landslide. And we may not be able to correctly simulate the run-out
315 process if the SFT is neglected. Additionally, although the single-phase model cannot actually
316 depict the complicated physical process of the SFT, the above simple method could improve the
317 simulation results of those landslides involving such a complex SFT phenomenon by adopting
318 frictional strengths for the sliding mass under the two different states (solid and fluid states).



319

320 **Fig. 5** Digital elevation differences between pre-failure and post-failure topographies in the

321 landslide zone obtained from simulations and field survey

322

323 4.2 Velocity and entrainment

324

325 The total average velocity and entrainment time curves of the landslide in the three simulations
326 (S1-S3) are illustrated in Fig. 6. The four turning points shown on the average velocity curve of
327 simulation S3 (Fig. 6a) indicate the first velocity peak (t_1), the turning point between the first
328 deceleration stage and the second acceleration stage (t_2), the second velocity peak (t_3), and the time
329 when the motion of the landslide basically stops (t_4). From 0 s to t_1 , the landslide accelerates rapidly
330 after it detaches from the bedrock and propagates to the steep slope just below the detachment area.
331 Then the landslide reaches a low slope gradient area (lower part of zone B, Fig. 4), resulting in a
332 dramatic drop of the average velocity from t_1 to t_2 . The landslide enters the gully head at around t_2 .
333 Here the model predicts a second slight acceleration stage (from t_2 to t_3) which should be attributed
334 to both SFT and the steep topography in the downstream part of the gully head. Finally (from t_3 to t_4)
335 the landslide comes to rest gradually. The difference between simulations S1 and S3 is that the
336 second acceleration stage does not exist in S1, since in this case the reduction in frictional strength
337 is not taken into account. Therefore, in S1 the landslide stops quickly after entering the gully
338 showing a small run-out distance. By contrast, in S2 the landslide runs too fast and too distant, and
339 the predicted velocity and entrainment are clearly overestimated.

