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HAL Authorization

1 **Water Flows in Rockwall Permafrost: a Numerical Approach Coupling Hydrological**
2 **and Thermal Processes**

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7 **Key Points:**

- 8 • Coupling thermal and hydrological processes is necessary to better understand rockwall
9 permafrost dynamics and geomorphic processes
- 10 • Water flows into bedrock fractures substantially affect permafrost aggradation and
11 degradation patterns
- 12 • Water accumulation into ice-cemented bedrock may cause significant hydrostatic
13 pressures favoring rockwall destabilization
14

Abstract

Rockwall permafrost is extremely sensitive to climate change and its degradation is supposedly responsible for the recent increase in periglacial rock slope failures. Investigations of rockwall permafrost dynamics and mechanics have so far neglected possible hydrogeological processes acting in bedrock fractures. In this study, we propose the first numerical approach to couple thermal and hydrological processes in alpine rockwall permafrost and show that the latter have major effects on permafrost (thermal) dynamics and mechanics when the fractures and/or rock matrix are saturated. Water flows into fractures favor deep-reaching of the permafrost body by driving cold water top-down. Ice-filled fractures delay permafrost thawing in a first stage due to latent heat consumption but then accelerate it when the ice starts to melt. Thus, frozen fractures may subsist in thawed bedrock while thawing corridors may form in frozen bedrock. As a result, temperature gradients are exacerbated. When connected fractures thaw, bottom-up permafrost degradation can occur through upwards propagation of thawing wedges delineated by these fractures. High hydraulic head values are associated to perched water table over or within the impermeable permafrost body, and correspond to hydrostatic pressures that can reach critical values in terms of rockwall stability. These results bear strong implications to understand permafrost response to climate signals, periglacial geomorphology and hazards assessment as well as alpine hydrothermal processes.

Plain Language Summary

Permafrost (i.e. ground that remains at or below 0°C for at least 2 consecutive years) affecting the steep rock slopes in alpine or polar environments is extremely reactive to climate change. Its degradation (i.e. warming and thawing) is supposedly responsible for the increase in rock falls frequency and magnitude recently observed. Research conducted during the past decade have improved understanding of rockwall permafrost response to climate change and its role in rockwall destabilizations. However, the role of water flows into bedrock fractures that may result of rainfall, snow or glacier melt, has been overlooked because of the lack of understanding of these processes and the difficulty to study them. Based the recent developments in numerical tools, we here propose the first study that model water flows into rockwall permafrost to assess their role on permafrost distribution and evolution as well as in destabilization. The results show that permafrost distribution and evolution is more complex than originally thought when water flows were ignored. Permafrost can form well deeper due to cold water infiltrating from the mountain top, down in the rock mass. The ice contained in bedrock fractures may significantly delay permafrost thawing in th on hand, but can also provoke thawing acceleration when its start to melt in the other hand by forming thawing corridors and wedges along and between fractures. Water columns developing over or within the frozen rock mass may provoke high hydrostatic pressures that could favor bedrock destabilization. This study invites to rethink current knowledge on rockwall permafrost reaction to climate change, rockwall permafrost destabilization processes and alpine hydrogeology.

53

54 **1 Introduction**

55 Rockwall permafrost has been investigated since the early 2000s in conjunction with the
56 observed increase in periglacial rockwall failures (e.g. Allen et al., 2009; Gruber et al., 2004). So
57 far, researches have been mostly driven by the hypothesis that permafrost degradation, i.e.
58 deepening of the active layer (seasonal thaw) or warming of the permafrost body, provokes
59 bedrock destabilization by altering ice-joints in bedrock fractures (Gruber & Haeberli, 2007;
60 Krautblatter et al., 2013). Laboratory experiments have largely supported this hypothesis. Davies
61 et al. (2001) have for example shown that the factor of safety of an ice-joint decreases with
62 increasing temperature from -5 to -0.5°C where it reaches its minimal value. More detailed
63 quantification of the temperature dependency of ice-joint strength was recently done by Mamot
64 et al. (2018), while a comprehensive model of bedrock destabilization in relation with permafrost
65 dynamics was proposed by Krautblatter et al. (2013).

66 In parallel, thermal modeling has shown the enhanced sensitivity of alpine rockwall permafrost
67 to air temperature signal (Magnin et al., 2017; Myhra et al., 2017; J. Noetzli & Gruber, 2009;
68 Jeannette Noetzli et al., 2007). Permafrost models were used to assess thermal conditions at
69 location of periglacial rockwall failures, confirming that many rockfall events occurred in warm
70 permafrost conditions (i.e. permafrost > -2°C; Deline et al., 2013; Frauenfelder et al., 2018;
71 Ravanel et al., 2017). Transient temperature models generally account for heat conduction
72 processes in a saturated, homogeneous and isotropic rock media. But rockwall permafrost
73 models also showed that rockfalls may occur in cold permafrost areas and prior to the maximum
74 active layer depth (Gruber et al., 2004; Luethi et al., 2015; Ravanel et al., 2017). Assuming that
75 such rockfalls likely result of ice-joint warming and thawing, this implies that non-conductive
76 heat transfers may play a key role in permafrost degradation such as found for more porous
77 terrains constituted of non-consolidated material and allowing significant water infiltration
78 (Luethi et al., 2017; Scherler et al., 2010). This latter hypothesis is supported by the investigation
79 from Hasler et al. (2011a) on water percolation along an experimental ice-filled cleft cementing
80 two bedrock compartments. This study showed accelerated permafrost degradation and cleft-ice
81 erosion due to advective heat transport by water flows. But such findings are difficult to verify
82 through field observations. Direct temperature measurements into boreholes by Phillips et al.
83 (2016) recorded short and intermittent temperature increase after snow melt, hinting at water
84 percolation down to the thermistor chain. Geophysical soundings from Krautblatter & Hauck
85 (2007) and Keuschnig et al. (2017) have suggested deep-reaching cleft-water systems in
86 rockwall permafrost during the thawing season. Observations of rockfall scars with ice-coat and
87 water flow marks are also frequently reported (Fischer et al., 2010; Frauenfelder et al., 2018;
88 Geertsema et al., 2006; Ravanel et al., 2017; Walter et al., 2020) and strengthen the idea that
89 water circulation may play a key role in the triggering of periglacial rockwall failure; either by
90 cleft ice erosion and a resulting loss of bonding at the rock-ice interface or through hydrostatic
91 pressures.

