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1 **Permafrost extension modeling in a rock slope since the Last Glacial**
2 **Maximum: application to the large Séchilienne landslide (French**
3 **Alps).**

4

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10

11 **Abstract**

12

13 Recent dating performed on large landslides in the Alps reveal that the initiation of instability
14 did not immediately follow deglaciation but occurred several thousand years after ice down-
15 wastage in the valleys. This result indicates that debuttressing is not the immediate cause of
16 landslide initiation. The period of slope destabilization appears to coincide with the wetter and
17 warmer Holocene Climatic Optimum, indicating a climatic cause of landslide triggering,
18 although the role of seismic activity cannot be ruled out. A phenomenon which may partly
19 explain the delay between valley deglaciation and gravitational instability is the temporal
20 persistence of thick permafrost layers developed in the Alps since the Last Glacial Maximum
21 (LGM). This hypothesis was tested through 2D thermal numerical modeling of the large
22 Séchilienne landslide (Romanche valley, French Alps) using plausible input parameter values.
23 Simulation results suggest that permafrost vanished in the Séchilienne slope at 10 to 11 ka,
24 3,000 to 4,000 years following the total ice down-wastage of the Romanche valley at 14.3 ka.
25 Permafrost persistence could have contributed to the failure delay by temporally
26 strengthening the slope. Numerical simulations also show that the permafrost depth expansion
27 approximately fits the thickness of ground affected by gravitational destabilization, as
28 deduced from geophysical investigations. These results further suggest that permafrost
29 development, associated with an ice segregation mechanism, damaged the rock slope and
30 influenced the resulting landslide geometry.

31

32 **Keywords:** Permafrost modeling; TTOP model; Last Glacial Maximum; Large landslide;
33 French Alps; Séchilienne

34

35 **1. Introduction**

36

37 The triggering of large gravitational movements in mountainous areas, following the last
38 Pleistocene glacial retreat, has been a debated question for many years (for a recent review,
39 see Sanchez et al., 2010). Glacial slope steepening and subsequent debutressing (lateral stress
40 release resulting from ice melting) have been frequently proposed as major causes of rock-
41 slope failures (Cruden and Hu, 1993; Augustinus, 1995; Cossart et al., 2008), although the
42 role of other phenomena like cleft-water pressure, seismic activity and climatic changes have
43 also been invoked (Ballantyne, 2002; Hormes et al., 2008; Ivy-Ochs et al., 2009; Le Roux et
44 al., 2009). Local factors like relief and favorable fracture patterns also play a role in
45 predisposing slopes to fail (Korup et al., 2007). In the last ten years, dating methods, mainly
46 the ^{14}C and cosmic ray exposure (CRE) techniques, have provided chronological constraints
47 on the failure time for major large alpine landslides (e.g., Bigot-Cormier et al., 2005;
48 Deplazes et al., 2007; Prager et al., 2009; Ivy-Ochs et al., 2009; Le Roux et al., 2009). In the
49 Alps, surface exposure age measurements in the above studies show that large landslides
50 initiated around the early to mid-Holocene: Fernpass (Austria, 4.1 ka), Flims (Switzerland,
51 8.9 ka), Kandertal (Switzerland, 9.6 ka), Köfels (Austria, 9.8 ka), La Clapière (France, 10.3
52 ka), Séchilienne (France, 6.4 ka) and Valtellina (Italy, 7.4 ka). The time interval following
53 total melting of ice in valleys during which the slope endures the new state of stress before the
54 initiation of failure (pre-failure endurance; Ballantyne, 2002) was estimated at least between
55 2,000 and 5,400 years (Le Roux et al., 2009), implying that these events are not an immediate
56 consequence of debutressing. Moreover, they often coincided with the Climatic Optimum
57 period, which is characterized in the Alps by increased mean temperatures of 1–2°C (Davis et
58 al., 2003), forest cover density (de Beaulieu, 1977) and lake levels due to heavy annual
59 precipitation (Magny, 2004, 2007). These data suggest that climatic changes play a major role

60 in landslide triggering (Ivy-Ochs et al., 2009; Le Roux et al., 2009). Recently Sanchez al.
61 (2010) applied the CRE technique on glacial, tectonic and gravitational surfaces in the SW
62 Alps. The resulting dates of 11 to 8 ka clearly show that the main tectonic activity postdates
63 deglaciation and corresponds to gravity destabilization. This interpretation is a probable
64 consequence of the post-glacial rebound and the enhanced pore water pressure, the inferred
65 cause of widespread slope fracturing. This tectonic phase was followed by rock weathering
66 during the Climatic Optimum. The development of large gravitational mass movements could
67 be related to the combined effects of intense tectonic activity and climatic change from cold
68 and dry (Pleistocene) to warm and wetter (Holocene) phases. Although the validity of this
69 scenario to the whole Alpine range has still to be documented, these results illustrate the
70 complexity of the interaction among tectonic, climatic and gravitational processes. The
71 question of the pre-failure endurance in the Canadian Rockies was addressed by Cruden and
72 Hu (1993) who proposed an exhaustion model, which assumes that the overall probability of
73 failure occurring within a given area diminishes exponentially with time elapsed since a
74 deglaciation. As outlined by Ballantyne (2002), however, this model is difficult to calibrate
75 and apply, particularly in zones characterized by gentle slopes.

