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Modelling the mobility and dynamics of a large Tahitian landslide using runout distance

Julien Gargani^{1,2}

¹Univ. Paris-Saclay, CNRS, Geops, Orsay, France

²Univ. Paris-Saclay, Centre d'Alembert, Orsay, France

julien.gargani@universite-paris-saclay.fr

Highlights

- One or a few landslide events created the large scar and mass deposits in Tahiti.
- Peak slide velocity reached 125–250 m/s.
- The high mobility related to an effective basal friction of 0.2–0.3 under submarine conditions.

Abstract

Runout distance was used here to determine whether the large landslide deposit formed several thousand years ago in northern Tahiti was caused by a single or multiple events. Using modelling to quantify the dynamics of this event suggested that a single event or a small number of events ($n < 10$) were responsible, and that the maximum slide velocity was high (> 125 m/s) under partially submarine conditions. Such submarine propagation favoured a slower dynamic but a longer runout. The effective basal friction under submarine conditions ranged from $0.2 < \mu < 0.3$.

Keywords : runout, volcano, modelling, landslide, Tahiti

1. Introduction

28 Various hypotheses have been proposed to explain the triggering of large landslides: (i)
29 ground acceleration due to earthquakes (Keefer, 1994; Roberts and Evans, 2013) or volcanic
30 eruptions (Carracedo et al., 1999), (ii) new stress triggered by dyke intrusion or magma reservoir
31 deformation (Carracedo et al., 1999; Gargani et al., 2006; Hampel and Hetzel, 2008; Le Corvec
32 and Walter, 2009), (iii) increased pore pressure associated with higher rainfall, sea level or lake
33 level (Muller-Salzburg, 1987; Mc Mutry et al., 2001; Cervelli et al., 2002; Kilburn and Petley,
34 2003; Veveakis et al., 2007; Quidelleur et al., 2008; Crozier, 2010; Gargani et al., 2014; Iverson et
35 al., 2015), and (iv) slope change due to deep erosion or vertical movement. The complexity and
36 inter-relatedness of the processes involved often prevent clear conclusions regarding the precise
37 causes of slope failures.

38 The nature of large landslide dynamics along the slide path is still under debate because the
39 mobility of these landslides is often larger than expected (Legros, 2002; Lucas and Mangeney,
40 2007; Iverson et al., 2015). In this study, the mobility of a landslide that occurred in Tahiti, French
41 Polynesia has been investigated. It may be necessary to consider complex processes of various
42 origins to explain such long-distance sliding, such as high basal pressure (Goren and Aharonov,
43 2007), high-fluidity layer or air-layer lubrication (Johnson, 1978), ground vibration or
44 hydroplaning (Hürlimann et al., 2000), presence of ice or evaporites (Bigot-Cornier and
45 Montgomery, 2007; Crosta et al., 2018), and submarine or aerial propagation (De Blasio, 2011b).
46 The complexity of these processes also leads to difficulties in modelling landslides. For example,
47 the friction angle of actual materials is often higher than that of materials used in simulations. A
48 modelling approach will be conducted to discuss and discriminate the triggering of the landslide.

49 Several studies have proposed that deposit geometry could provide indications of landslide
50 dynamics (Puga-Bernabéu et al., 2017; Crosta et al., 2018; Collins and Reid, 2020). For example,
51 runout distance may relate to landslide dynamics and can be used to determine the processes
52 involved (Crosta et al., 2018) as it may depend on the effective friction during sliding (Staron and

53 Lajeunesse, 2009). In this study, the runout distance will be used to discuss the dynamic of the
54 Tahiti landslide. The spreading of landslide material may reflect the processes involved (Puga-
55 Bernabéu et al., 2017; Collins and Reid, 2020). Although morphologic features represent a good set
56 of observable data for past landslides, in the case of large landslides, it may be difficult to
57 discriminate between the specific mobility of the material and the normal spreading of a large
58 volume of material when using morphological data. In addition, large landslide dynamics are
59 difficult to understand due to a lack of accurate data, as deformation and erosion may have affected
60 their morphological characteristics following landslide occurrence (Menendez et al., 2008).
61 Moreover, the rarity of these events reduces the chances of direct observation and thus hinders
62 understanding; dedicated studies are primarily based on past events (Puga-Bernabéu et al., 2017).
63 Large landslides have been observed and studied since several decades, first in aerial cases, for
64 example for the Saidmarreh landslide in Iran (Harrison and Falcon, 1937 ; Roberts and Evans,
65 2013) or the Blackhawk landslide in California (Shreve, 1968; Johnson, 1978), then under
66 submarine conditions. Large landslide morphology is often recorded only in part due to the
67 difficulty of access, as most cases are located underwater (Abril and Periáñez, 2017), are covered by
68 more recent sediments or lava flows, or are located on other planets like Mars (Quantin et al.,
69 2004). In the present case, the landslide deposits are located under water and the scar is now full of
70 lava. Furthermore, landslide scar and mass transport deposit could represent the consequences of
71 several events triggering the development of complex morphologies (Johnson and Campbell, 2017).
72 Given these observational difficulties, it is not always easy to discriminate morphologies generated
73 by a single landslide from those generated by multiple events. Based on a modelling approach, the
74 study aim is to discriminate between scenarios that advocate for a single events from those that
75 propose that multiple events take place in the case of the Tahiti landslide deposits.

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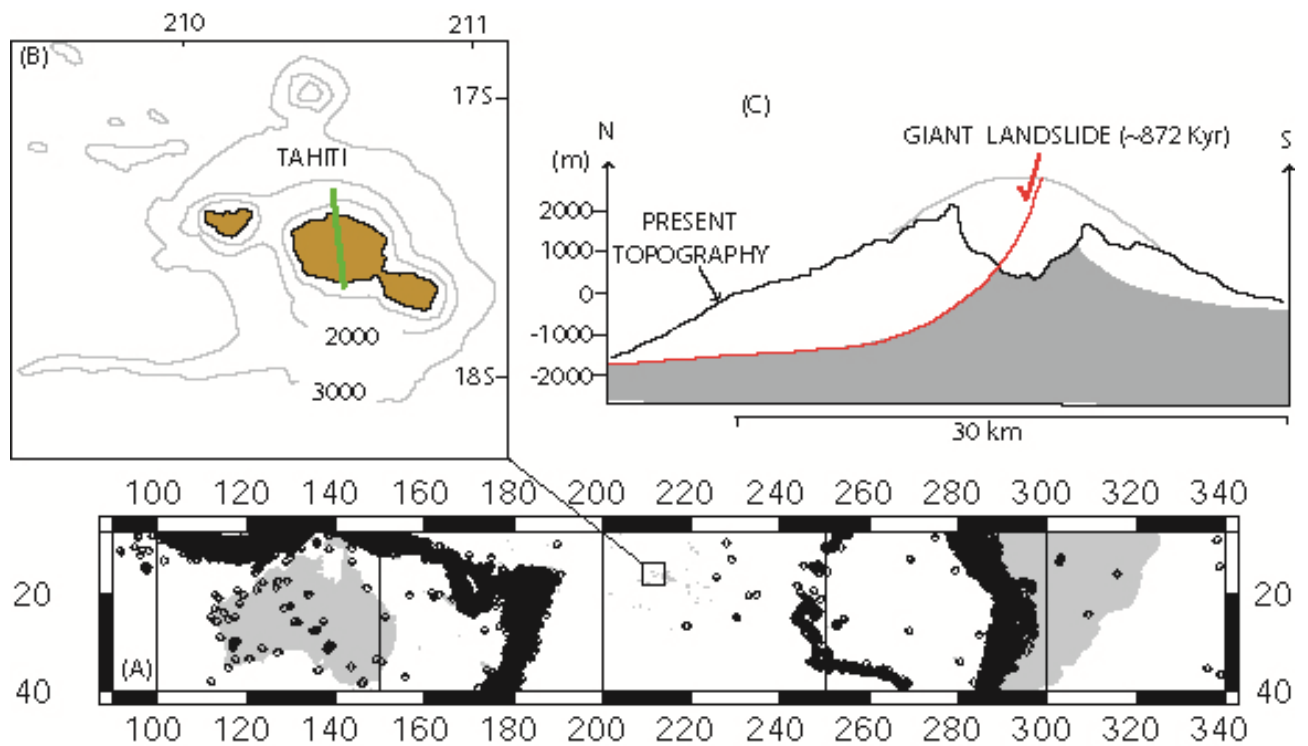
77 Field studies, geochronological results, and bathymetric data suggest that a large mass
78 displacement took place in Tahiti 872 ± 10 kyr ago (Clouard et al., 2001; Hildenbrand et al, 2004,