92 Coupling thermal, hydrological and mechanical processes is a major research perspectives
93 pointed out by Krautblatter et al. (2012) to go over current knowledge limits. Hydrological
94 processes remain the least understood because they are non-linear and are difficult to observe.
95 However, in the recent years, a wealth of numerical codes have been developed to fully couple
96 thermal and hydraulic processes (Grenier et al., 2018). They have been so far mostly designed
97 and used for high-latitude flat regions (e.g. Bense et al., 2012; McKenzie & Voss, 2013;

98 Rowland et al., 2011) but they offer new opportunities to explore and conduct systematic
99 investigations on the role of hydrological processes in rockwall permafrost as well.

100 In this study, we claim that rockwall permafrost dynamics should be understood as a
101 hydrogeological problem and we thus propose to rethink former thermal modeling approach by
102 adopting a hydrogeological conceptualization to allow water input and flow at a scale of a
103 mountain flank.

104 We aim at (i) proposing an appropriate numerical procedure to fully couple thermal and
105 hydrological processes in alpine rockwall permafrost, (ii) testing the sensitivity of rockwall
106 permafrost to fluid flows, and (iii) pointing out possible research directions hinted at by these
107 new developments.

108 We first describe our conceptual and numerical approach to couple thermal and hydrological
109 processes in saturated and unsaturated conditions. We then run four simulations with various
110 saturation levels and water input into a fractured bedrock medium. Our simulations show the
111 major importance of fluid flows into bedrock thermal dynamics and fluid pressure distribution,
112 under saturated conditions. We finally highlight that developing such models is a key to improve
113 understanding of rockwall permafrost (thermal) dynamics, alpine hydrogeology and
114 morphodynamics (mechanics).

115 **2 Hydrothermal simulations**

116 2.1. Preliminary concerns

117 Magnin et al. (2017) have shown that 2D thermal models running over pluri-decadal time
118 scale and forced with local air temperature measurements are able to reproduce measured
119 borehole temperature in alpine rock wall at a depth of > 8-10 m. Such models follow the
120 approach designed by Noetzli et al. (2007) which consider typical alpine topography and the
121 related topoclimatic control (sharp variation in elevation and sun-exposure resulting in important
122 air temperature lapse rate and highly variable incoming short-wave solar radiation). They
123 consider a saturated and homogeneous rock media with relatively high porosity compared to
124 most rock types in order to indirectly account for fractures. In these approaches, water flows are
125 neglected. However, water flows into frozen bedrock have been investigated for several decades
126 in the frame of frost weathering studies. These investigations mostly focus on water migration
127 through the pore space as a result of temperature change and freezing (see Matsuoka & Murton
128 (2008) for a review). Such processes are essential to understand bedrock fracturing and the
129 preparation to rock slope failures. However, to address the triggering of rockwall failures, macro-
130 scale processes acting at the already existing failure plan are more relevant.

131 Recently, the study of Hasler et al. (2011a) have set up a conceptual model describing water
132 circulation patterns in fractures of frozen rockwalls to question their thermal effect. This
133 conceptual model first pointed out that non frozen fractures are preferential flow paths as
134 moisture migration through bedrock pores has a minor role in the thermal regime of low-porosity
135 rock. Then, it stated that high hydraulic gradients related to slope angle and high permeability
136 due to macroscopic clefts results in unsaturated conditions and that bedrock permeability
137 depends on the cleft ice content.

138 Because cleft saturation results of complex processes involving repeated freeze and thaw as well
139 as discharge and loading cycles, the level of saturation of rockwall permafrost is a poorly known

140 parameter. In this respect, we propose to steadily upgrade existing thermal models by first
 141 integrating macro and micro-scale water flows in a “traditional” approach assuming a fully
 142 saturated porous and fractured bedrock medium. Then, we initiate a step towards the
 143 consideration of unsaturated conditions by setting up an unsaturated model domain and running
 144 simulations with limited or enhanced water flows. To do so, we adopt a hydrogeological
 145 conceptualization by setting up a water table in an initially unfrozen mountain flank.

146 2.2 Mathematical approach

147 The equation for transient flow through an anisotropic 3D porous medium is obtained by
 148 plugging the Darcy law into the continuity equation as follow:

$$149 \frac{\partial}{\partial x} \left(K_{Sxx} \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{Syy} \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_{Szz} \frac{\partial h}{\partial z} \right) = S \frac{\partial h}{\partial t}. \quad (1)$$

150
 151 with K_S the hydraulic conductivity at saturation, h the hydraulic head, S the specific storage, t the
 152 time and $x/y/z$ the three axes of space.