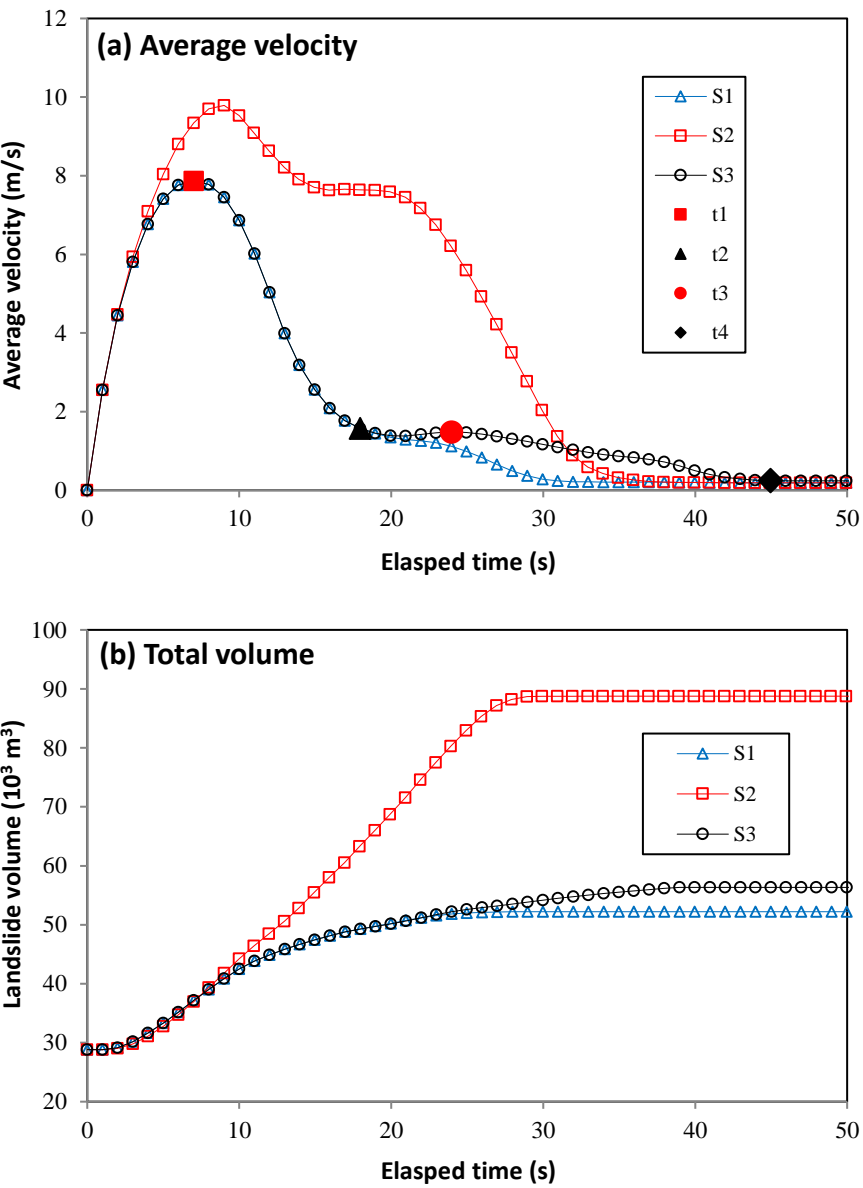
340

341 The total volume curve of simulation S3 (Fig. 6b) indicates that the landslide is likely to have
342 entrained a large amount of loose soil before reaching to the gully head. This extra volume from
343 entrainment may potentially generate the source material for the mass flow in the gully. The volume
344 of the landslide probably doubled (from approximately 28,000 m³ to around 56,000 m³) through
345 entrainment.

346

347 The thickness and entrainment distributions of the landslide (in group S3) at the above four
348 moments (t_1 - t_4) are presented in Fig. 7. The thickness distributions at the four moments support our
349 above analysis toward the velocity change process of the landslide. By contrast, the entrainment
350 distribution characteristic of the landslide is relatively simple (Fig. 7), indicating the landslide may

351 entrain almost all the superficial loose mass on the slope wherever the landslide runs over.
352



354
355 **Fig.6** Simulated time curves of the (a) average velocity and (b) total volume of the Verghereto landslide
356