79 2006). However, the cause of the original slope destabilisation and the resulting landslide dynamics
80 have never been precisely constrained. First, it is not clear whether this landslide scar and
81 associated deposits are the consequence of a single large event, a few moderate events, or many
82 small events. Second, the velocity of the Tahitian slide has never been quantified. Field
83 investigations and seismic observations have constrained the geometry of past landslides but have
84 not focused on quantitatively describing their dynamics, while modelling allows the quantitative
85 assessment of slide velocity along the path. Landslide speed has often been estimated for smaller
86 landslides (Rodriguez et al., 2017; Salmanidou et al., 2018), but data and simulations are less
87 common for large landslides. Third, the influence of submarine versus aerial propagation on the
88 velocity of large landslides is unclear, as the former results in increased drag force and decreased
89 frictional force due to the incorporation of water into the slide; interactions between these factors
90 can generate diverse dynamics.

91 Numerical experiments could provide a complementary approach to previous analyses of the
92 Tahitian landslide scar and related deposits, potentially allowing the validation of a specific
93 scenario and providing quantitative insights. Therefore, this study aimed to (i) discriminate between
94 three hypotheses for this landslide's dynamics (single large event, few moderate events, or many
95 small events) and (ii) estimate the maximum slide velocity.

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98
 99 *Fig. 1: General context for the large Tahitian landslide. (A) Location of Tahiti (17.70° N,*
 100 *149.40° E), circles represent the seismicity from 1997 to 2007 with a magnitude $M > 4$*
 101 *(seismological data from the USGS Earthquake Database). (B) Tahiti-Nui Island and location of*
 102 *cross-section, (C) Schematic cross-section of the giant landslide.*

104 2. Geomorphic patterns and geological structure

105 The study area is located in the southern Pacific Ocean (Fig. 1A). Several submarine
 106 landslides occurred in French Polynesia (Clouard and Bonneville, 2004), whose archipelago
 107 includes the volcanic island of Tahiti. Regionally, the Society volcanic chain consists of numerous
 108 volcanic islands and seamounts are aligned from northwest to southeast, with ages increasing to the
 109 northwest (Patriat et al., 2002). Tahiti is composed of two coalescent volcanoes, Tahiti-Nui to the
 110 northwest and Tahiti-Iti to the southeast (Clouard et al., 2001) with steep slopes. These two
 111 volcanoes are connected by an isthmus. Tahiti-Nui (Fig. 1B) constitutes the youngest extinct
 112 volcano of the Society volcanic chain and is located near the Society hotspot (Duncan and
 113 McDougall, 1976).

114 Lateral collapses are visible on the northern and southern slopes of the main shield volcano
115 (Clouard et al., 2001). The southern collapse occurred between 650 and 850 kyr (Clouard et al.,
116 2001), while the northern collapse occurred around 0.872 ± 0.01 Ma ago (Hildenbrand et al., 2004).
117 These landslides are debris avalanche along listric faults (Clouard and Bonneville ; Hildenbrand et
118 al., 2008). The southern landslide deposits extend up to 60 km away from the island's shore
119 (Clouard and Bonneville, 2004). The maximum runout distance of the northern landslide front
120 reached ~70–80 km, while the runout distance of the centre of mass was ~40 km (Fig. 2). The
121 northern landslide involved volcanic rock several hundreds of metres thick, with a maximum
122 deposit width of ~80–90 km. The volume of the scars are smaller than the volume of the deposits,
123 respectively ~300-460 km³ and 800-1150 km³ (Clouard et al., 2001 ; Clouard and Bonneville, 2004
124 ; Hildenbrand et al., 2008). The Tahitian landslides may have been triggered by a paleoclimatic
125 change (Quidelleur et al., 2008) or by the repeated intrusions of magma inside the main rift zone
126 (Clouard and Bonneville, 2004).

127 Following the landslide, a second shield developed in the northern depression and dykes
128 were generated on the depression's rims (Hildenbrand et al., 2004). Lava and breccias flowed into
129 these depressions, filling them progressively (Hildenbrand et al., 2008). The present slope of the
130 volcanic edifice ranges from 8–12°, similar to conditions before the large collapse (Hildenbrand et
131 al., 2004 and 2006). The slope is steeper than in Hawaii, probably due to the viscosity of the lava,
132 which is higher for those in Tahiti (Craig, 2003). Rocks from the volcanic system are basalts
133 transitional to tholeiites, alkali basalts, basanites, picrites and evolved lavas (Duncan et al., 1994).

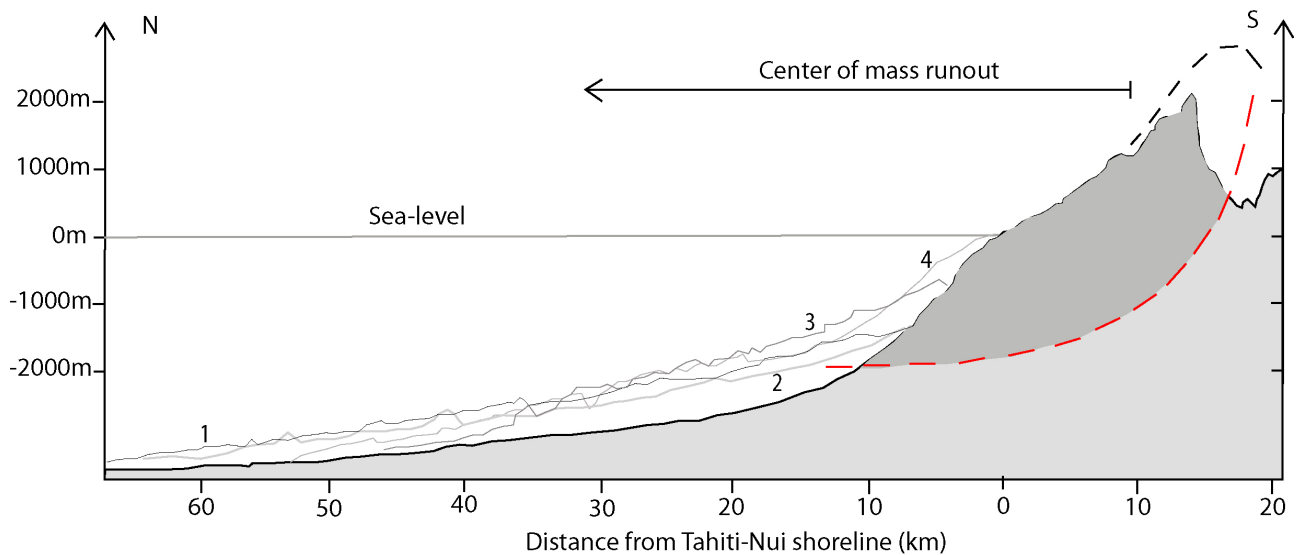
134 Erosion process are still active on the island. The initial formation of valleys take place
135 rapidly and erosion is faster here than in Hawaii (Craig, 2003). In Tahiti, the recession of the fluvial
136 knickpoints was initiated by a sea level drop of around 135 m, 20 ky ago (Ye et al., 2013). Littoral
137 cliffs have been shaped by marine erosion 7kyr ago (Ye et al., 2013). Climatic conditions are
138 characterized by heavy rainfall (ORSTOM, 1993). The rainwater percolates down through the
139 substrate (Menard, 1986) covered in lava and scoria. Groundwater seeps downward in the upper