153 Into unsaturated media, the pressure head ψ is negative and $h = z + \psi$ is lower than the same
 154 hydraulic head at the same altitude under saturated conditions. The first equation for unsaturated
 155 media thus becomes:

$$156 \frac{\partial}{\partial x} \left(K_{\psi xx} \frac{\partial \psi}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{\psi yy} \frac{\partial \psi}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_{\psi zz} \left(\frac{\partial \psi}{\partial z} + 1 \right) \right) = C(\psi) \frac{\partial \psi}{\partial t} \quad (2)$$

157
 158 with $C(\psi) = \frac{d\theta}{d\psi}$, θ being the moisture content. This second equation reduced to 1D vertical flow
 159 is known as Richard’s equation properly: $\left(K_{\psi zz} \left(\frac{\partial \psi}{\partial z} + 1 \right) \right) = C(\psi) \frac{\partial \psi}{\partial t}$. (3)

160
 161 K_{ψ} designs the hydraulic conductivity for a given ψ . The values of K_{ψ} are obtained through the
 162 classical van Genuchten – Mualem relationship :

$$163 s_e = \begin{cases} \frac{1}{[1+(A|\psi|)^n]^m} \text{ for } \psi < 0 \\ 1 \text{ for } \psi \geq 0 \end{cases}, K_R = s_e^{\frac{1}{2}} \left\{ 1 - \left(1 - s_e^{\frac{1}{m}} \right)^m \right\}^2, \quad (4)$$

164
 165 with the effective saturation s_e given by $s_e = \frac{\theta - \theta_r}{\theta_s - \theta_r}$, where θ corresponds to a given ψ , θ_s is the
 166 moisture content at saturation, θ_r is the residual moisture content, K_R the relative hydraulic
 167 conductivity (relative to K_S), and A , m and n are van Genuchten parameters. For this study, we
 168 kept the van Genuchten – Mualem relationship, but other relationships could be used
 169 (Haverkamp, Brooks & Corey, etc.).

170 The flow velocities obtained from the previous calculations are integrated into the advective
 171 dispersive-diffusive heat transport equation which is usually expressed as follows:

$$172 \frac{\partial \left[(\varphi \rho_C)_L + (1 - \varphi) (\rho_C)_S \right] T}{\partial t} = -\nabla \cdot \left[(\rho_C)_L q T - \Lambda \nabla T \right] \quad (5)$$

173
 174 with φ the porosity (a-dimensional), ρ_{CL} and ρ_{CS} the volumetric heat capacities ($\text{J} \cdot \text{m}^{-3} \cdot \text{K}^{-1}$) of the
 175 liquid and solid phases respectively, Λ the hydrodynamic thermal dispersion tensor, ($\text{J} \cdot \text{m}^{-1} \cdot \text{s}^{-1} \cdot \text{K}^{-1}$)

178 ¹) that includes thermal conductivity, T the temperature (K) and q the apparent flow velocity
 179 from Darcy or Richards equation ($\text{m}\cdot\text{s}^{-1}$). Then, the ice phase is included in the solid phase to
 180 modify only one parameter of thermal conductivity (and not the one related to fluids). The solid
 181 thermal conductivity λ ($\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$) remains:

$$182 \quad \lambda_s = \lambda_{s,0} + \frac{\varepsilon_i (\lambda_i - \lambda_s)}{1 - \varepsilon} \quad (6)$$

183 with ε_i the bulk fraction of the ice and ε the bulk fraction occupied both by water and air. In the
 184 solid, the thermo-dispersion tensor is linked to the thermal conductivity through the solid bulk
 185 volume fraction, and is sufficient here, the fluid convection being beyond the scope of this study.
 186 Into the fractures, we used the Hagen-Poiseuille flow formulation that characterize laminar flow
 187 that is accounted for as follow (modified from Diersch 2004):

$$188 \quad Q = -\frac{a^3}{12\mu} \|\overrightarrow{\text{grad}P} - \rho_w \vec{g}\| \quad (7)$$

189 where Q is the water discharge, a the aperture of the fracture in m, μ is the dynamic viscosity, P
 190 the water pressure, ρ_w the density of water and g the gravity's acceleration.

191 Concerning the addition of the ice in the whole medium, it is expressed throughout the bulk volume
 192 as: $\varepsilon_a + \varepsilon_w + \varepsilon_i + \varepsilon_r = 1$, with ε_a the bulk fraction of air, ε_w the bulk fraction of water, ε_i the bulk
 193 fraction of ice and ε_r the bulk fraction of rock. A relation is established between ice and liquid.
 194 This relation, called the freezing function (Clausnitzer & Mirnyy, 2015) links the mass fraction per
 195 bulk volume of the unfrozen liquid to the total liquid mass:

$$196 \quad F = \frac{\varepsilon_w \rho_w}{\varepsilon_w \rho_w + \varepsilon_i \rho_i} \quad (8)$$

(3)

198 where ρ is the density of the corresponding phase (i for ice and w for water). This function F
 199 decreases with the fraction of ice. With a freezing point T_0 , the ice forms gradually within a
 200 predefined temperature interval $\left[T_0 - \frac{\Delta T}{2}, T_0 + \frac{\Delta T}{2} \right]$ of the length ΔT .