76 Another factor that could contribute to explain pre-failure endurance is the persistence of
77 permafrost in the rock mass. Indeed, a thick permafrost layer developed in the Alps during the
78 Early Holocene and probably reached more than 150 m deep, as suggested by numerical
79 modeling (Wegmann et al., 1998) and a permafrost/glacier evolution study (Guglielmin et al.,
80 2001). The first effect of permafrost is to stabilize slopes by increasing mechanical properties.
81 Comparing the deformation and strength properties of frozen and unfrozen crystalline rocks,
82 Krivonogova (2009) has shown that the presence of ice increases the Young modulus and
83 cohesion by a factor of about 2, while the friction angle remains similar. Permafrost
84 development contributes to slope reinforcement, thus stabilizing surfaces. With significant

85 variations of temperature over the last 21,000 years, permafrost thickness has varied with time
86 disappearing in low-elevation slopes, similar to the one affected by the Séchilienne landslide
87 in the French Alps, whose crown is at about 1100 m a.s.l. Ice disappearance has probably
88 created favorable conditions for low-elevation slope failure, as suggested by the increasing
89 evidence of destabilization at present (see Gruber and Haeberli, 2007 for a review). The
90 sensitivity of permafrost to anthropomorphic climate change and its influence on natural
91 hazards are now recognized, and numerical modeling is increasingly used for investigating
92 the effect of climate variability and topography on permafrost temperature and extension
93 (Riseborough et al., 2008; Noetzli and Gruber, 2009).

94 On the other hand, the presence of permafrost lasting millennia allowed the accumulation of
95 ice-rich layers at the top and bottom of a frozen layer (Matsuoka et al., 1998), through the ice
96 segregation mechanism. That occurs when liquid water migrates through a porous medium
97 towards freezing surfaces, resulting from temperature gradient-induced suction in freezing or
98 frozen ground (Murton et al., 2006). Laboratory experiments simulating rock freezing
99 produce fractures containing segregated ice layers near the permafrost table (Murton et al.,
100 2001). These results demonstrate that ice segregation is an important rock degradation
101 process, as suggested by other authors (see Matsuoka and Murton, 2008 for a review). With
102 permafrost boundary variations in rock slopes over long time-scales, ice segregation may
103 have acted as a contributory factor producing rock mass fractures, preferentially parallel to
104 slope, to a depth of a few tens of meters or more (Matsuoka et al., 1998). Modeling the
105 thermal evolution of the Konkordia ridge (Switzerland) since the end of the Little Ice Age,
106 Wegmann et al. (1998) demonstrated permafrost penetration into the first decameters of rock
107 as a consequence of temperate glacier retreat. Considering climatic variations in northern
108 Fennoscandia and using the TTOP model (Temperature at the Top Of Permafrost;
109 Riseborough et al., 2008) with constant n -factors, Kukkonen and Safanda (2001) showed that

110 the permafrost thickness experienced considerable variations during the Holocene, with a
111 maximum permafrost penetration between 100 and 250 m for low porosity rocks and
112 temperate glacier conditions. In conclusion, they stressed that vegetation and snow cover
113 changes during the Holocene should be taken into account in the model.

114 The present paper investigates the potential role of permafrost extension and persistence in
115 development of a large landslide during the period between deglaciation and failure initiation.
116 The 2D thermal response of the Séchilienne slope (Western Alps, France) during the last
117 21,000 years was computed using the TTOP model for two scenarios: cold and temperate
118 glaciers. The influence of long-term freeze-thaw action on slope fracturing was estimated by
119 comparing the computed deeper permafrost extension to the present-day deconsolidated zone
120 imaged by P-wave seismic tomography (Le Roux et al., 2011). The modeling has also
121 permitted evaluation of the persistence effect of permafrost on slope evolution, in addition to
122 the other involved processes like glacial debuttressing and climatic change.

123

124 **2. Geological and kinematic contexts**

125

126 The lower Romanche valley is located in the Western Alps (southeast of France), about 20 km
127 SE of Grenoble City (Fig. 1). It borders the southern part of the Belledonne massif (external
128 crystalline massifs), which is divided into two main lithological domains, the external one to
129 the west and the internal one to the east (Fig. 1) (Guillot et al., 2009). These two geological
130 units are separated by a major Late Paleozoic near-vertical fault so-called Belledonne Middle
131 Fault (BMF in Fig. 1). During the Quaternary, the Romanche Valley was subjected to many
132 cycles of glaciation and deglaciation including the Last Glacial Maximum (LGM) around 21
133 ka (Clark et al., 2009) when the Romanche and Isère valleys were covered with ice to an
134 elevation of 1200 m a.s.l (Montjuvent and Winistörfer, 1980) (Fig. 1). The relief of the lower

135 Romanche valley shows a strong glacial imprint (van der Beek and Bourgeon, 2008; Le Roux
136 et al., 2010; Delunel et al., 2010) such as steep slopes dipping 35° to 40° , overdeepened
137 troughs and glacial deposits. These characteristics suggest that the thermal regime of the
138 glacier was temperate, although the majority of glaciers are polythermal (Owen et al., 2009).
139 Moreover, the right bank of the Romanche valley is overlooked by a glacial plateau (Mont
140 Sec plateau) at an elevation higher than 1100 m a.s.l (Fig. 1). This plateau is locally overlain
141 with relict peat bogs (Muller et al., 2007) that developed quickly in a cold and wet
142 environment after the disappearance of ice. The steep slopes in the external domain of the
143 Belledonne massif, which mainly consists of micaschists unconformably covered with
144 Mesozoic sediments and Quaternary deposits, is affected by several active or dormant large
145 gravitational movements (Fig. 1).