140 slopes via shallow hollows, macropores and flow paths (Craig, 2003). Saturated subsurface flow
141 cause an erosional break in the soil where a channel will begin (Craig, 2003).

142 This volcano is located in an area where no significant seismicity has been recorded due to
143 its intraplate location (Fig. 1A) and where tectonic deformation is negligible. Subsidence has been
144 measured as 0.25 mm/yr (Bard et al., 1996) and 0.37 mm/yr (Fadil et al., 2009; Thomas et al.,
145 2012). These conditions favour reef formation. There have been no dyke intrusions or volcanic
146 eruptions recorded during the past thousands years, nor any uplift of the volcanic edifice due to
147 magma reservoir activity.

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150

151 *Fig. 2: Longitudinal profiles along the large Tahitian landslide, from which the runout of the centre*
152 *of mass was estimated as 40 ± 5 km (modified from Hildenbrand et al., 2008).*

153

154 **3. Modelling approach**

155 The mobility of the landslide was modelled in order to study various processes explaining
156 the significant runout distance L and to assess the different event scenarios. The model analysed the
157 propagation conditions for three cases: (i) completely subaerial, (ii) completely subaqueous, and
158 (iii) subaerial initiation and subsequent subaqueous propagation.

159 The role of landslide volume on mobility has been investigated by several studies (Goren
 160 and Aharonov, 2007; Brunetti et al., 2009; De Blasio, 2011b), which showed that runout is
 161 proportional to material volume and that there is an apparent decrease of basal friction ($\mu \sim H/L$,
 162 where H is the height) in relation to increased rock volume. Nevertheless, even if complex thermo-
 163 poro-elastic mechanisms (lubrication by high pressure at the base, the presence of evaporites or ice,
 164 local melting, ground vibration, trapped air, etc.) are used to explain this phenomena, other studies
 165 have shown that slope path geometry and unconsolidated mass spreading play a significant role in
 166 describing mobility (Staron and Lajeunesse, 2009). In other words, mobility is controlled primarily
 167 by sliding processes in relation to classical friction laws and spreading processes in the case of large
 168 volume landslide as in other cases (Staron and Lajeunesse, 2009; Salmanidou et al., 2018).

169 Subaerial sliding along a slope θ was modelled using the weight W of the sliding block and
 170 the frictional force R at the base of the sliding block. To simulate the effects of a subaqueous
 171 environment, hydrostatic uplift due to buoyancy and the drag force F_D due to seawater were also
 172 incorporated (Fig. 3). The latter was based on a classical fluid mechanics approach (Hürlimann et
 173 al., 2000):

$$174 \quad F_D = 0.5K \rho_f v^2 A \quad (1)$$

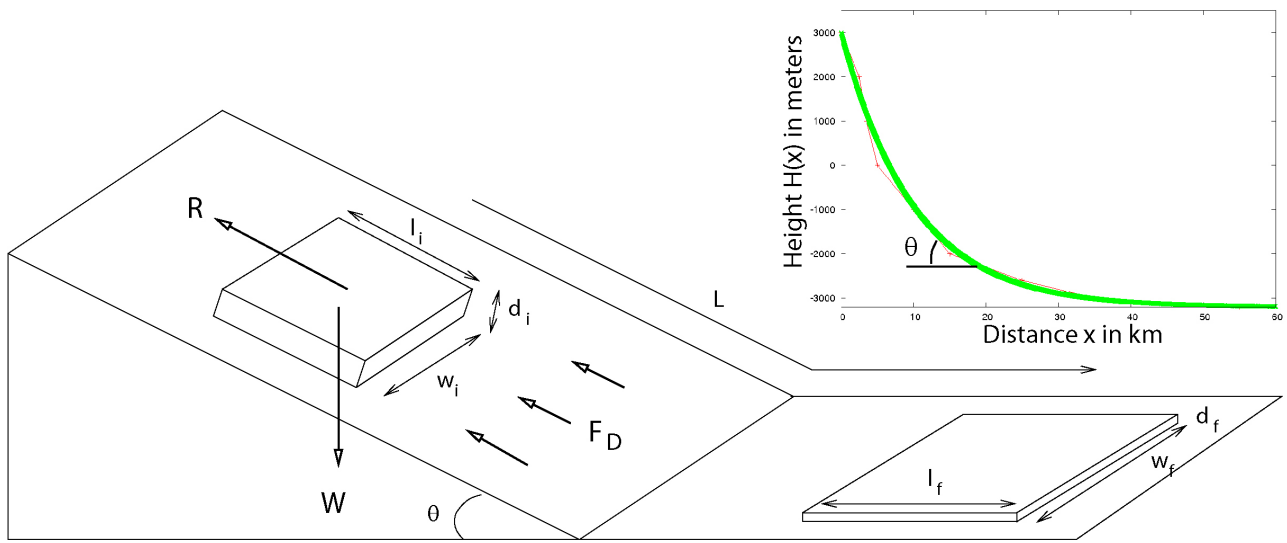
175 where K is a dimensionless shape coefficient for drag and skin friction, ρ_f is the density of the fluid
 176 (i.e. water density if subaqueous or air density if subaerial), v is the sliding block velocity and A is
 177 the surface of the sliding block in contact with the fluid opposing the movement. The value of K
 178 ranges from 0–2 (Hürlimann et al., 2000).

179 The equation of motion (Hürlimann et al., 2000) was defined as:

$$180 \quad \frac{dv}{dt} = [g(\rho - \rho_f)(\sin \theta - \mu \cos \theta) / \rho] - [0.5 (K / l) (\rho_f / \rho) v^2] \quad (2)$$

181 where t is time, g is gravity, θ is the slope, μ is the coefficient of friction, l is the length of the body,
 182 and ρ is the density of the sliding block.