201 2.3 Numerical approach

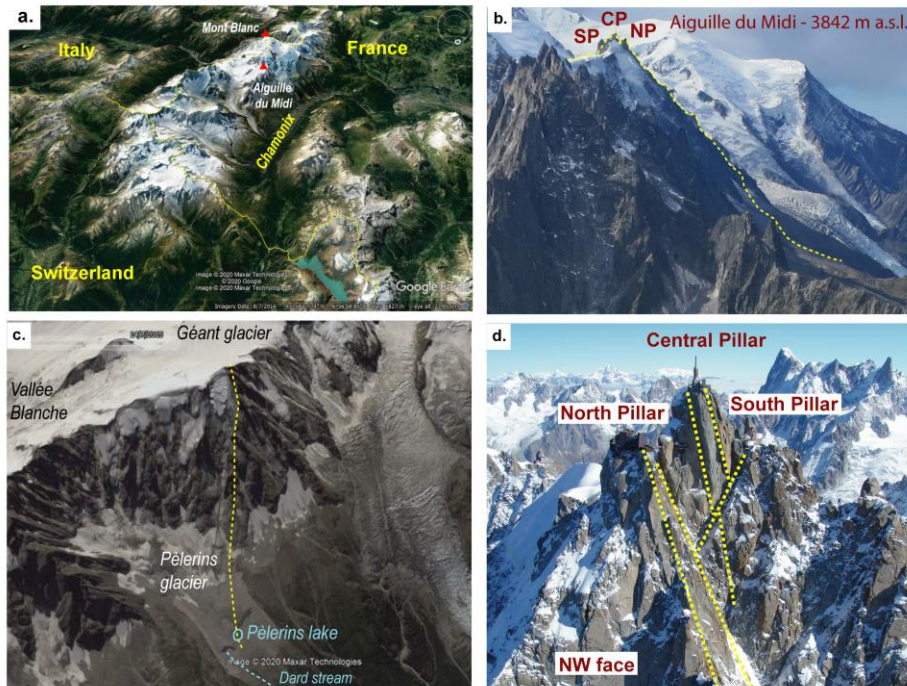
202 We implement the hydrothermal model described above with Feflow® 7.0 and 7.2 (DHI-
 203 WASY GmbH), a finite elements numerical code for simulation of saturated or unsaturated flow
 204 coupled with mass and/or heat transport. In the recent years, the *Pi-Freeze* plug-in was added to
 205 take into account freezing and thawing processes. The use of Feflow and *Pi-Freeze* for
 206 hydrothermal modeling applied to permafrost cases has been successfully benchmarked in a
 207 model intercomparison study (Grenier et al., 2018).

208 2.3.1 Study site and model domain

209 2.3.1.1 Study site settings

210 In our study, we take the Aiguille du Midi (Mont Blanc massif, northwestern European
 211 Alps) as a study case. The Aiguille du Midi is a set of three granitic pillars culminating at 3842
 212 m a.s.l. on the western margin of the Mont Blanc massif in France (northwestern European Alps;
 213 Fig. 1a-b). Its South pillar stands above the Glacier du Géant by about 200 m while its North
 214 Pillar dominates Chamonix, extending over a height of about 1400 m down to the Pèlerins

215 glacier (Fig. 1c). The site is accessible by a cable car and host about half a million tourists every
 216 year (Fig. 1d). It is the starting point of the Vallée Blanche, a very popular ski route and one of
 217 the normal route to the Mont Blanc summit (4809 m a.s.l.). The Aiguille du Midi bedrock is
 218 structured by various fracture sets with predominance of N050 and N150 (Fig. 1d).
 219 Thermal modeling with Feflow and *Pi-Freeze* was already applied on the top part of this site and
 220 evaluated against borehole temperature measurements by Magnin et al. (2017) considering
 221 saturated and homogeneous conditions with no water flow. We here extend the model domain to
 222 ensure coherence with hydrogeological concerns and design four study cases to test rockwall
 223 permafrost sensitivity to water flow.



224
 225 **Figure 1.** Study site and main features for determining model domain. **a.** The Mont Blanc
 226 massif. **b.** The Aiguille du Midi with the topographical profile considered in this study. SP:
 227 South Pillar, CP: Central Pillar, NP: North Pillar. **c.** The topographical profile considered in this
 228 study and main water outlet. **d.** Pictures of the Aiguille du Midi site with some of the main
 229 fractures identified to draw the model geometry (*Pic: S. Gruber*).

230

231

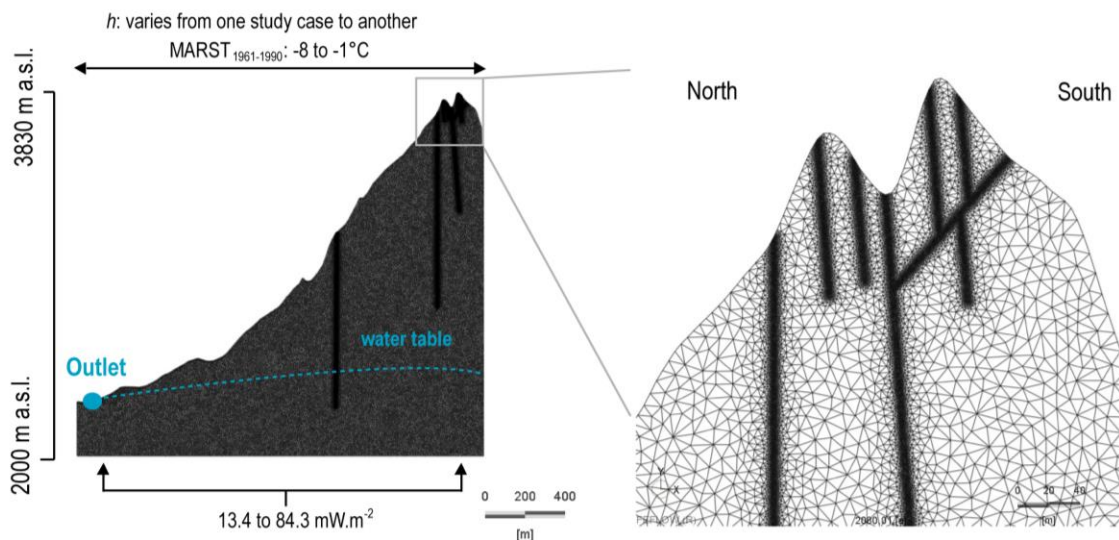
2.3.1.2 General features of the model domain