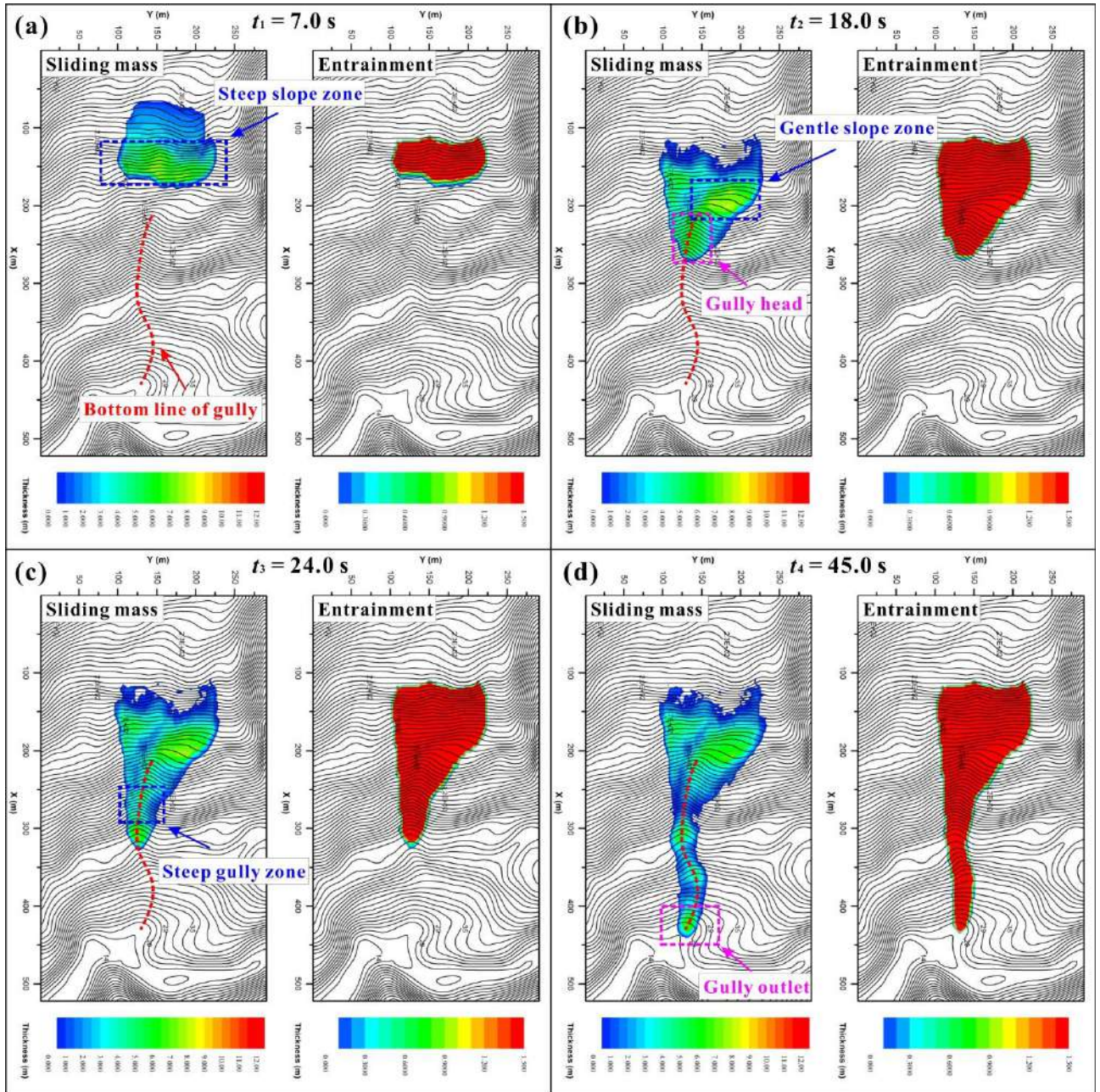


Fig. 7 Thickness and entrainment distributions of the landslide at four moments in S3. t_1 corresponds to the time when the landslide has the highest kinetic energy, t_2 is the time when part of the landslide starts to enter the gully head, t_3 is the time when the landslide reaches the second velocity peak, and t_4 corresponds to the time when the landslide basically comes to rest.

4.3 Influence of topography

Four numerical gauge points (P1-P4) in the landslide area are selected to analyze the influence of topography (slope gradient and gully channel) on the simulated dynamic characteristics (thickness,

367 velocity and Froude number) of the landslide at four different locations, and the simulation results
368 of group S3 is used to conduct this analysis. The locations of these gauge points are shown in Fig.
369 5c. P1 and P2 are located in region 1 where the landslide is in ‘solid state’, while P3 and P4 are
370 points located in region 2 where the landslide has transformed into a debris flow. P1 and P3 are
371 approximately in the middle part of the steep slope gradient zones in region 1 and region 2 (in
372 gully), respectively. For comparison, P2 and P4 are selected from the low slope gradient zones in
373 region 1 and region 2, respectively. The flow thickness h and depth-averaged velocity v are directly
374 available from the simulation results, while the Froude (Fr) number is calculated using $Fr = v/\sqrt{gh}$.
375 Fr number is a dimensionless variable reflecting the relationship between flow inertia and gravity.
376 At P1 and P3 where the slopes are steep, the thickness of the landslide (Fig. 8a) increases fast when
377 the front of the landslide arrives, and then decreases gradually to less than 1 m. The final DEM
378 differences (net change in elevation) in these zones are less than zero, so the deposit there has an
379 appearance of entrainment. However, at P2 and P4 the sliding mass accumulates and finally stops
380 propagating, demonstrating an opposite appearance of deposition. Actually, entrainment should
381 occur in both steep and gentle slope gradient regions, but the apparent entrainment is only revealed
382 in steep slope zones. The thickness of the landslide at P3 remains at a relatively stable and thin level
383 (about 1.5 m) which lasts for around 15 s after the arrival of landslide front, while at P1, the
384 thickness decreases quickly after the arrival of landslide front. These different thickness curves
385 indicate that on the steep slopes in Region 1 (P1) the landslide propagates like a surge wave, while
386 on the steep slopes in Region 2 (P3), the landslide probably behaves like a plug flow due to the
387 confinement of lateral propagation from the gully channel. The velocity curves (Fig. 8b) illustrate
388 that the velocity of landslide is generally higher when it propagates on steep slopes than on gentle
389 slopes (Fig. 8b). And the peak velocity of landslide on steep slopes (around 8.0 to 9.0 m/s) is about
390 twice of the peak value on gentle slopes (approximately 4.0 to 5.0 m/s). The Fr number curves at P2
391 and P4 are similar. At P2 and P4, the Fr number peaks at the arrival of landslide front, and then
392 decreases quickly because the sliding mass accumulates and comes to rest on the gentle slopes. By
393 contrast, the Fr numbers at P1 and P3 show some different tendencies. At P1 and P3, the Fr number
394 peaks when the landslide front arrives, and then the number drops quickly until it rises up again.
395 After reaching at the first valley value, at P1 the Fr number increases rapidly to a second peak larger