183



185

186

187 *Fig. 3: Geometrical parameters and forces applied to the landslide model along with the simulated*
 188 *path geometry, based on the equation $H(x) = H_0 \exp(x/l_0)$, where $H_0 = 6.2$ km and $l_0 = 10$ km.*

189

190

191 In the model, the transition from subaerial to subaqueous sliding was immediate without loss
 192 of energy. For simplicity, the slope path was represented by an exponential function, $H(x) =$
 193 $H_0 \exp(x/l_0)$, where $H(x)$ is the height of the terrain and x is distance along the path (Fig. 3), H_0 is the
 194 initial height, and l_0 is a length that allows to the curvature to fit the slide path (De Blasio, 2011a).
 195 These geometric parameters were set as $H_0 = 6.2$ km and $l_0 = 10$ km and allow to fit the real slope
 196 (Fig. 3). The density of the sliding block (composed of volcanic rocks) was considered constant at
 197 2800 kg/m^3 (Hürlimann et al., 2000). The density of the fluid ρ_f was set as 1000 kg/m^3 for
 198 subaqueous sliding and 0 kg/m^3 for subaerial sliding (Hürlimann et al., 2000). The geometry of the
 199 sliding body was characterised by an initial geometry (initial width w_i , initial length l_i , and initial
 200 thickness d_i), a final geometry (final width w_f , final length l_f , and final thickness d_f), and a sliding
 201 length L (also referred to as runout) (Fig. 3). These parameters were estimated from
 202 geomorphologic and bathymetric data (Hildenbrand et al., 2004, 2006). The runout considered here

203 was that of the centre of mass, not the maximum runout of the slide front.

204 The large landslide's initial conditions were estimated as width $w_i = 25$ km, length $l_i = 20$
205 km, and thickness $d_i = 2000$ m, the final geometry was estimated as $w_f = 90$ km, $l_f = 70$ km, and $d_f =$
206 160 m, and the sliding distance of the centre of mass was estimated as $L = 40 \pm 5$ km. The volume
207 of the sliding block was identical at the initial and final states. The temporal evolution of the
208 geometrical parameters from the initial state to the final state was assumed to be linear. Due to the
209 short duration of the events considered ($300 \text{ s} < t < 850 \text{ s}$), as defined by the simulations, this
210 assumption did not significantly influence the maximum velocity and runout distance. Even if the
211 real value was uncertain, the influence of the geometrical parameters was less significant (in the
212 range of values tested) with respect to the mobility than the friction μ and the hydrodynamic factor
213 K .

214 The parameters selected for the mobility analysis were the ratio between the hydrodynamic
215 coefficient and the length of the sliding block (K/l) and the friction μ . The first parameter is known
216 to strongly influence sliding dynamics under subaqueous conditions (Hürlimann et al., 2000; De
217 Blasio, 2011a). Assuming a parallelepipedic shape of length l , width w , and thickness d , $K/l = C_D/l +$
218 $C_s/d + 2C_s/w$, where C_D is the drag coefficient and C_s is the skin coefficient. The drag coefficient
219 can be calculated using $C_D = 1.95 - 0.77d/w$ (De Blasio, 2011a). The skin coefficient ranges from
220 $0.002 < C_s < 0.006$ (De Blasio, 2011a). Under these assumptions, the parameter K/l ranges from
221 $5 \cdot 10^{-5} \text{ m}^{-1}$ to $5 \cdot 10^{-4} \text{ m}^{-1}$, similar to those used by Hürlimann et al. (2000) ($K/l = 2 \cdot 10^{-4} \text{ m}^{-1}$ and $K/l =$
222 $2 \cdot 10^{-3} \text{ m}^{-1}$). Under subaerial conditions, the coefficient $K/l = 0 \text{ m}^{-1}$.

223 The value of friction μ can range from 0.6–0.85 (Bayerlee's law) for dry subaerial conditions
224 of intact rocks and has been estimated as low as 0.0025–0.08 for subaqueous conditions in specific
225 cases (Hürlimann et al., 2000; De Blasio, 2011a) such as hydroplaning or the presence of
226 evaporites. In this study, the basal friction μ was set at an intermediate value of 0.2–0.3 under
227 subaqueous conditions instead of 0.6 under subaerial conditions. The former value corresponded to
228 an angle of friction of $\phi = 11\text{--}17^\circ$ in accordance with the value considered appropriate for highly

229 weathered and fragmented volcanic rock (Rodriguez-Losada et al., 2009).

230 Alternatively, a viscous rheology could be considered to simulate the complex behaviour of
231 the landslide using the Bingham law, in which the shear stress:

$$232 \quad \tau = \eta \, dv/dz + \tau_0, \text{ if } \tau > \tau_0$$

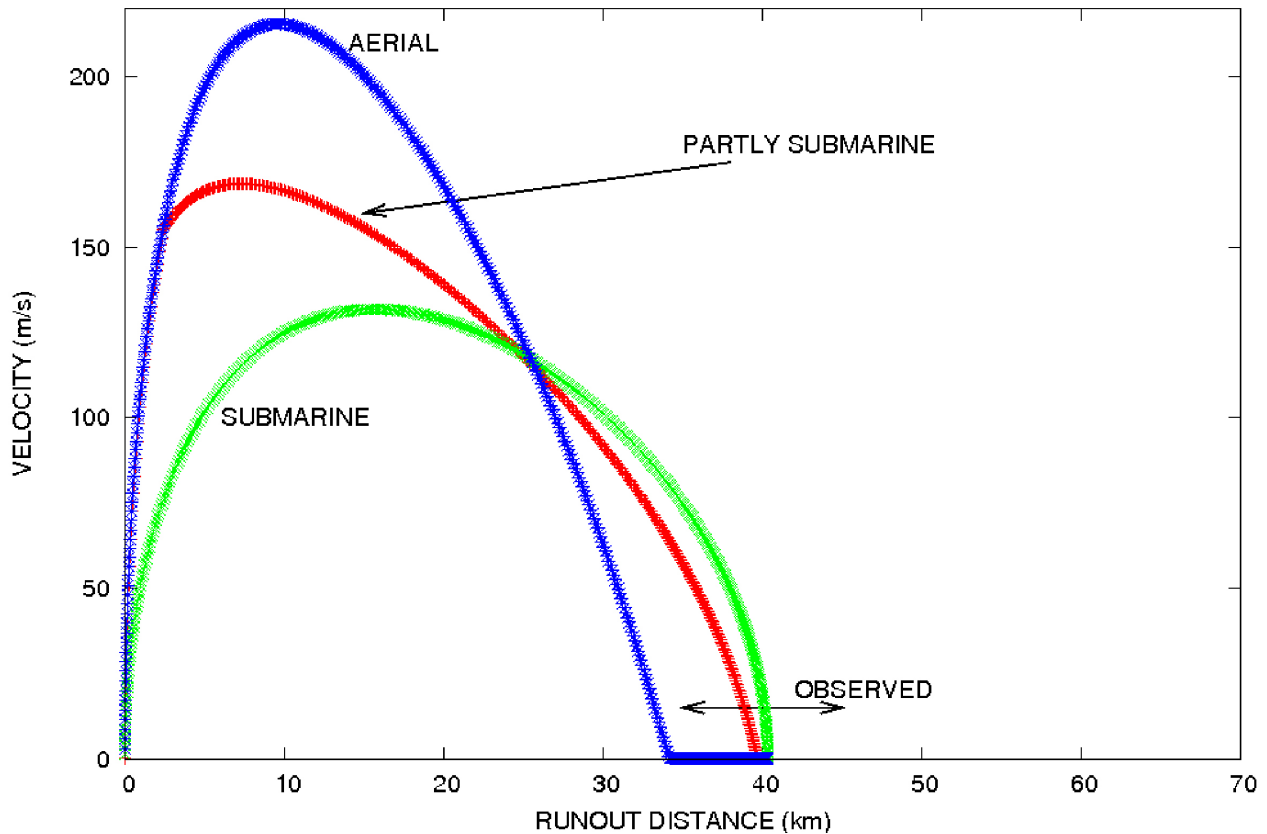
$$233 \quad \text{and } \tau = 0, \text{ if } \tau < \tau_0,$$