232 The topographical profile was extracted from a 4-m-resolution DEM provided by the
 233 regional authority (*Régie de Gestion de Donnée Haute Savoie*) to draw a 2D model domain. It
 234 extends from the bottom of the Central Pillar, at the edge of the Géant glacier, over the Central
 235 and North Pillars down to the foot of the glacier des Pèlerins where a lake forms intermittently
 236 (Fig. 1c). The profile was closed at its bottom and discretized into a network of 140,509 triangular
 237 elements and 70,879 nodes (Fig. 2). The mesh was refined around fractures because the high
 238 hydraulic gradient in between the fractures and the rock matrix could challenge the calculation.

239 Fractures were drawn to determine preferential flow paths by selecting a couple of the main
 240 fracture sets of the Aiguille du Midi based on various photos and field observations (Fig. 1d).

241 Hydraulic conductivity is mostly influenced by fracture connectivity, density and aperture
 242 (Maréchal et al., 2004; Zhao, 1998, Long and Witherspoon 1985, Snow 1979). In a permafrost
 243 context, the hydraulic conductivity is furthermore dependent on the temperature as it is related to
 244 the unfrozen water content (Burt & Williams, 1976). Interaction of the fracture characteristics and
 245 high temperature variability in alpine rockwalls is supposedly responsible for high and complex
 246 hydraulic gradients, which are also induced by the large elevation difference between the summit
 247 and the outlet. However, as a preliminary investigation, our study aims at ensuring results
 248 transparency and readability in order to facilitate interpretation of water flow effects in permafrost
 249 dynamics. We thus ignored fracture density and aperture and designed bedrock fractures with a
 250 concern of (i) connectivity, (ii) representativeness of the thermally-induced hydraulic gradient, and
 251 (iii) simplicity.

252 We selected a couple of the main fracture sets easily perceptible (Fig. 1d) and crossing both the
 253 rather colder north-exposed areas and warmer south-exposed areas (Fig. 2). Fracture aperture
 254 was set to 5 cm for long ones that represent faults and supposedly have an effect on water flow at
 255 the mountain flank scale and 2.5 cm for short ones that rather have an effect at the outcrop scale.
 256 We assumed constant aperture through depth. Thus, compared to real-world cases, we consider a
 257 realistic but largely simplified model geometry.
 258



259 **Figure 2:** Model domain and initial conditions. The MARST was extracted from Magnin et al.
 260 (2015) and is representative of the 1961-1990 period. h is the hydraulic head. The water table was
 261 set after a host of trial and error run.
 262
 263

264 2.3.2 Model parameters and consistent boundary conditions

265 In thermal modeling, main boundary conditions are the initial surface temperature and
 266 geothermal heat flux. Following previous thermal modeling approach of Magnin et al. (2017), the
 267 initial rock surface temperature was extracted from the temperature map of Magnin et al. (2015)
 268 implemented on the same 4 m resolution DEM as the one used for the topographical profile. The
 269 map represents the Mean Annual Rock Surface Temperature (MARST) calculated after the
 270 statistical model calibrated by Boeckli et al. (2012) which explains the MARST according to the

271 mean air temperature of the period 1961-1990 and the potential incoming solar radiation. The
272 initial MARST ranged from -8°C to -1°C according to elevation and sun-exposure (see Fig. S1).

273 At the lower boundary, a geothermal heat flux was applied. We used the heat flux value calculated
274 in the study of Magnin et al. (2017) at an elevation of 2000 m a.s.l when applying $85\text{ mW}\cdot\text{m}^{-2}$ at 5
275 km below the Aiguille du Midi summit.

276 The thermal conductivity of the rock was set to $3\text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ which stands for a conservative value
277 for saturated granitic rock (Cho et al., 2009) and the heat capacity of the rock mass was set to 1.8
278 $\text{MJ}\cdot\text{m}^3\cdot\text{K}^{-1}$. A bedrock porosity of 5% was assumed for the intact rock mass which accounted for
279 a greater fracture density than the considered one. For the water, the thermal conductivity was set
280 to $0.65\text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ and the heat capacity was set to $4.2\text{ MJ}\cdot\text{m}^3\cdot\text{K}^{-1}$. Freezing was setup to occur at
281 -1°C , the latent heat of fusion was $334\text{ kJ}\cdot\text{kg}^{-1}$.

282

283 2.3.3. Study cases with variable saturation and fluid flows

284 We then test four study cases with contrasted hydrological setup (Tab. 1) so that a large
285 panel of behaviors can be analyzed.

286 2.3.3.1 Two cases in saturated conditions

287 Case *Sa-NF* is a “traditional” modeling approach with saturated conditions and no fluid
288 flow (only conductive heat transport, similar to Magnin et al. (2017)). A hydraulic head of 3830
289 m a.s.l. is applied at all surface nodes (Tab. 1), which corresponds to the highest elevation point
290 in the profile, and thus provokes saturated conditions and the absence of fluid flows.

291 *Sa-Fl* was also saturated but accounted for fluid flow (allowing forced convection) by setting up
292 a hydraulic head equal to the surface elevation for all surface nodes and was forced to remain
293 constant. Water flows are provoked by the hydraulic processes related to topographical and
294 thermal-related hydraulic gradient, and are further enhanced by a constant recharge and
295 discharge at the rock surface to maintain a constant head. No outlet was specifically designed but
296 was accounted for by the software itself.