396 than the first one, and then the number slumps to a low value. However, at P3 the Fr number
 397 increases only slightly until reaching at a relatively steady value (around 1.5) which lasts
 398 approximately 10 s, and then gradually decreases to a low value. These difference tendencies on Fr
 399 number between P1 and P3 is probably caused by the presence of the gully channel and its
 400 influence on the dynamic process of a landslide. This influence from gully will be discussed in the
 401 Discussion section. In summary, the topography on the path has a significant influence on the
 402 dynamic characteristic of this landslide. At locations on steep slopes, the landslide passes over
 403 quickly and finally shows entrainment. Conversely, at low slope gradients regions, the landslide
 404 comes to rest fast and eventually produces deposition. The existence of a gully channel also alters
 405 the dynamic characteristic of the landslide.

406

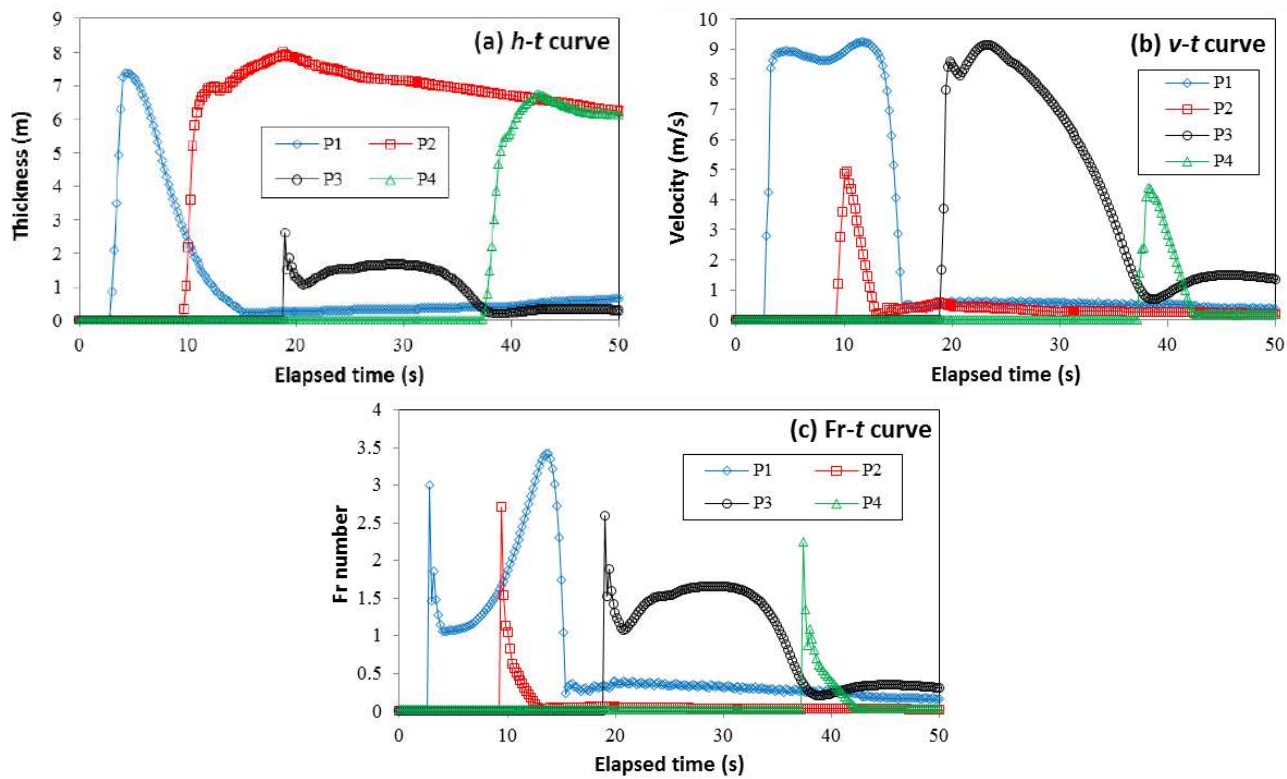


Fig. 8 Time curves of thickness, velocity and Froude (Fr) number of sliding mass at four locations P1 to P4. P1 is on the steep slope of Region 1, P2 on the low slope gradient zone in Region 1, P3 on the steep slope at the gully head and P4 on gully outlet.