146 Among these movements, the best known and most active is the Séchilienne landslide (Fig.
147 1), whose 40 m high head scarp affects the southern edge of the Mont Sec glacial plateau
148 (Fig. 2a). Below the head scarp, a moderately sloping depletion zone between 950 and 1100
149 m a.s.l exhibits a series of large depressions and salient blocks (Fig. 2a,c), while the lower
150 part of the landslide, between 450 and 950 m a.s.l, shows steep convex slopes ($> 40^\circ$, Fig. 2c)
151 and is interpreted as an accumulation zone (Vengeon, 1998). The Séchilienne slope is cut by
152 three main sets of near-vertical open fractures oriented N20, N70 and N110 to N120 (Fig. 2b).
153 This structural framework results in linear scarps and troughs filled by rock debris and topsoil
154 (Fig. 2a), which delineate rock blocks displaying downslope motion. The N20 fractures are
155 near-parallel to the BMF and their orientation fits the main foliation plane measured in the
156 micaschists over the slope. The N70 set corresponds to a major regional fracture set
157 evidenced on both sides of the BMF, in the micaschists and the amphibolites, and is probably
158 inherited from the regional tectonics (Le Roux et al., 2010). In the accumulation zone, these
159 wide open fractures delineate near vertical slabs locally toppling downhill and have been

160 progressively filled with coarse scree deposits. Finally, the N110–120 fracture set, which is
161 also interpreted as tectonically inherited (Le Roux et al., 2010), is dominant in the depletion
162 zone (Fig. 2). Additional structural data were provided by the north–south oriented
163 exploration gallery (G in Fig. 2a). The gallery description (Vengeon, 1998) shows a
164 succession of pluri-decametric compact blocks separated by meter-to-decameter crushed
165 zones filled with soft clay materials, trending N50 to N70 with 80° northwestward dip. These
166 undeformed blocks are affected by few near-vertical N0 and N90 fractures and by a dense set
167 of N75-oriented short fractures dipping 40–50°S, near-parallel to the slope. These fractures
168 are also visible on the slope surface (Fig. 3) and were recently observed in the first 100 m of a
169 150 m deep borehole drilled in the accumulation zone (labeled B in Fig. 2a; Bièvre et al.,
170 2012).

171 The cross-section of Fig. 2c summarizes the main structural features evidenced at the surface
172 and at depth along a survey gallery. At the hectometer to kilometer scale, the main set of
173 fractures, near-vertical and trending N70, cuts the whole mass and appears as V-shaped
174 troughs filled with soil deposits at the surface and as crushed zones in the gallery. This major
175 fracture family, which favors the toppling mechanism in the accumulation zone, is cut by
176 numerous pluri-metric fractures dipping near-parallel to the slope. These two sets of fractures
177 result in a stepped geometry that probably controls the downward movement (Fig. 2c).
178 Fracturing parallel to the slope has been commonly observed in sites previously covered by
179 glaciers, and the origin of these fractures has usually been associated with the stress release
180 resulting from deglacial unloading (e.g. Ballantyne and Stone, 2004; Cossart et al., 2008).
181 Eberhardt et al. (2004) documented such fractures in the gneissic slope of the Randa valley
182 where a major rockslide occurred in 1991. Modeling the glacial rebound process at this site,
183 they showed that these tensile fractures parallel to topography could be induced up to a depth
184 of 200 m. However, as mentioned before, the permafrost expansion with time could also have

185 played a role in fracturing the rock mass, preferentially parallel to the slope (Matsuoka and
186 Murton, 2008).

187 The Séchilienne landslide has been the subject of multiple investigation campaigns for fifteen
188 years (for a recent review, see Le Roux et al. 2011). The combination of the
189 geomorphological and geological analysis, displacement rate values and deep geophysical
190 investigation allowed delineation of the area covered by the landslide (Fig. 2a). The volume
191 affected by the landslide was estimated from deep seismic profiles, bracketed between 48×10^6
192 m^3 and $63 \times 10^6 \text{ m}^3$ by P-waves velocity (V_p) thresholds at 3000 and 3500 m s^{-1} , respectively
193 (Le Roux et al. 2011). The two landslide limits are shown in the cross-section (Fig. 2c).

194 Cosmic ray exposure (CRE) dating in the area showed that the glacier retreat occurred at 16.6
195 ± 0.6 $^{10}\text{Be ka}$ at 1120 m a.s.l (Le Roux et al., 2009). By transposing to the Romanche valley
196 the chronological constraints from the large alpine valley of Tinée (Bigot-Cormier et al,
197 2005), located 130 km to the South, Le Roux et al., (2009) proposed that the total down-
198 wastage of the Romanche valley at 400 m a.s.l occurred at 13.3 ± 0.1 ka. More closely,
199 Delunel (2010) calculated a vertical glacier ablation rate between 0.30 and 0.37 m year^{-1}
200 (mean value of 0.335 m year^{-1}) in the valley of Vénéon, filled with a 670 m thick glacier.
201 Applying these ablation rate values to the 760 m high Romanche glacier, extending from the
202 bottom of the valley (380 m) to the Mont Sec plateau (1140 m), provide an earlier total down-
203 wastage estimate of the Romanche valley about 14.3 ± 0.3 ka. Therefore, the Séchilienne
204 slope head scarp failure initiation, dated 6.4 ± 1.4 $^{10}\text{Be ka}$ (Le Roux et al., 2009) during the
205 warmer and wetter Holocene Climatic Optimum period, occurred at least 6,200 years after
206 glacial retreat. Slope destabilization does not, then, appear to have been an immediate
207 consequence of the Romanche valley debuitressing event, the observed delay at least partly
208 related to the permafrost persistence. This hypothesis is examined in the following sections.