297

298 2.3.3.2 Two cases in unsaturated conditions

299 We then run simulations in an unsaturated bedrock medium by setting up a water table
300 with an initialization run (sect. 2.3.4). Little is known about water table levels in alpine
301 environments (Cochand et al., 2019). Tunneling work and boreholes in crystalline rock showed
302 that the unsaturated zone is generally several hundred meters below the bedrock surface
303 (Maréchal, 1998; Masset & Loew, 2010). We determined the water table by a set of trial and
304 error runs. To do so, we set up a specific outlet that we determined by focusing on possible
305 evidence of perennial water offspring that were found on the forefield of the Pèlerins glacier
306 where water streams exist and a lake sometimes forms in the morainic material (Fig.1b). To
307 simplify our model domain and calculations, we neglected subglacial outlets. The recharge area
308 is the ground surface to which we applied a water flow constrained by the hydraulic head h
309 (Neuman conditions automatically turned into Dirichlet conditions). Trial and errors simulations
310 were run in unfrozen conditions with water input in meter per day (Neuman conditions) until a

311 realistic and steady water table allowing further water infiltrations and preventing from either
 312 water outburst or substantial drainage was found. This initial water table was about 1000 m
 313 below the highest point of the topographical profile (Fig. 2). The resulting hydraulic head ranged
 314 between 2260 (elevation of the outlet) and 3826 m a.s.l.

315 The case *uS-LF* relates to unsaturated conditions determined as describe here above and with
 316 very little water flow because the low hydraulic conductivity, even at saturation and the low
 317 porosity do not allow large flow into the rock matrix. The hydraulic head corresponding to the
 318 steady water table was applied as boundary condition (Dirichlet condition) replacing the Neuman
 319 condition used for trial and error simulations described above. These boundary conditions varied
 320 within a limited hydraulic head range (see sect. 2.3.4) in order to maintain unsaturated conditions
 321 during freezing and thawing phases, and conversely to *Sa-Fl*, this resulted in very little fluid
 322 flows because only a very low recharge was necessary.

323 Finally, a fourth case named *uS-Fl*, aimed at testing the effect of enhanced water flow (discharge
 324 and recharge) into bedrock fractures. Because of some numerical limits (see discussion) and the
 325 aim of results transparency, water flow was enhanced into specific fractures and their vicinity,
 326 and at different periods (either during freezing or during thawing, see sect. 2.3.4). To do so, the
 327 hydraulic head varied during transient simulations at the top of the concerned fractures. This
 328 variable water flow forcing aimed at better see what happens into progressively suturing bedrock
 329 medium to refine characterization of water flow patterns and effects.
 330

	<i>Sa-NF</i>	<i>Sa-Fl</i>	<i>uS-LF</i>	<i>uS-Fl</i>
<i>h</i>	3830 m a.s.l.	Elevation of the surface	2260 to 3826 m a.s.l.	
outlet	None	Created <i>ad hoc</i> by the simulation process (handled by the software)	Designed at the snout of the Pèlerins glacier	
Water flow	None	Constrained by steady hydraulic head provoking hydraulic gradient	±0.25% of the initial hydraulic head	± 0.25% of the initial hydraulic head and additional water input (+0.7 m) until saturation is reached at selected fractures during freezing and thawing

331 **Table 1.** Hydrological settings and forcing for the four study cases.

332

333

2.3.4 Initialization and transient simulations

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335

2.3.4.1 Initialization

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The models were first run for 3000 years until the water table (*uS-LF* and *uS-Fl* cases only) and the bedrock temperature reached a steady state. A value of +7°C was applied to the MARST of the 1961-1990 period to generate a non-frozen steady state. For *Sa-Fl*, the water input is controlled by the Dirichlet boundary conditions while for *uS-LF* the Neuman boundary conditions are adjusted such as a steady unsaturated zone is maintained through time.

344 2.3.4.2 Transient simulations and forcing data

345 Transient simulations are run at an automatically adapted and variable time step and cover
346 a freezing period and a thawing period. They were run from 0 AD (results of the initialization run)
347 to 2100 AD. Using calendar dates was not necessary for our experiment as we do not intend to
348 discuss past or future permafrost evolution patterns but we used them as the temperature time
349 series was already created in a previous study (Magnin et al., 2017), that it represent realistic
350 climate evolution, and that the dates facilitate results description.

351 In a first step, the entire bedrock is steadily frozen from 50 AD to 1550 AD (beginning of the Little
352 Ice Age). To do so, a progressive temperature decrease of 8°C is applied at the surface, such as the
353 MARST in 1550 AD was 1°C lower than the MARST of the 1961-1990 period. In a second step,
354 the freezing is kept constant (constant MARST) until the end of the Little Ice Age (1850). Within
355 these two first steps, the initial hydraulic head is steadily reduced by -0.25% (Dirichlet condition)
356 for *uS-LF* and *uS-Fl* as soon as the bedrock freezes, considering that no flow enters into the frozen
357 rock and to prevent saturation when ice seals fractures. As the freezing is not uniform because of
358 variable MARST, the decrease in h is not uniform either. For *uS-Fl*, water is added (Neumann
359 condition) in the only fracture reaching the water table during freezing, until saturation is reached.
360 The freezing begins in 850 AD in this fracture and the head starts to increase in 852 AD at the top
361 of the fracture until saturation ($+0.7$ m) in order to simulate infiltration.