234 where η is the viscosity, τ_0 is the yield stress, v is the velocity of the landslide, and z is the vertical
235 distance. For a thickness h and a depth-averaged velocity v , the shear rate can be estimated by dv/dz
236 $\approx 2v/h$ (Pelletier et al., 2008). The resulting shear stress thus becomes $\tau = 2\eta v/h + \tau_0$. Therefore, an
237 increase in viscosity will increase the difficulty of landslide propagation, as expected. Under this
238 assumption, landslide dynamics can be modelled by:

$$239 \quad \frac{dv}{dt} = [g(\rho - \rho_f) \sin \theta / \rho] - [2\eta v / (\rho h^2) + \tau_0 / (\rho h)] - [0.5 (K / l) (\rho_f / \rho) v^2] \quad (3)$$

240 where the first term on the right side is due to the weight, the second term is due to the viscous
241 behaviour of the landslide, and the third term is due to water drag.

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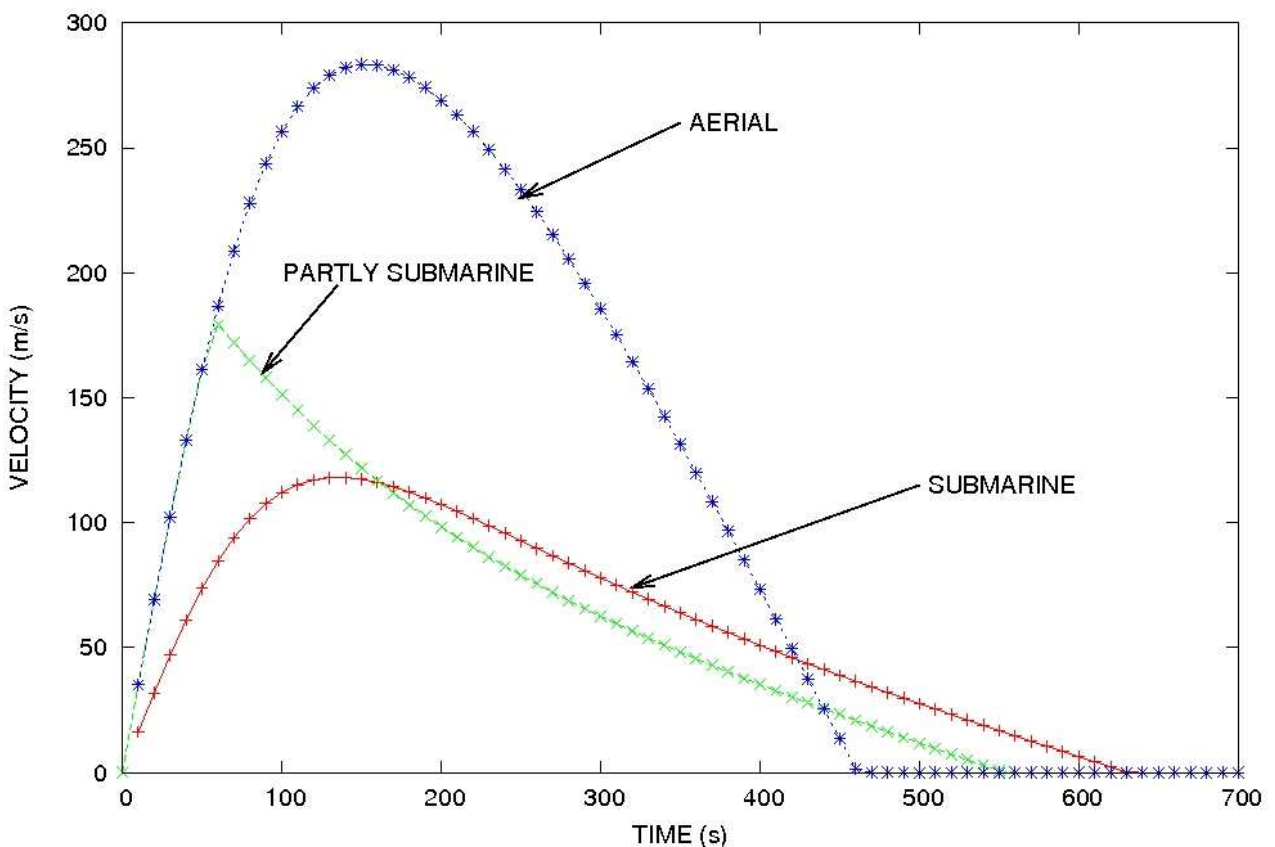
244 *Fig. 4: Influence of three simulated propagation conditions: aerial, submarine, and partly*
 245 *submarine (5 km aerial, then submarine). $K/l = C_D/l + C_S/d + 2C_S/w$ for all cases. $\mu = 0.27$ ($\phi =$*
 246 *15.5°) under submarine conditions and $\mu = 0.6$ under aerial conditions. $w_i = 25$ km, $l_i = 20$ km, and*
 247 *$d_i = 2$ km. Observed runout of the landslide's centre of mass is assumed to be 40 ± 5 km.*

248

249 **4. Results**

250 The landslide propagated mostly under subaqueous conditions although the slide may have
 251 initiated under subaerial conditions. This study's model assumed that the slide material entered into
 252 the sea after 5 km and compared the effects of subaqueous and subaerial propagation. The
 253 landslide's submarine and partly submarine mobility was of the same order of magnitude (Fig. 4).
 254 The runout distance and slide time could be higher under subaqueous conditions than under
 255 subaerial conditions, assuming a reduction of basal friction under submarine conditions from 0.6 to
 256 0.3 (Fig. 4). The subaerial propagation of the landslide showed a higher maximum velocity than

257 submarine propagation, despite the reduction of effective basal friction simulated in the latter (Fig.
 258 4). When the effect of hydroplaning was not considered (μ was constant), aerial propagation was
 259 longer than submarine propagation due to the increased hydrodynamic drag force. The maximum
 260 aerial velocity was ~ 215 m/s, but only 170 m/s for a partly submarine landslide and 125 m/s for a
 261 submarine landslide (Fig. 4). Entering the water generally resulted in decreased velocity (Fig. 5).
 262 However, if the volume of material was significant ($L = 20$ km, $V = 1000$ km³), the velocity of the
 263 landslide was able to continue increasing after entering the water (although at a smaller rate) for ~ 3
 264 km due to the inertia of the sliding material (Fig. 6). The slide time was higher under subaqueous
 265 conditions than under subaerial conditions (Fig. 5).
 266