209

210 3. Ground thermal evolution model

211

212 As the climatic and surface conditions prevailing in the study area over the last 21,000 years
213 are poorly known, the simple TTOP model (Smith and Riseborough, 1996) was chosen and
214 coupled with the heat transfer equation in a 2D finite element code for simulating the
215 permafrost temperature variations in the Séchilienne slope. Following Riseborough et al.
216 (2008), the temperature profile is divided in five distinct layers, from top to bottom (Fig. 4):
217 the lower atmosphere, the surface layer (from the base of the lower atmosphere to the Earth
218 surface), the active layer (from the Earth surface to the permafrost table), the permafrost body
219 and the deep ground. The corresponding boundary temperatures are the mean annual air
220 temperature (T_{maa}), the mean annual ground surface temperature (T_{mag}), the mean annual
221 temperature at the top of the permafrost body (T_{top}) and the mean annual temperature at the
222 bottom of the permafrost body (T_{bot}). The differences between T_{maa} and T_{mag} on the one hand,
223 and T_{mag} and T_{top} on the other hand are called surface offset and thermal offset, respectively
224 (Smith and Riseborough, 1996). The TTOP model combines the processes occurring in the
225 surface layer and in the active layer to estimate the temperature T_{top} . The surface offset (Fig.
226 4) depends on the isolating and albedo effects of different ground conditions (vegetation,
227 snow cover, forest floor, mineral soils, etc.) and could be estimated by calculation of the
228 surface energy balance. In the TTOP model, these complex processes within the surface layer
229 are simplified and accounted for by two factors, i.e. the freezing and thawing factors (n_{F} and
230 n_{T} , respectively). The n_{T} factor incorporates all microclimatic effects (radiation, convection,
231 evapotranspiration, etc.) due to vegetation, while n_{F} is mainly controlled by the influence of
232 snow cover (Smith and Riseborough, 1996). The TTOP model is detailed in Appendix 1.

233 In and below the permafrost, a simple heat transfer model (Williams and Smith, 1989) is used
234 to relate T_{top} to T_{bot} , considering the geothermal flux and the latent heat phase changes.

235 Fluctuation of permafrost thickness, however, changes the thermal regime by consuming or
236 releasing large amounts of latent heat during freeze/thaw processes, respectively. Following
237 Mottaghy and Rath (2006), the latent heat phase change is accounted for by introducing an
238 effective heat capacity c_e in the heat transfer equation (see Appendix 2).

239

240 **4. Air temperature reconstruction**

241

242 The thermal response simulation of the S echilienne slope requires the mean annual air
243 temperature curve (T_{maa}), from the Late Glacial Maximum (21 ka, Clark et al., 2009) to the
244 present day, as well as the seasonal temperature fluctuations (ATA) that are deduced from the
245 mean annual temperatures of the coldest and warmest months (T_{mco} and T_{mwa} , respectively).

246 The three T_{maa} , T_{mco} and T_{mwa} curves over the time period (21 to 0 ka) were reconstructed for
247 the S echilienne site by compiling curves of several origins and spanning different time
248 intervals (Fig. 5). The following data were considered: (1) recent temporal climatic series
249 characterizing the studied area from 1960 to the present (InfoClimat, 2011); (2) the
250 Greenland ice core records providing the T_{maa} curve evolution from 40 to 0 ka (Alley, 2000);
251 (3) quantitative pollen climate reconstructions for Central Western Europe giving thermal
252 anomalies of T_{maa} , T_{mco} , and T_{mwa} with respect to the present-day temperature since 12.0 ka
253 (Davis et al., 2003); and (4) T_{mwa} deduced from chironomids and pollen data from 14.0 to 10.8
254 ka (Ilyashuk et al., 2009). For the present period, the temperature series measured at the
255 Grenoble Saint Geoirs meteorological station between 1960 and 2010 was used to produce the
256 T_{maa} , T_{mwa} and T_{mco} curves. This station, located 50 km NW of the S echilienne slope at an
257 elevation of 384 m a.s.l, required a lapse rate correction to account for the elevation difference
258 to the top of the S echilienne slope (1140 m a.s.l). Thus, we applied the altitudinal temperature
259 decrease of $5.7^{\circ}\text{C km}^{-1}$ proposed by Ortu et al. (2008). The obtained temperatures were taken

260 as present day reference values at 1140 m a.s.l. The temperature variations determined for
261 Central Western Europe between 10.8 to 0 ka (Davis et al., 2003) were applied to compute the
262 T_{maa} , T_{mco} , and T_{mwa} temperature curves at the top of the S echilienne slope (1140 m a.s.l.)
263 during the same period (Fig. 5). The T_{maa} curve was extended to 21 ka by using the Greenland
264 ice core records (Alley, 2000), while the T_{mwa} curve was constrained from 10.8 to 14.0 ka by
265 using chironomids (Ilyashuk et al., 2009). Determining *ATA* from the T_{mwa} and T_{maa} curves
266 allowed the T_{mco} curve to be computed during the same period of time (Fig. 5). Finally, the
267 only missing data (T_{mco} and T_{mwa} curves between 21 and 14 ka) were estimated by assuming a
268 linear relationship between T_{maa} and *ATA* values. These composite temperature curves (T_{mwa} ,
269 T_{mco} , and T_{maa}) were used as input data in the thermal modeling of the S echilienne slope from
270 21 to 0 ka. Despite a substantial uncertainty, they provide a plausible estimate of the
271 temperature variation at the study site during that period of time. In Fig. 5, four thermal
272 periods were distinguished (labeled A to D) from the temperature curve fluctuations: a cold
273 period A from 21 ka (Last Glacial Maximum) to 14.7 ka with a mean T_{maa} around -8°C ; a
274 warmer period B from 14.7 to 13.0 ka with a T_{maa} between -4.0°C to $+5.5^{\circ}\text{C}$; a short colder
275 period C until 11.6 ka with a mean T_{maa} around -10°C ; and a warmer period D from 11.6 to 0
276 ka (Holocene) with a T_{maa} between $+1.5^{\circ}\text{C}$ to $+7.5^{\circ}\text{C}$.