407

408 5. Discussion

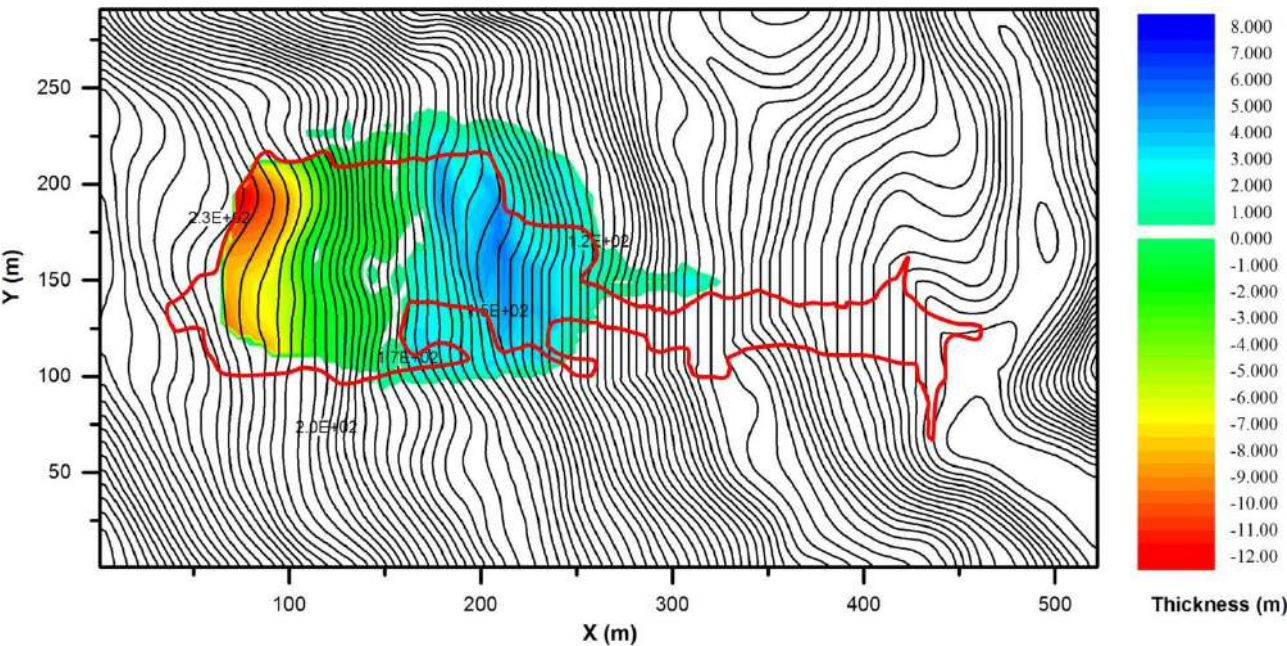
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410 As mentioned above, the existence of a gully channel may play an important role in determining the
411 dynamic characteristic of a landslide. In field, we observed an interesting phenomenon that the
412 existence of a gully seems to increase the final run-out distance of a rockslide. In rockslides with
413 similar geological setting, those rockslides have a gully on the slope, similar to the Verghereto
414 landslide in this study, tends to have a larger run-out distance which usually extends to the slope toe,
415 while the rockslides without a gully normally deposit in the middle part of the slope which is far
416 away from the slope toe.

417
418 To illustrate the influence from channel, in this section we simulate the Verghereto landslide in the
419 condition of without a channel. Then four imaginary numerical tests (S1-nTnG, S2-nTwG,
420 S3-wTnG and S4-wTwG) are conducted to investigate the generalized scenarios.

421
422 The simulation setups of the Verghereto landslide without the presence of a channel are the same as
423 those in S3 expect for the topography. In this simulation, the channel on the slope is artificially
424 removed by adjusting the elevation around the gully. The result is present in Fig. 9, which obviously
425 shows a reduction in the run-out distance due to the absence of a channel.