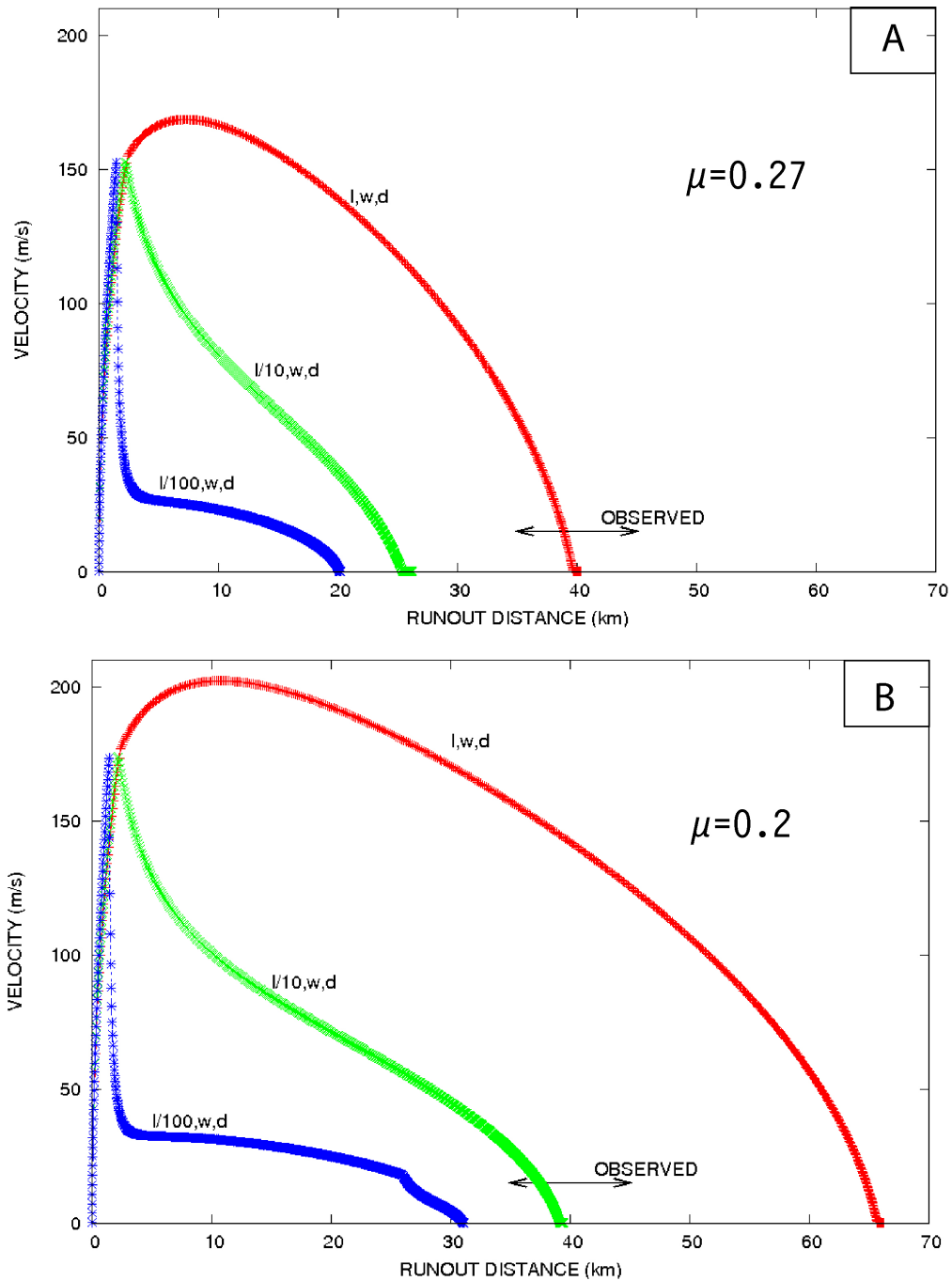
267
 268 *Fig. 5: Landslide velocity over time under three different propagation conditions: aerial,*
 269 *submarine, and partly submarine (5 km aerial, then submarine). The drag friction was set equal to*
 270 *$K/l = 0.0004$. $\mu = 0.2$ ($\phi = 11.3^\circ$) under submarine conditions and $\mu = 0.6$ under aerial conditions.*

271 $w_i = 25 \text{ km}$, $l_i = 20 \text{ km}$, and $d_i = 2 \text{ km}$.

272 A landslide with a large initial volume V of 1000 km^3 (volume $V = w \times l \times d = 25 \text{ km} \times 20$
273 $\text{km} \times 2 \text{ km}$) allowed the runout distance of $\sim 40 \text{ km}$ observed offshore of Tahiti with a basal friction
274 $\mu = 0.27$ (Fig. 6A), but not with a basal friction $\mu = 0.2$ (Fig 6B). Smaller landslides with a length
275 $l/10$ or $l/100$ had a runout L smaller than a bigger landslide with an initial length l under partly
276 submarine conditions. For a repetition of $n = 10$ small landslides with a length $l/10$, a width w , and a
277 thickness d ($w \times l/10 \times d = V/10$), the total deposit volume ($V_{total} = 10 \times V/10$) was equivalent to a
278 unique landslide with a volume V . Ten landslides with a volume $V/10$ had an expected runout of
279 $\sim 25 \text{ km}$ when the basal friction $\mu = 0.27$, below the observed runout distance observed (Fig. 6A).
280 However, when the basal friction $\mu = 0.2$, the runout distance of landslides with volume $V/10$ fit the
281 observed runout data (Fig. 6B). For a volume of each landslide of $V/100$ (i.e. the length of each
282 landslide became $\sim l/100$, but the initial width w as well as the thickness d was the same as for
283 landslides with volumes V and $V/10$), the calculated runout distance was $\sim 20 \text{ km}$ for $\mu = 0.27$ and
284 30 km for $\mu = 0.20$, below the observed runout distance (Fig 6). For $l/100$, only a very small basal
285 friction $\mu \ll 0.2$ was able to fit the observed runout of the centre of mass. The number of landslides
286 with a volume $V/100$ necessary to obtain the observed volume of deposits was 100. In all
287 simulations, the maximum velocity of the landslides ranged from 180–200 m/s.

288 In the Tahitian case considered here, the final deposit's centre of mass was located $\sim 40 \text{ km}$
289 from its initial position and the front of the slide propagated $\sim 80 \text{ km}$. An effective basal friction μ of
290 $0.2\text{--}0.3$ under submarine conditions allowed the simulation to fit the observed runout distance.
291 Reduced basal friction permitted the simulation of hydroplaning effects. Conversely, increasing
292 aerial propagation increased the maximum velocity of the slide but slightly reduced the runout
293 distance of the centre of mass.