277

278 **5. Numerical model definition**

279

280 *5.1. Thermal scenarios*

281 Because of various interpretations of thermal and surface conditions prevailing in the
282 S echilienne region over the last 21,000 years, four models were defined, implying two glacier
283 thermal regimes and two ground thermal sets of parameters. First, as glaciers are often
284 polythermal (Owen et al., 2009), two glacier thermal regimes were considered: a cold glacier

285 (regime C) and a temperate glacier (regime T), with a base temperature equal to T_{top} and 0°C ,
286 respectively. Second, thermal ground parameters were usually set constant in numerical
287 modeling (e.g. Kukkonen and Safanda, 2001), although the vegetation and snow cover
288 conditions controlling the n -factors significantly varied during the succession of different
289 thermal periods (Fig. 5). Two ground condition scenarios were then considered. In the first
290 one, the n -factors were kept constant with time and the ranges of values ($0.40 \leq n_T \leq 1.30$;
291 $0.20 \leq n_F \leq 1.00$) were derived from the works of Lunardini (1978), Jorgenson and Kreig
292 (1988) and Juliusen and Humlum (2007), with the same default values ($n_T = 0.70$ and $n_F =$
293 0.50) as those used by Smith and Riseborough (1996). In the second scenario, the n -factors
294 were defined for each of the four thermal periods (Table 1). During the cold periods A and C
295 (Fig. 5), the n_F factor was computed using the relation proposed by Riseborough and Smith
296 (1998), assuming a snow cover between 0.2 and 1.0 m and an average T_{maa} value of -8°C and
297 -10°C for periods A and C, respectively. The corresponding n_T factor values were derived
298 from Juliusen and Humlum (2007) for barren ground surfaces. Under warm periods, both n_F
299 and n_T are controlled by vegetation and the default values and ranges were defined from
300 Lunardini (1978), Jorgenson and Kreig (1988) and Juliusen and Humlum (2007). Finally,
301 glacier-covered areas were characterized by $n_F = 1$ and the n_T values for cold periods (Table
302 1).

303 Considering the two glacier thermal regimes (T and C) and the two thermal ground conditions
304 (1 and 2, with constant and time-variable n -factors, respectively), we numerically simulated
305 four models (labeled 1C, 1T, 2C and 2T). The initial conditions prevailing at 21.0 ka were a
306 surface temperature $T_{\text{maa}} = -10^{\circ}\text{C}$ (Fig. 5) and a glacier level at 1200 m a.s.l (Montjuvent and
307 Winistörfer, 1980). Exploiting the CRE data of Le Roux et al. (2009) and Delunel (2010),
308 glacier ablation rates of 0.014 and 0.335 m yr^{-1} were applied before and after 16.6 ka,
309 respectively.

310

311 *5.2. Methods and the geometrical model*

312 The 2D thermal evolution in the Séchilienne slope was numerically simulated during the last
313 21,000 years, by implementing the permafrost model of Fig. 4 in the 2D finite-element
314 Comsol software (<http://www.comsol.com>). First, the slope geometry before destabilization
315 was approximately reconstructed along the N–S cross-section of Fig. 2c by balancing the
316 depletion and accumulation surfaces (Fig. 6). It resulted in a simple 40° slope cut by the Mont
317 Sec plateau and the valley at 1140 and 380 m a.s.l, respectively. This model was laterally and
318 vertically extended to reduce the boundary effects and was gridded (Fig. 7), using a mesh
319 composed of 1758 triangular elements with a size between 65 and 135 m. The temperature
320 evolution in the slope over the last 21,000 years was simulated with a time step of 4.2 years.
321 A null horizontal heat flux was applied at both vertical boundaries of the model, while a
322 constant vertical upward heat flow of 65 mW m^{-2} , similar to the present-day flux (Lucazeau
323 and Vasseur, 1989; Goy et al., 1996), was imposed at the bottom of the model.

324

325 *5.3. Parameters*

326 The model was supposed homogeneous and the parameter values used for the thermal
327 simulation are given in Table 2 (default values and ranges of variation). Porosity and bulk
328 density values (ϕ and ρ_d) were determined by previous laboratory tests performed on
329 micaschist samples (Le Roux et al., 2011), with default values of 3.7% and 2730 kg m^{-3} ,
330 respectively. Porosity was bracketed between 0.9% and 5.3%. Although micaschists are
331 thermally anisotropic, a unique thermal conductivity value k_T of $2.5 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$ was
332 considered in thawed rock (Goy et al., 1996), while a specific heat capacity value of $c = 800 \text{ J}$
333 $\text{kg}^{-1} \text{ }^\circ\text{C}^{-1}$ was taken from the literature (Stacey and Davis, 2008). A ground conductivity ratio
334 $r_k = k_T/k_F$ between 0.25 and 1 was considered, with a default value fixed at 0.5 (Smith and

335 Riseborough, 1996). A freezing interval parameter θ of 0.3°C was taken between the solidus
336 and liquidus temperatures (Wegmann et al., 1998). Finally, as ground temperatures also
337 depend on the solar radiation (Blackwell et al., 1980), which is controlled by the slope angle
338 and orientation, we also considered a scenario with a temperature correction of $+0.4^{\circ}\text{C}$
339 applied to the south-facing 40° Séchilienne slope (Safanda, 1999; Table 1). The T_{top} values at
340 the ground surface were calculated from the air temperature curves of Fig. 5, using Eq. (1) in
341 Appendix 1. A thermal gradient of $5.7^{\circ}\text{C km}^{-1}$ was considered for elevation corrections (Ortu
342 et al., 2008).