362 After this steady freezing, a progressive thawing starts by 1850 AD. The MARST first increases
363 linearly by $+1^{\circ}\text{C}$ until 1990 AD (to match the MARST of the period 1961-1990) and then varies
364 from a time step to another following the air temperature anomaly registered by *Météo France* in
365 Chamonix and compared to the mean air temperature of 1961-1990. This approach strictly follows
366 the one from Magnin et al. (2017) which has been shown reliable to reconstruct realistic permafrost
367 evolution. Simulation are run until 2100 AD and temperature anomaly is then calculated according
368 to the IPSL_LR model of the CMIP5 project from 2015, assuming a greenhouse gas emission
369 scenario of $+4.5$ $\text{W}\cdot\text{m}^2$ by the end of the 21st century (RCP4.5, see Fig. 4 in Magnin et al. (2017))
370 for summary of the air temperature anomaly). Within the thawing period which lasts from 1850 to
371 2100 AD, the hydraulic head increases by 0.25% for *uS-LF* and *uS-Fl* as soon as the bedrock starts
372 to thaw. For *uS-Fl*, water input ($+0.8$ m) is added in two other interconnected fractures until
373 saturation is reached.

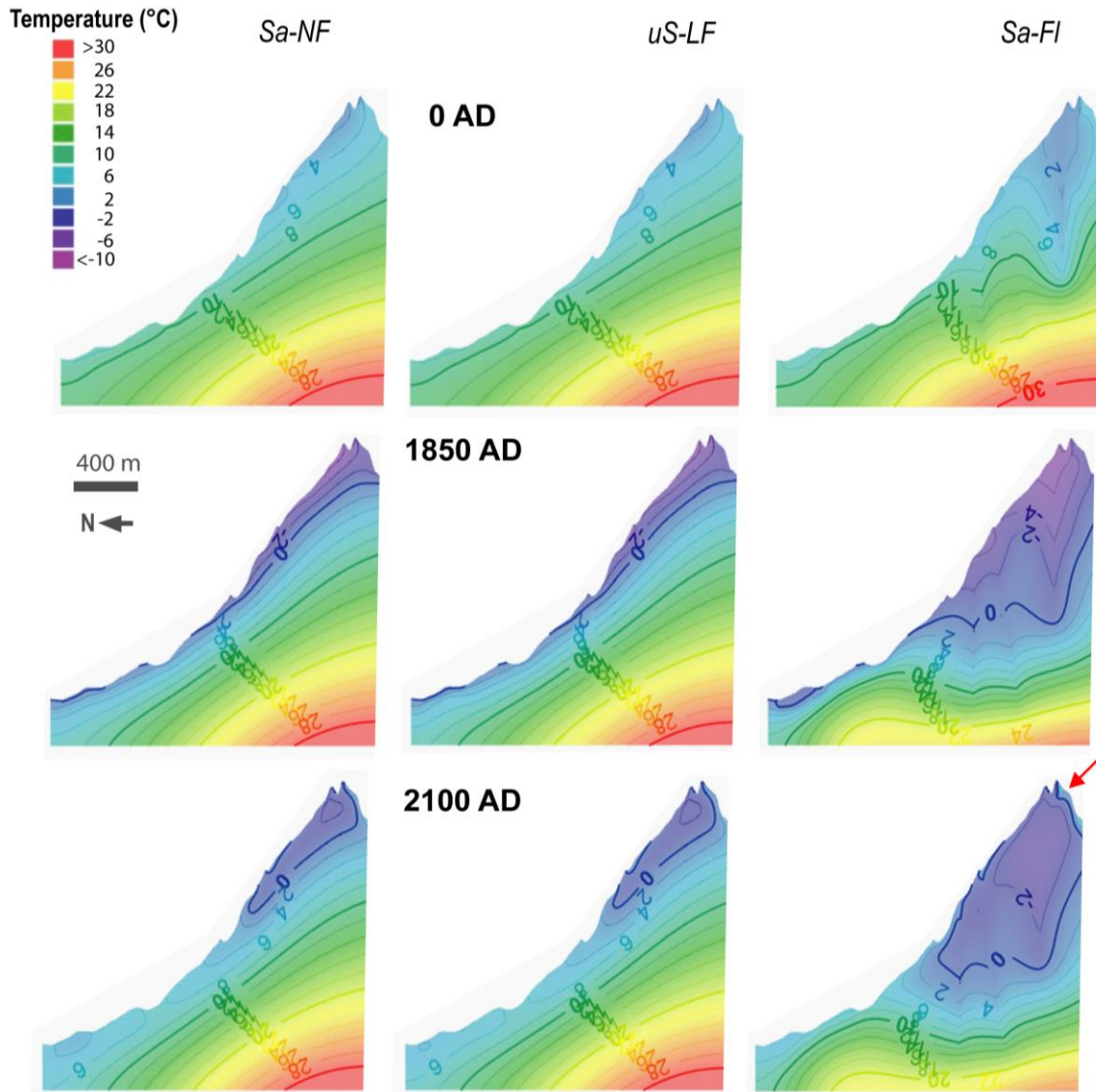
374 **4 Results**

375 4.1 Temperature fields

376 Figure 3 shows the temperature fields evolution after initialization (0 AD), freezing (1550
377 AD) and thawing (1850 AD) for the *Sa-NF*, *Sa-Fl* and *uS-LF* and Figure 4 focuses on specific
378 timing and space scales of cases with enhanced fluid flows (*uS-Fl* and *Sa-Fl*) to better show the
379 water flow effect when thawing and freezing under variable saturation level.