426



427
428 Fig. 9 Simulation result of the Verghereto landslide in the condition without the presence of a
429 channel on the slope

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The simulation setups for the other four generalized numerical tests are listed in Table 2. The schematic diagram of these numerical tests is shown in Fig. 10. In these tests, a 10 m thick, 80 m wide and 60 m long rock block is assumed to detach from the bedrock in a rock scarp and forms a rockslide. Then the rockslide propagates on a 25° slope next to the rock scarp. The landslide area is divided into two regions similar to what we have done in the simulation of the Verghereto landslide. The slope above the gully head ($x > 200$ m) is region 1, while the slope below the gully head is region 2. The pore pressure coefficients in these two regions have different combinations in different groups (Table 2).

440 **Table 2** Parameters for simulating ideal soil collapse experiments

Group	Existence of A channel	Pore pressure Coefficient in R1 r_{u1}	Pore pressure Coefficient in R2 r_{u2}	Basal effective cohesion c' (kPa)	Basal effective friction angle φ'
S1-nTnG	No	0.25	0.25	5	30
S2-nTwG	Yes	0.25	0.25	5	30
S4-wTnG	No	0.25	0.40	5	30
S4-wTwG	Yes	0.25	0.40	5	30
Notes: R1 and R2 refer to Region 1 and Region 2, respectively.					

441

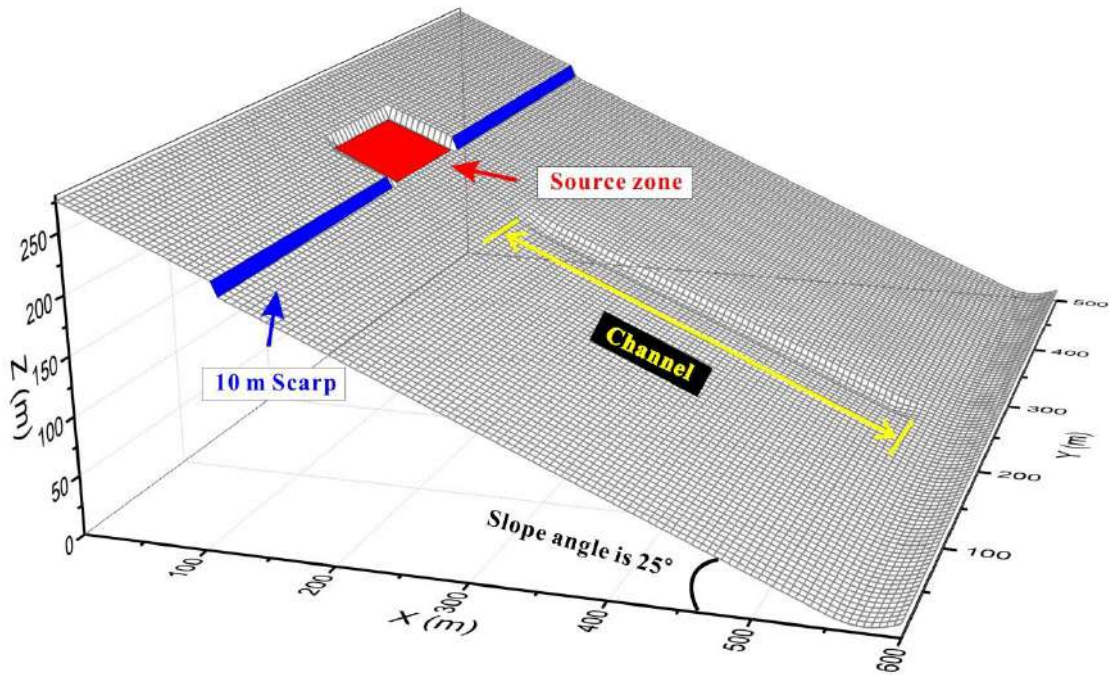
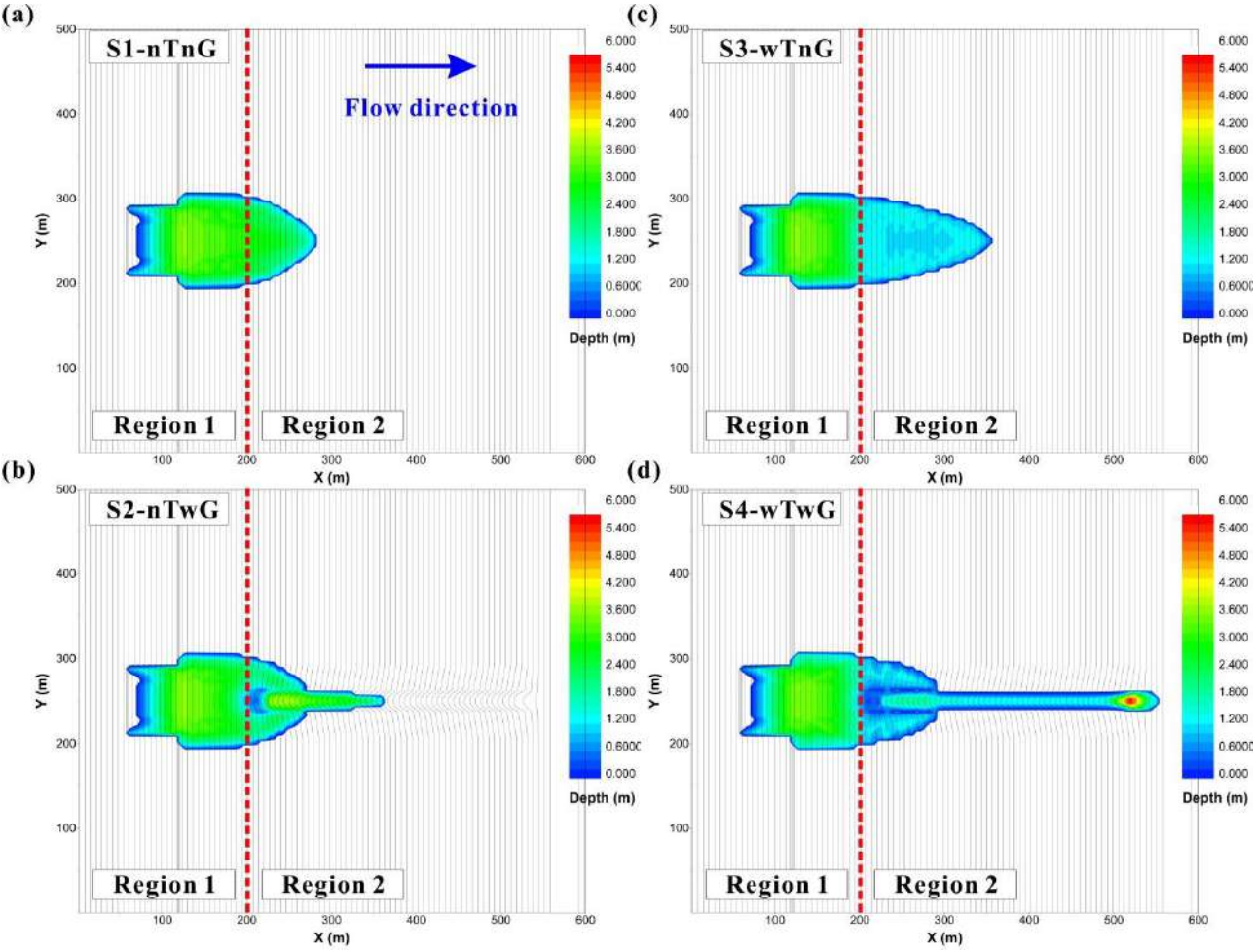