343

344 **6. Modeling results**

345

346 The temperature evolution in the slope was simulated for the four models, considering the
347 default values shown in Table 2. The temperature distributions computed for Model 1T
348 (constant n -factors and temperate glacier) at seven different times (Fig. 8b) are plotted with an
349 interval of 2°C (Fig. 8a), along with the extension of the permafrost (dark blue) and the
350 glacier (light blue). Permafrost depths measured perpendicular to the slope at four points (P1
351 to P4, Fig. 8a) are given in Table 3 for five different times with an accuracy of ± 10 m.
352 Notably, the permafrost limits are roughly parallel to the slope surface. At 21.0 ka (time 1 in
353 Fig. 8b), there was no permafrost since the long-term temperature at the bottom of the
354 temperate glacier was zero. During the cold period A (times 2 and 3, 16.6 and 15.0 ka), the
355 permafrost gradually spread into the upper and middle parts of the ice-free slope, following
356 the glacier lowering. The maximum permafrost thickness reached about 190 m (Table 3).
357 During the warmer period B (time 4, 14.1 ka), the permafrost thinned and just an iced core
358 remained in the upper part of the slope. During the following cooler period C (times 5 and 6,
359 13.0 and 12.0 ka), the permafrost developed into the slope to reach again a maximum depth of

360 190 m. At the beginning of the Holocene (period D, time 7, 10.0 ka), the temperature rose
361 quickly before stabilizing, provoking the quick melt of the permafrost that vanished.

362 Simulations for the other three models 1C, 2C and 2T are shown in Fig. 9 at the same
363 periods/times, and corresponding permafrost depths are given in Table 3. Under cold glacier
364 conditions, the slope was initially frozen to a depth varying between 150 and 350 m for model
365 1C and 100 and 225 m for model 2C (Fig. 9). During the cold period A (21.0 to 15.0 ka), the
366 glacier progressively lowered and the permafrost volume slightly decreased to reach a
367 thickness ranging from 95 to 315 m for model 1C and 45 to 165 m for model 2C. During the
368 warmer period B (14.1 ka), the glacier disappeared from the valley and the permafrost was
369 reduced to a thick core in the upper part of the slope, with a much larger extension for model
370 1C. At 13.0 ka, the cold thermal period C initiated a new growing of the permafrost along the
371 slope, which reached a depth between 120 and 255 m (Model 1C) and from 70 to 170 m
372 (Model 2C) at 12.0 ka. The permafrost disappeared at 10.3 ka for Model 1C and about 1,000
373 years earlier for Model 2C.

374 Finally, the permafrost evolution with time for Model 2T (variable n -factors in Table 1) is
375 similar to that described for Model 1T (Fig. 8), with lower permafrost depths. In Table 3,
376 permafrost depths at P1 to P4 for the four models are compared at the five different times in
377 Table 3, along with the permafrost disappearance age. A striking feature is that the permafrost
378 totally melts in the same time range (11.0 to 10.0 ka) for all simulations. The comparison of
379 the permafrost depths along the slope shows that the maximum extension of permafrost (330
380 m at the top of the slope) was obtained for Model 1C, while the more limited extension was
381 observed for Model 2T (125 m at the same site). For the same glacier conditions, accounting
382 for time-dependent n -factors resulted in less development of the permafrost than the extension
383 computed with constant n -factors. Notably, for cold glacier conditions (Models 1C and 2C),

384 the maximum permafrost depths were reached during the first cold period A, while they were
385 observed during the second cold period (C) under temperate glacier conditions.

386

387 **7. Discussion**

388

389 The maximum depth reached by the permafrost along the slope is plotted in Fig. 10 for the
390 four models. Three models (1T, 2T and 2C) yield relatively similar results while a significant
391 deviation in permafrost depth (330 m) is observed for Model 1C. Although cold glacier
392 conditions cannot be locally excluded, the strong glacial erosion observed in the Western Alps
393 (Owen et al., 2009) is in favor of a temperate regime at Séchilienne. In particular, Model 1C
394 (constant n -factors and cold glacier) is the least plausible among the considered models and
395 has been discarded. The permafrost penetration obtained for Model 1T (105 to 195 m) is
396 compared to the thickness values (100 to 250 m) computed by Kukkonen and Safanda (2001)
397 in northern Fennoscandia, using the TTOP model under the same conditions. In both studies,
398 depth values are of the same order of magnitude. As concluded by Kukkonen and Safanda
399 (2001), depth estimations could be improved by accounting for the changes in snow and
400 vegetation cover. In order to define the most impacting parameter on the modeling results, a
401 study of the sensitivity was performed for models 1C and 1T, through varying the five poorly
402 constrained parameters (ξ , r_k , n_T , n_F and s_c) in the range indicated in Table 2. The results (not
403 shown) indicate that the predominant parameter is n_F , underlining again the need to better
404 precise the n -factor values for modeling. The effect of n -factor fluctuations with climate was
405 investigated in Model 2T and it turned out that the permafrost penetration was about 30% less
406 in this case (70–135 m; Table 3 and Fig. 10). Sensitivity tests were made for the same Model
407 2T, focusing on the n -factor variations in the range shown in Table 2. The maximum observed

408 effect is a permafrost persistence variation of 600 years and a depth fluctuation of 30 m with
409 respect to the default values.

410 The maximum depth reached by the permafrost for the three models (1T, 2T and 2C) is
411 compared (Fig. 10) to the depth affected by the landslide along the same cross-section,
412 considering the two V_p threshold limits (3000 and 3500 m s^{-1}) proposed by Le Roux et al.
413 (2011). The maximum permafrost depths computed along the slope are of the same order of
414 magnitude (100 to 190 m) as the thickness of the damaged zone imaged by the seismic
415 investigation (Le Roux et al., 2011). This comparison suggests that the long-term permafrost
416 front fluctuations during the last 21,000 years could have played a role in mechanically
417 degrading the slope through ice segregation, a mechanism suggested by Wegmann et al.
418 (1998) and Kukkonen and Safanda (2001) in other regions. This hypothesis is supported by
419 the observation of meter-size fractures nearly parallel to the slope, both at the surface and in
420 the first 100 m of borehole B. The common explanation for this fracture pattern is the stress
421 release following deglacial unloading (Balantyne and Stone, 2004; Cossart et al., 2008). The
422 penetration and intensity of fracturing during debuttressing strongly depend upon rock
423 mechanical characteristics (Augustinus, 1995), which then could have been controlled by the
424 permafrost-induced slope weakening.

