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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  $^{14}\text{C}$  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  $35^\circ$  to  $40^\circ$ , 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-decamic 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  $48 \times 10^6$   
192  $\text{m}^3$  and  $63 \times 10^6 \text{ m}^3$  by P-waves velocity ( $V_p$ ) thresholds at 3000 and 3500  $\text{m s}^{-1}$ , 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  $\pm 0.6 \text{ }^{10}\text{Be ka}$  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  $13.3 \pm 0.1 \text{ ka}$ . More closely,  
199 Delunel (2010) calculated a vertical glacier ablation rate between 0.30 and 0.37  $\text{m year}^{-1}$   
200 (mean value of 0.335  $\text{m year}^{-1}$ ) 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  $14.3 \pm 0.3 \text{ ka}$ . Therefore, the Séchilienne  
204 slope head scarp failure initiation, dated  $6.4 \pm 1.4 \text{ }^{10}\text{Be ka}$  (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 ( $T_{\text{maa}}$ ), the mean annual ground surface temperature ( $T_{\text{mag}}$ ), the mean annual  
221 temperature at the top of the permafrost body ( $T_{\text{top}}$ ) and the mean annual temperature at the  
222 bottom of the permafrost body ( $T_{\text{bot}}$ ). The differences between  $T_{\text{maa}}$  and  $T_{\text{mag}}$  on the one hand,  
223 and  $T_{\text{mag}}$  and  $T_{\text{top}}$  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  $T_{\text{top}}$ . 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 ( $n_{\text{F}}$  and  
230  $n_{\text{T}}$ , respectively). The  $n_{\text{T}}$  factor incorporates all microclimatic effects (radiation, convection,  
231 evapotranspiration, etc.) due to vegetation, while  $n_{\text{F}}$  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  $T_{\text{top}}$  to  $T_{\text{bot}}$ , 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 ( $T_{\text{maa}}$ ), 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 ( $T_{\text{mco}}$  and  $T_{\text{mwa}}$ , respectively).

246 The three  $T_{\text{maa}}$ ,  $T_{\text{mco}}$  and  $T_{\text{mwa}}$  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  $T_{\text{maa}}$  curve evolution from 40 to 0 ka (Alley, 2000);  
251 (3) quantitative pollen climate reconstructions for Central Western Europe giving thermal  
252 anomalies of  $T_{\text{maa}}$ ,  $T_{\text{mco}}$ , and  $T_{\text{mwa}}$  with respect to the present-day temperature since 12.0 ka  
253 (Davis et al., 2003); and (4)  $T_{\text{mwa}}$  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  $T_{\text{maa}}$ ,  $T_{\text{mwa}}$  and  $T_{\text{mco}}$  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  $T_{\text{maa}}$ ,  $T_{\text{mco}}$ , and  $T_{\text{mwa}}$  temperature curves at the top of the Séchilienne slope (1140 m a.s.l.)  
263 during the same period (Fig. 5). The  $T_{\text{maa}}$  curve was extended to 21 ka by using the Greenland  
264 ice core records (Alley, 2000), while the  $T_{\text{mwa}}$  curve was constrained from 10.8 to 14.0 ka by  
265 using chironomids (Ilyashuk et al., 2009). Determining  $ATA$  from the  $T_{\text{mwa}}$  and  $T_{\text{maa}}$  curves  
266 allowed the  $T_{\text{mco}}$  curve to be computed during the same period of time (Fig. 5). Finally, the  
267 only missing data ( $T_{\text{mco}}$  and  $T_{\text{mwa}}$  curves between 21 and 14 ka) were estimated by assuming a  
268 linear relationship between  $T_{\text{maa}}$  and  $ATA$  values. These composite temperature curves ( $T_{\text{mwa}}$ ,  
269  $T_{\text{mco}}$ , and  $T_{\text{maa}}$ ) were used as input data in the thermal modeling of the Séchilienne 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  $T_{\text{maa}}$  around  $-8^{\circ}\text{C}$ ; a  
274 warmer period B from 14.7 to 13.0 ka with a  $T_{\text{maa}}$  between  $-4.0^{\circ}\text{C}$  to  $+5.5^{\circ}\text{C}$ ; a short colder  
275 period C until 11.6 ka with a mean  $T_{\text{maa}}$  around  $-10^{\circ}\text{C}$ ; and a warmer period D from 11.6 to 0  
276 ka (Holocene) with a  $T_{\text{maa}}$  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 Séchilienne 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  $T_{\text{top}}$  and  $0^{\circ}\text{C}$ ,  
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  $T_{\text{maa}}$  value of  $-8^{\circ}\text{C}$  and  
297  $-10^{\circ}\text{C}$  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  $0.335 \text{ m yr}^{-1}$  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  $65 \text{ mW m}^{-2}$ , 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 ( $\phi$  and  $\rho_d$ ) were determined by previous laboratory tests performed on  
329 micaschist samples (Le Roux et al., 2011), with default values of 3.7% and  $2730 \text{ kg m}^{-3}$ ,  
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  $\theta$  of  $0.3^{\circ}\text{C}$  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  $40^{\circ}$  Séchilienne slope (Safanda, 1999; Table 1). The  $T_{\text{top}}$  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  $2^{\circ}\text{C}$  (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  $\pm 10$  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 ( $\xi$ ,  $r_k$ ,  $n_T$ ,  $n_F$  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<sup>-1</sup>) 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 (Tg<sup>f</sup>), 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 (Tp), 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 (Td<sup>i</sup>), 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  $T_{\text{maa}}$  and  
 690  $T_{\text{mag}}$ . 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,  $I_{\text{FA}}$  and  $I_{\text{TA}}$  are the air seasonal freezing and thawing  
 697 degree-day indexes,  $I_{\text{FS}}$  and  $I_{\text{TS}}$  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,  $\rho$  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,  $\rho_w$  ( $=1000 \text{ kg m}^{-3}$ ) is the density of water,  $\rho_d$  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  $\theta$  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  $3000 \text{ m s}^{-1}$  (dotted red line) and  $3500 \text{ m s}^{-1}$   
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).  $T_{\text{bot}}$ : mean annual temperature at  
736 the bottom of the permafrost.  $T_{\text{top}}$ : mean annual temperature at the top of the permafrost.  $T_{\text{maa}}$ :  
737 mean annual air temperature.  $T_{\text{mag}}$ : 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  $^{14}\text{C}$  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.  $T_{\text{mwa}}$ : mean annual temperature  
745 curve for the warmest months.  $T_{\text{maa}}$ : mean annual air temperature curve.  $T_{\text{mco}}$ : 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<sup>-1</sup>) 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.  $T_{\text{top}}$ : Temperature deduced from the TTOP model and imposed at the surface.  
757  $T_{\text{bg}}$ : 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<sup>-1</sup> (dotted red line) and 3500 m s<sup>-1</sup> (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<sup>f</sup>: age of the glacier retreat in the  
781 valley at 380 m; Tp: age of permafrost disappearance inferred from thermal modeling; and  
782 Td<sup>i</sup>: 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.