Fig. 10 Schematic diagram of the ideal rock collapse experiment with a channel on the slope

The simulation results illustrate that the existence of a channel can obviously increase the run-out distance (Fig. 11), no matter there is SFT or not. Without SFT and the channel (Fig. 11a), the front of final deposit reaches to $x = 280$ m, while the final landslide front can reach to $x = 360$ m if there is a channel on the slope (Fig. 11b). Similarly, when there is SFT, the rockslide will have a larger run-out distance than that one without SFT (Fig. 11a), but the movement of the rockslide still will stop on the middle part of the slope while there is no channel on the slope (Fig. 11c). By contrast, the landslide may reach to the slope toe if there is a channel (Fig. 11d). These results indicate that SFT may not be the only factor contributing to the large run-out distance of the landslide. The existence of a gully can also promote the run-out distance. The promotion effect of a gully on the run-out distance of a landslide may simply because the gully constrained the lateral spreading of the landslide. When the landslide propagates on a relatively uniform slope (Fig. 11a and 11c), it propagates forward and laterally simultaneously. The lateral spreading process will consume part of the kinetic energy, so the deposit has a smaller run-out distance but a larger lateral spreading area. Conversely, the channel reduces the lateral spreading and the energy consumption caused by it, so the landslide reaches a larger run-out distance but a much smaller lateral spreading area. This conclusion could be helpful for us to conduct a quick prediction on the risk of potential rockslides