425 Our results are synthesized in Fig. 11, which shows the chronological constraints on the
426 events that could have affected the Séchilienne slope. From CRE dating, the final total down-
427 wastage of the Romanche valley was estimated at 14 ka (T_g^f), at least 6,200 years before the
428 initiation of Séchilienne head scarp. This delay can be considered as a minimal pre-failure
429 endurance corresponding to the time interval following the disappearance of the glacier
430 during which the slope endures the new state of stress before the initiation of failure. Thermal
431 modeling results suggest that permafrost vanished in the Séchilienne slope between 10 to 11
432 ka (T_p), i.e. at least 2,000 to 3,000 years before the Séchilienne head scarp failure. These

433 results suggest that the permafrost disappearance did not directly cause the failure but its
434 persistence could have delayed the rupture by a few thousand years, by mechanically
435 strengthening the slope. Finally, the head scarp destabilization occurred at 6.4 ka (Tdⁱ), during
436 the warmer and wetter Climatic Optimum period (Magny, 2004; Davis et al. 2003). This
437 suggests that increases in temperature and precipitation during the Middle Holocene
438 significantly contributed to the Séchilienne slope destabilization. Fig. 11 emphasizes that the
439 permafrost expansion and degradation since 21,000 years played a key role in the Séchilienne
440 slope development, in a multi-process phenomenon including glacial debuttressing and
441 Pleistocene to Holocene climate change.

442

443 **8. Conclusions**

444

445 The thermal numerical modeling of the Séchilienne slope during the last 21,000 years showed
446 that permafrost vanished around 10 to 11 ka and therefore persisted at least 3,000 to 4,000
447 years after total ice down-wastage in the Romanche valley. The strengthening effect of ice can
448 only partly explain the 6,200-yr delay measured between glacial retreat and instability
449 initiation of the head scarp, which occurred during the wet and warm Climatic Optimum
450 period. These results support the interpretation of a predominant role of climate on slope
451 destabilization, although the effect of seismic activity cannot be ruled out completely. This
452 study also reveals that, under the most plausible conditions (temperate glacier and time-
453 dependent *n*-factors), the permafrost below the Séchilienne slope since the Last Glacial
454 Maximum (LGM) reached a maximum thickness of 70 to 135 m, which corresponds to the
455 destabilization depth inferred from seismic prospecting. These observations suggest that
456 permafrost expansion weakened the Séchilienne slope and controlled the thickness of ground
457 fractured after glacial unloading.

458 Permafrost development and longevity has turned out to be factors controlling slope stability,
459 in addition to those usually proposed such as glacial debulking, climate changes and active
460 tectonics. In particular, deep permafrost expansion is shown to play a significant role in the
461 development of deep-seated landslides in previously glaciated areas. The effect of permafrost
462 is, however, hard to show from direct field observations, and its importance in comparison
463 with that of the other factors is still difficult to assess. Understanding complex gravitational
464 movements requires further investigation combining CRE dating and thermo-mechanical
465 finite element modeling.

466

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468

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473

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686

687 **Appendix**

688 **Appendix 1:**

689 In the TTOP model the n_T and n_F factors are applied as transfer functions between T_{maa} and
 690 T_{mag} . In the active layer, the thermal offset, which results from the difference in thermal
 691 conductivity values between frozen and thawed grounds, is related to ground thermal
 692 properties and to the ground surface temperature. The effects of active and surface layers are
 693 combined to obtain the following equation (Smith and Riseborough, 1996):

$$694 \quad T_{\text{top}} = \frac{k_T n_T I_{\text{TA}} - k_F n_F I_{\text{FA}}}{Pk^*} \quad \text{with} \quad k^* = \begin{cases} k_F & \text{if } k_T I_{\text{TS}} - k_F I_{\text{FS}} < 0 \\ k_T & \text{if } k_T I_{\text{TS}} - k_F I_{\text{FS}} > 0 \end{cases} \quad (1)$$

695 where k_F and k_T are the thermal conductivity values for the frozen and thawed ground, n_F and
 696 n_T are the freezing and thawing factors, I_{FA} and I_{TA} are the air seasonal freezing and thawing
 697 degree-day indexes, I_{FS} and I_{TS} surface seasonal freezing and thawing-degree days indexes,
 698 and P is the period (365 days) of temperature fluctuations. Air seasonal indexes can be
 699 deduced from the mean annual air temperature curve and the annual temperature amplitude
 700 (Smith and Riseborough, 1996).

701

702 **Appendix 2:**

703 The effective heat capacity c_e is introduced in the heat transfer equation:

$$704 \quad -\text{div } \vec{q} = \rho c_e \dot{T} \quad \text{with} \quad c_e = c + \frac{L}{1 + \left(\frac{1}{2} - 1\right) \frac{\rho_d}{\rho_w}} \cdot f \quad (2)$$

705 where \vec{q} is the conductive heat flux density, ρ is the total rock density, T is the temperature, c
 706 is the specific heat capacity of rock at constant pressure, L ($=3.35 \times 10^5 \text{ J kg}^{-1}$) is the latent
 707 heat of fusion for water, ρ_w ($=1000 \text{ kg m}^{-3}$) is the density of water, ρ_d is the dry bulk density
 708 and f is the frozen content of water.

709

710 f is given by the following equation:

711

$$712 \quad f = H(T_T - T) \cdot \frac{2T}{\theta^2} e^{-\left(\frac{T_T + T}{\theta}\right)^2} \quad (3)$$

713

714 where H is the Heaviside function, T_T is the melting point and θ is the freezing interval.