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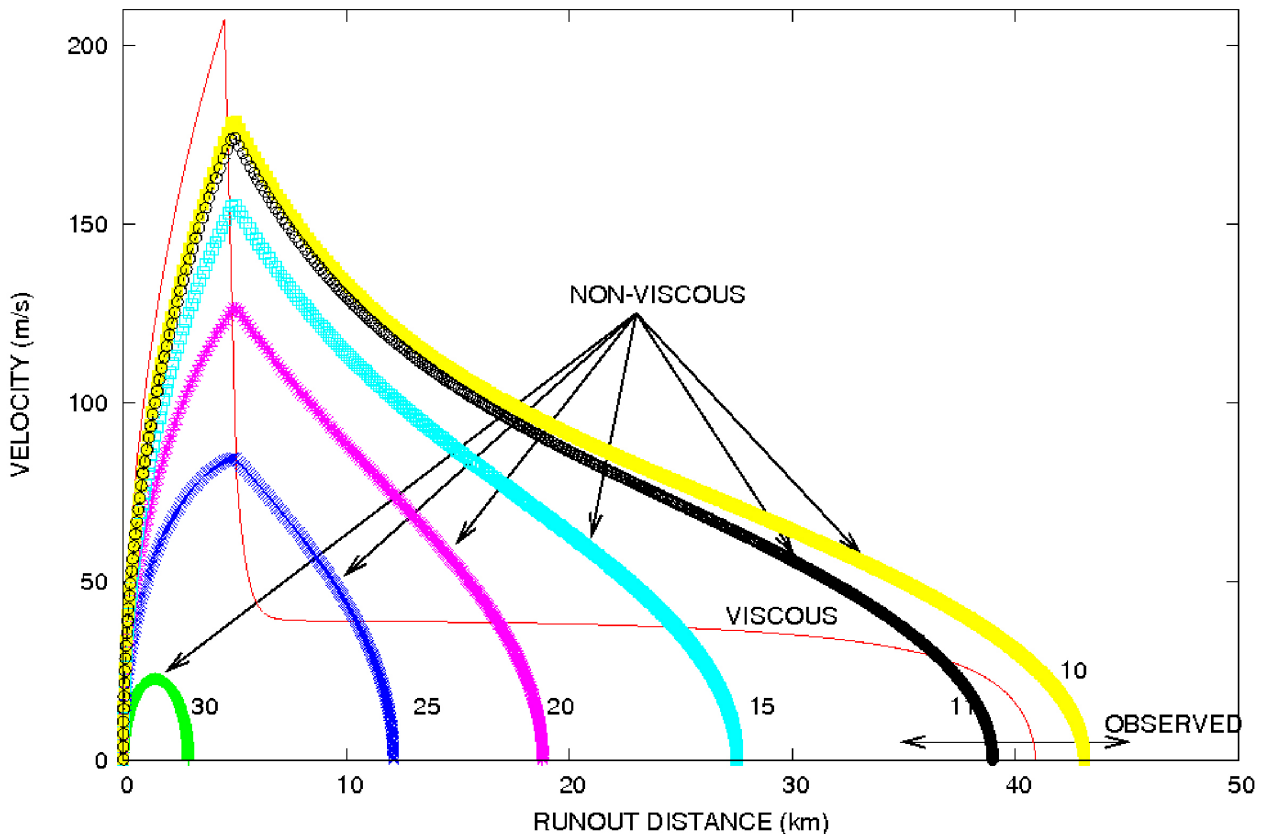
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Fig. 6: Effect of landslide volume on landslide dynamics for three volumes and two friction values: (A) $\mu=0.27$ ($\phi=15.5^\circ$) under submarine conditions and (B) $\mu=0.2$ ($\phi=11.3^\circ$) under submarine conditions. In both cases, propagation initiated under aerial condition but transitioned to submarine conditions after 5 km. The water effective drag coefficient was defined as $K/l=C_D/l+C_S/d+2C_S/w$, where $w_i=25$ km, $d_i=2$ km, and $l_i=20$ km, 2 km, or 0.2 km. The observed runout of the landslide centre of mass was 40 ± 5 km. Under aerial conditions, the basal friction $\mu=0.6$.

303 The mechanical behaviour of the highly fragmented sliding material was accurately
 304 simulated by low effective friction and allowed explanation of its significant mobility. Alternative
 305 modelling experiments including more complex rock behaviour may also explain the dynamics of
 306 this landslide (Fig. 7). For example, viscous behaviour could also explain the observed runout of the
 307 centre of mass (Fig. 7). There is an abrupt decrease of the velocity from the aerial part to the
 308 submarine part. In this case, the submarine velocity was < 50 m/s for almost all slide paths (Fig. 7)
 309 and smaller than for non-viscous conditions; this is near the peak velocity obtained for other
 310 submarine landslides with a smaller size (L'heureux et al., 2013; Rodriguez et al., 2017; Abril and
 311 Periañez, 2017; Salmanidou et al., 2018; Sawyer et al., 2019).



312
 313 *Fig. 7: Modelled landslide velocity vs. runout distance ($\eta = 5$ MPa.s, $\tau_0 = 0.5$ Mpa) for viscous and*
 314 *various non-viscous rheologies for different angles of internal friction ($\phi = 30^\circ, 25^\circ, 20^\circ, 15^\circ, 11^\circ,$*
 315 *and 10°). In all cases, the landslide propagates aurally for the first 5 km, then transitions to*
 316 *submarine conditions. $K/l = C_D/l + C_S/d + 2C_S/w$. $w_i = 25$ km, $l_i = 2$ km, $d_i = 2$ km. The observed*
 317 *runout of the landslide centre of mass is 40 ± 5 km.*