461 similar to the Verghereto rockslide. Those with a gully channel on the slope may pose higher risk on
 462 the infrastructures in the gully outlet (slope toe), while the potential rockslides without a gully on
 463 slope mainly endanger the properties in the middle part of the slope. The average velocity curves of
 464 these numerical tests agree with the above analysis (Fig. 12). The existence of a channel can reduce
 465 the rate of deceleration and produce a larger run-out distance.
 466



467
 468 **Fig. 10** Simulation results of ideal rockslides on a slope with different parametric and topographic
 469 conditions. The four graphs correspond to (a) without both SFT a channel and, (b) without SFT but
 470 with a channel, (c) with SFT but without a channel, and (d) with both SFT a channel
 471

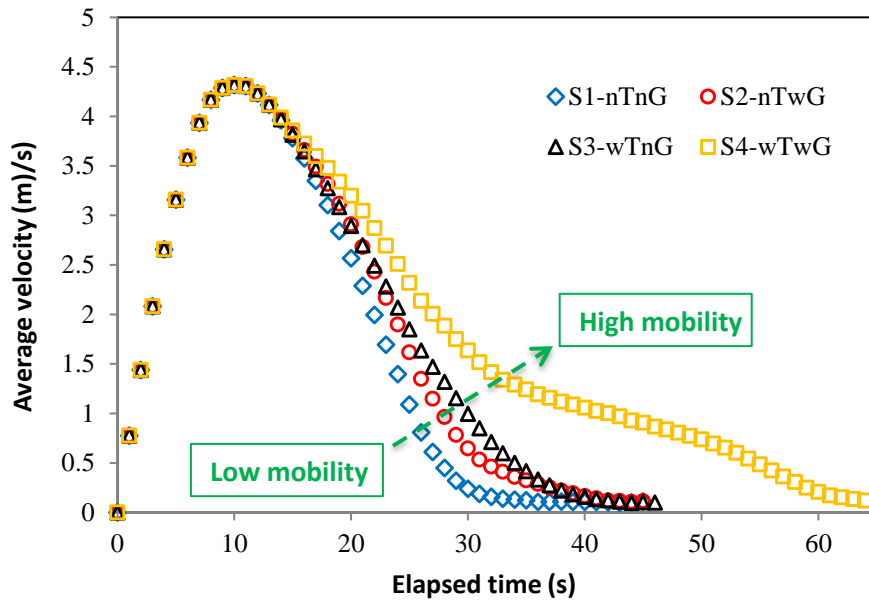


Fig. 11 Average velocity-time curves of the collapsed soil in the four simulation conditions. From S1 to S4 the mobility of the sliding mass increases gradually

6. Conclusions

The run-out process of a rockslide-debris flow in a layered rock slope is studied by an improved finite difference model. Field investigation and numerical simulations on this landslide are conducted to interpret the propagation process, and we obtain the following conclusions.

(1) The run-out process of the Verghereto landslide can be divided into three stages. In the first stage, the landslide detached from the bed rock sliding on the relatively gentle surface in the source zone. Then in the second stage, the landslide descended quickly in the steep slope zone next to the source zone before slumping heavily on the low slope gradient zone, and in the meantime, the volume of the landslide increased by entraining the loose mass on the slope and the rock mass disintegrated quickly. In the final stage, the disintegrated rock mass converged into the gully and transformed into a debris flow, and then the flow propagated along the gully until it stopped at the outlet of the gully.

(2) Simulation results show that the frictional strength change produced by the SFT process probably performs an important role in determining the dynamic characteristics of this landslide.

The run-out behavior and depositional characteristic of the landslide can be correctly simulated if we properly consider this friction strength change of sliding mass. The depth-averaged single-phase model adopted in this study performs well in the simulation of the Verghereto rockslide-debris flow.

(3) Topography may have a dominant impact on the depositional characteristic of the Verghereto landslide. In the landslide area where the slope is relatively steep, the final digital elevation difference shows entrainment. By contrast, in low slope gradient zones, the deposit shows accumulation and deposition. However, bed entrainment should occur on both steep and gentle slopes. Additionally, the existence of a gully channel on the slope could enlarge the run-out distance of the landslide. In the potential rockslides similar to the Verghereto landslide, those with a gully on the slope may pose higher risk to the infrastructures in the outlet of the gully (at slope toe).

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