715

716 **Figure captions**

717 Fig. 1. Geological and structural map of the lower Romanche Valley with the location of the
 718 S echilienne landslide.

719

720 Fig. 2. Geology and geomorphology of the S echilienne landslide. (a) Structural sketch map
 721 with the location of the investigation gallery (G) and the borehole (B). (b) Rose diagram of
 722 structural data for the S echilienne slope (modified from Le Roux et al., 2011). (c) North–
 723 south cross section with the two main inferred sets of fractures (near-vertical N70 oriented
 724 and near-parallel to the slope). The lower seismic limit of the zone affected by the landslide is

725 drawn, considering the two V_p threshold limits at 3000 m s^{-1} (dotted red line) and 3500 m s^{-1}
726 (plain red line).

727

728 Fig. 3. Photographs of characteristic structures observed in the S echilienne slope. (a) Meter-
729 size fractures dipping nearly parallel to the slope and intersecting the N70 oriented near-
730 vertical fractures. This geometry contributes to the downward motion of the slope. (b)
731 Fracture parallel to the slope in the depletion zone. (c) Penetrative fracture set parallel to the
732 slope in the accumulation zone.

733

734 Fig. 4. Permafrost model showing five distinct layers and the temperature vertical profile
735 curve (red line) (modified from Riseborough et al., 2008). T_{bot} : mean annual temperature at
736 the bottom of the permafrost. T_{top} : mean annual temperature at the top of the permafrost. T_{maa} :
737 mean annual air temperature. T_{mag} : mean annual ground surface temperature.

738

739 Fig. 5. Paleo-temperature curves from the last 21,000 years (see text for details). Temperature
740 data are in dotted lines, with different colors according to the authors: (1) blue: Davis et al.
741 (2003); (2) red: Ilyashuk et al. (2009); and (3) green: Alley (2000). Chronologies of Davis et
742 al. (2003) and Ilyashuk et al. (2009) are based on ^{14}C calibrated ages, whereas that of Alley
743 (2000) is based on the GISP2 ice core. Both our data and the reference data are plotted on the
744 Cal BP scale. Reconstructed temperatures are in solid lines. T_{mwa} : mean annual temperature
745 curve for the warmest months. T_{maa} : mean annual air temperature curve. T_{mco} : mean annual
746 temperature curve for the coldest months. ATA : annual temperature amplitude. Four climate
747 periods (labeled A to D) are distinguished. The melting of the Romanche glacier in the valley
748 bottom until 14.3 ka is also indicated.

749

750 Fig. 6. 2D reconstruction of the Séchilienne slope geometry before destabilization
751 corresponding to the cross-section in Fig. 2c, obtained by balancing the depletion and
752 accumulation surfaces. The uncertainty on the landslide base (threshold between 3000 and
753 3500 m s⁻¹) is shown with red lines.

754

755 Fig. 7. 2D Séchilienne slope model with the applied boundary conditions. The glacier at 15.6
756 ka is in blue. T_{top} : Temperature deduced from the TTOP model and imposed at the surface.
757 T_{bg} : Temperature at the base of the glacier. The thickness of the glacier varies between 0 and
758 820 m.

759

760 Fig. 8. Results of 2D thermal numerical modeling. (a) Temperature distributions simulated
761 for model 1T (constant n -factors $n_T = 0.7$ and $n_F = 0.5$ and temperate glacier) at the seven
762 different times shown in (b). The permafrost and glacier extensions are shown in dark and
763 light blue, respectively. P1 to P4 show the locations where permafrost thickness values were
764 extracted (Table 3).

765

766 Fig. 9. 2D temperature distributions simulated for the three models 1C, 2C and 2T (1:
767 constant n -factors; 2: variable n -factors; C: cold glacier; T: temperate glacier) at different
768 times, applying the temperature curves in Fig. 5. The permafrost and glacier extensions are
769 shown with deep and light blue colors, respectively. P1 to P4 show the locations where
770 permafrost thickness values were extracted (Table 3).

771

772 Fig. 10. Maximum permafrost depths computed for the four models (default values) along the
773 Séchilienne slope before destabilization. They are compared to the landslide geometry,
774 considering the two thresholds at 3000 m s⁻¹ (dotted red line) and 3500 m s⁻¹ (solid red line).

775

776 Fig. 11. Succession of kinematics events affecting the Séchilienne slope after the thermal and
777 chronological constraints. (a) Mean annual air temperature curve from the Last Glacial
778 Maximum (21 ka) up to the present day in the Séchilienne slope at 1140 m a.s.l, the Holocene
779 Climatic Optimum period is indicated in grey. (b) Chronological constraints. Tg: age of the
780 glacier retreat at 1100 m a.s.l. (Le Roux et al. 2009); Tg^f: age of the glacier retreat in the
781 valley at 380 m; Tp: age of permafrost disappearance inferred from thermal modeling; and
782 Tdⁱ: initiation phase of the head scarp destabilization (Le Roux et al. 2009). (c) Kinematics of
783 the Séchilienne slope deduced from the chronological data related to glacier melting.
784 Permafrost evolution and landslide activity yielded a minimal pre-failure endurance of 6,200
785 years.

786

787 Table 1. Default values and variation ranges of the *n*-factors used in the scenario 2 for the four
788 thermal periods (A to D) and ground surfaces covered by the glacier.

789

790 Table 2. Default values and variation ranges for the parameters used in the model. See text for
791 details.

792

793 Table 3. Depth of the permafrost base (in m) at four sites (P1 to P4) shown in Fig. 8. The
794 maximum permafrost depth reached at each site is indicated in bold for the four models. In
795 the last column, the age of the permafrost disappearance is given for each model.