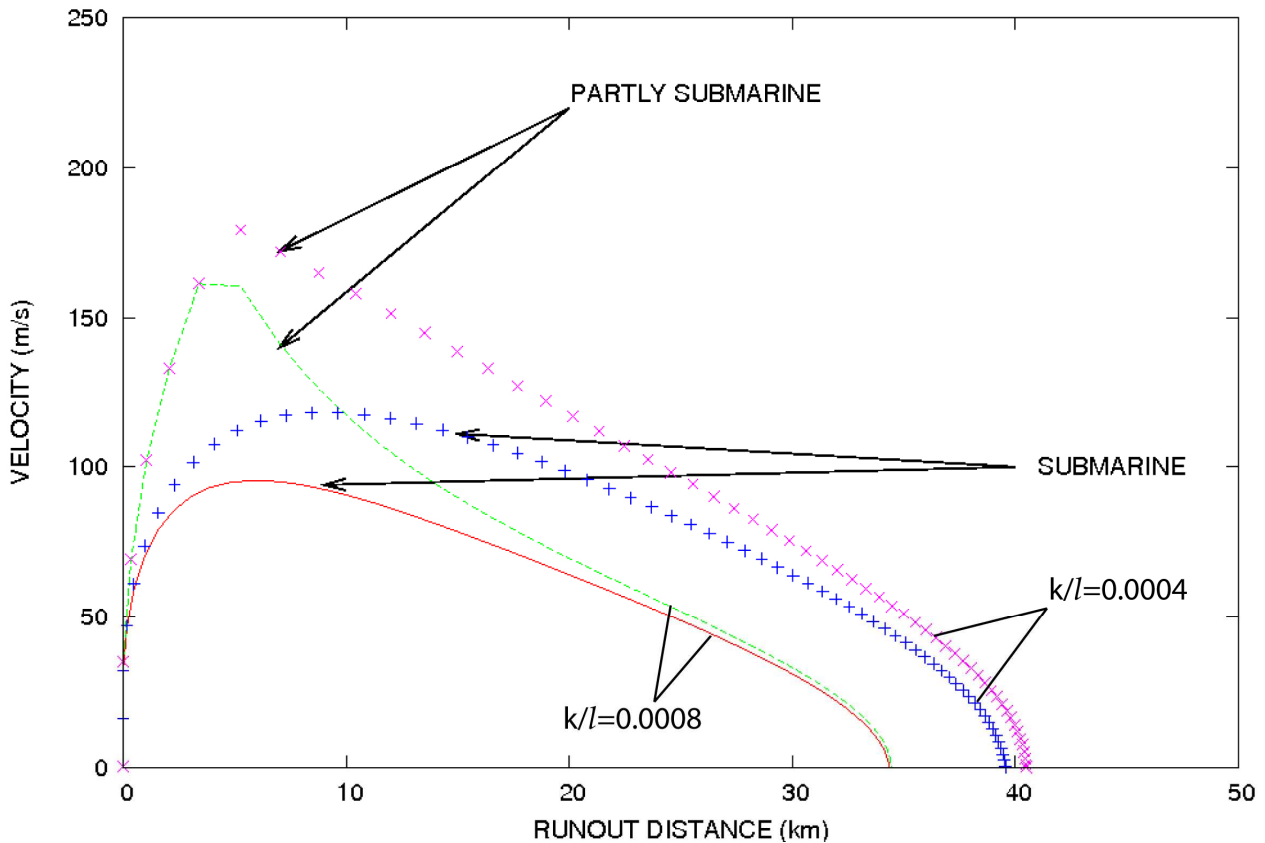
318

319 **5. Discussion**

320 Previous studies have suggested that landslide propagation under submarine conditions
321 could result in longer runout distances and the occurrence of hydroplaning (Locat and Lee, 2002;
322 De Blasio, 2011b; Hürlimann et al 2000). However, in several case where no hydroplaning
323 occurred, submarine propagation reduced the runout distance (Gargani et al., 2014). Other
324 lubrication effects could also explain decreases in basal friction under aerial (Johnson, 1978) or
325 submarine conditions. The spreading of a large volume of material could partially explain the long
326 apparent runout of the landslide's front (Johnson and Campbell, 2017).

327 This study's modelling showed that slide velocity decreased (Fig. 5) or reduced its
328 acceleration (Fig. 4) after entering the water, as the drag force generated by water is more efficient
329 than that of air; the former could be significant at high velocities. Previous studies suggested that
330 subaerial mass acceleration can be much greater than submarine mass acceleration yet (Kafle et al.,
331 2016). Furthermore, the presence of water could generate a counterintuitive effect by increasing the
332 runout distance (Fig. 4). Such significant runout under subaqueous conditions has been explained
333 by hydroplaning (De Blasio, 2011a). The competition between drag and friction force on one hand,
334 and gravity and buoyancy on the other, could explain the high runout under subaqueous conditions
335 seen in this study. Under such conditions, the propagation along the path was slower but lasted
336 longer (Fig. 5), as shown by previous studies (Hürlimann et al., 2000; De Blasio, 2011b). When the
337 volume and length of a landslide are large, the inertia of the sliding block could allow a reduced
338 effect of the drag force when entering the water, such that the sliding block speed is not reduced
339 abruptly when the propagation becomes submarine (Fig. 6).

340



341

342 *Fig. 8: Influence of the effective drag coefficient k/l on landslide dynamics for submarine and partly*
 343 *submarine propagation. For aerial propagation, the effective drag coefficient $k/l=0$. $\mu=0.2$ under*
 344 *submarine condition and $\mu=0.6$ under aerial condition.*

345

346 One of this study's main questions relates to whether the Tahitian landslide deposits
 347 considered here were formed through a single slide event or multiple events. If the former, the
 348 runout distance of ~40 km could be easily achieved with a friction $\mu = 0.27$ under submarine
 349 conditions (Fig. 6A). However, under this condition, other landslide geometries did not fit the
 350 observed runout distance. The simulated runout was lower for smaller landslide geometries, as an
 351 increasing length l decreases the ratio K/l . The drag coefficient is significantly reduced when the
 352 length of the sliding block is increased. Equation (2) shows that larger landslides are capable of
 353 propagating longer distances than smaller ones (Fig. 6). If all other parameters are equal, an
 354 increase in the effective drag coefficient K/l will decrease propagation distance (Fig. 8). A decrease

355 in K/l by a factor of two can increase the runout of the centre of mass by a factor of ~ 1.2 . In this
356 model, the width of the landslide had only a small influence on the runout because the skin
357 coefficient C_S was small in comparison with the drag coefficient C_D .

358 Nevertheless, it may be possible that the deposits were formed by a limited number of
359 landslide events ($n < 10$). Indeed, the observed runout distance can also be explained by the
360 occurrence of ten separate landslides with a length of ~ 2 km using a basal friction $\mu = 0.2$ under
361 submarine conditions (Fig. 6B). However, this multi-slide scenario ($n > 10$) is less realistic, because
362 it implies that the basal friction became very low ($\mu \ll 0.2$, $\phi < 10^\circ$) in order to fit the observed
363 runout. Although the occurrence of a single event was difficult to prove, the modelling showed that
364 a reduced number of events (< 10) was necessary to reproduce the observed runout distance.
365 Therefore, these results support the proposal that a single event or a reduced number of events
366 generated the large debris apron morphology. The ability of few events to form moderate to large
367 landslide morphology (i.e. scars and mass transport deposits) has been advocated in several specific
368 cases using more accurate geomorphological data and observations (Cauchon-Voyer et al., 2011 ;
369 Puga-Bernabéu et al., 2017 ; Collins and Reid, 2020). For example, some studies of the Oso
370 landslide have advocated for two large, distinct failure events separated by a few minutes, whereas
371 others studies suggested that the vast majority of landslide activity occurred within the initial
372 minute (Collins and Reid, 2020). In comparison, the Gloria Knolls landslide consisted of three
373 events (Puga-Bernabéu et al., 2017).

374 Extremely small basal friction values are not necessary to explain the large runout of the
375 Tahitian landslide's centre of mass, which primarily propagated under submarine conditions.
376 Values of $0.2 < \mu < 0.3$ corresponded to an angle of internal friction from $11\text{--}17^\circ$, as expected for
377 very weathered basalt or pyroclastic rocks (Rodriguez-Losada et al., 2009). To reach these values,
378 the landslide may have caused significant fragmentation of the material. The numerous fractures
379 and long polygenic history of the volcanic edifice probably contributed to the presence of weak
380 material (Hildenbrand et al. 2004). The humid climate of Tahiti could also explain the highly

381 weathered and altered nature of the rocks, which contributed to the presence of weak material from
382 a mechanical perspective.

383
384 The estimated slide peak velocity ranged from 125–250 m/s, within the range of other large
385 landslide peak velocities observed or calculated in volcanic contexts (Voight et al., 1983;
386 Hürlimann et al., 2000) but higher than that of smaller submarine landslides (20–50 m/s)
387 (L’heureux et al., 2013; Rodriguez et al., 2017; Abril and Periañez, 2017; Salmanidou et al., 2018;
388 Sawyer et al., 2019). The significant mass displacement may have triggered an isostatic rebound
389 (Gargani, 2004, 2010). A tsunami should also have occurred after the landslide as has been
390 estimated for other comparable events (Bohannon and Gardner, 2004; Rodriguez et al., 2017).

391 392 **6. Conclusion**

393 The velocity of the Tahitian landslide exceeded 125 m/s, allowing the fragmentation of
394 highly weathered rocks. The effective basal friction for submarine propagation was relatively low
395 ($0.2 < \mu < 0.3$), enabling significant mobility for the centre of mass ($\sim 40 \text{ km}^3$) and suggesting that
396 hydroplaning occurred under submarine conditions. The long runout length of the centre of mass
397 also suggested that the landslide deposit was the consequence of a single event or a reduced number
398 of events ($n < 10$). The initial length of the landslide varied between $l/10 < l_i < l$, where $l = 20 \text{ km}$ for
399 a total volume $V \sim 1000 \text{ km}^3$. The significant mobility of the material and the slide dynamics
400 constrained the geometry of the mass transport deposit.

401
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403
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409

410

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