380 Comparison of *Sa-NF* with *uS-LF* (Fig. 3) shows that minor water flows have no effect on
381 temperature fields distribution whatever the saturation level. Without sufficient fluid flows
382 resulting of sufficient hydraulic gradient or saturation, heat transfer is purely conductive
383 (Ingebritsen and Sanford, 1998). With a bedrock porosity of 5%, the thermal conductivity and the
384 heat capacity of the rock are the most influent parameters. Thus, temperature distribution mainly

385 results of the topoclimatic control on surface temperature (air temperature lapse-rate and sun-
 386 exposure) such as described in previous studies (e.g. Noetzli et al., 2007).

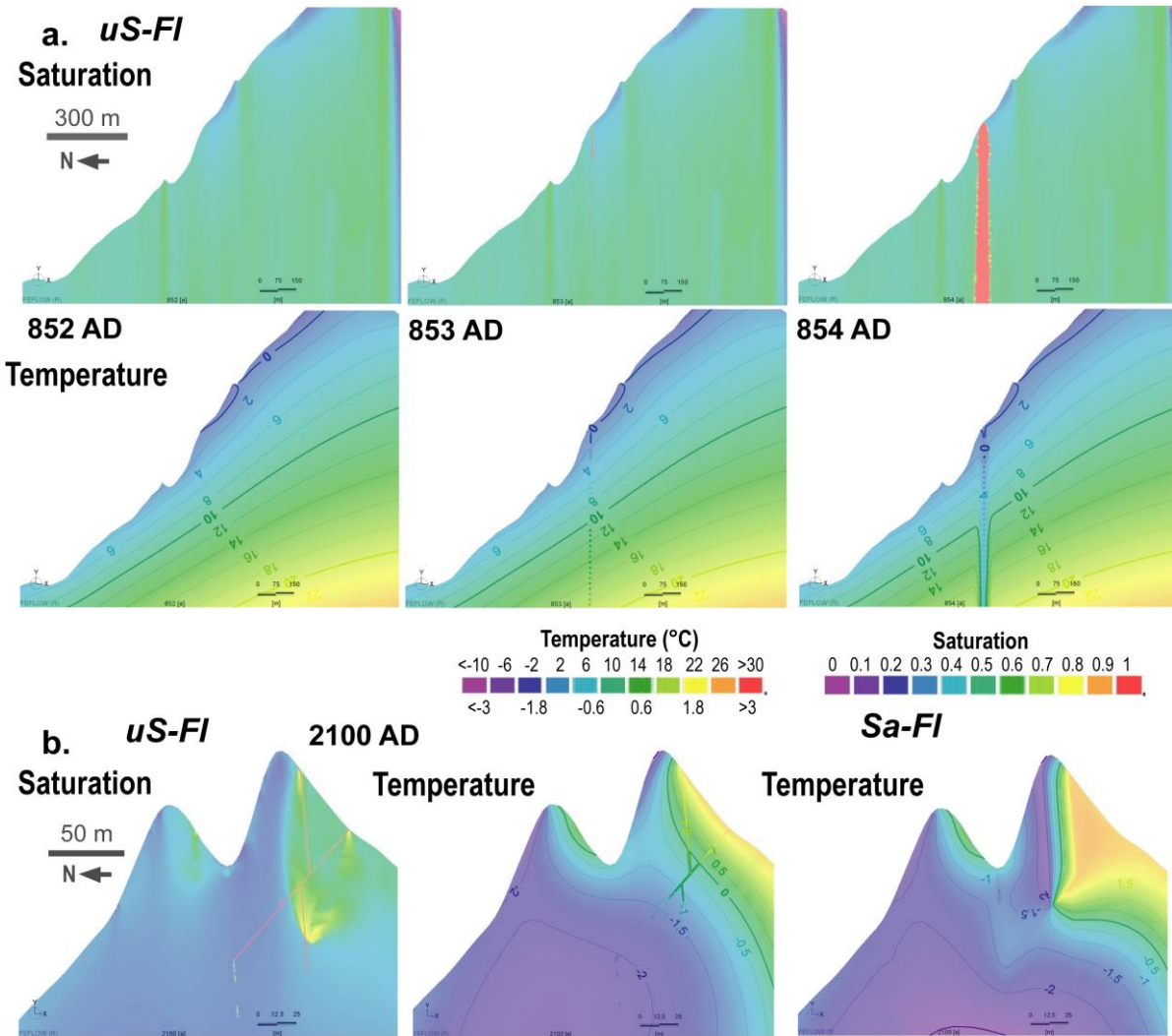


387

388 **Figure 3.** Temperature fields distribution in the mountain flank for steady and transient states
 389 when freezing and thawing. Red arrows point out area of interest reported in Figure 4b.

390 Conversely, water flows provoked by the combination of high hydraulic gradients and saturated
 391 conditions have major effects on bedrock temperature. For *Sa-Fl*, temperature fields are affected
 392 in the whole domain and isotherm distortion is visible at the three timings (0, 1850 and 2100 AD)
 393 along the main fractures (Fig. 3). This corresponds to water flow pathways (see Fig. S2 for Darcy
 394 fluxes: they are several order of magnitude greater in the fractures than in the rock matrix) which
 395 drives the cold water from high-elevated areas to lower and warmer areas throughout these main
 396 flow paths (Ingebritsen and Sanford, 1998; Gallino et al., 2009; Thiébaud et al. 2010; Dzikowski
 397 et al., 2016). Cold corridors thus develop along the major fractures and gradually affects the

398 surrounding rock matrix. These water flows stretch the permafrost body deeper than under no-flow
 399 or very low flow conditions such as for *uS-LF*. But saturation is a key parameter as exemplified
 400 by the *uS-FI* case (Fig. 4a): the freezing only starts in 850 AD at this low-elevated fracture and the
 401 steady increase of the hydraulic head provoked at this time until saturation is followed by a
 402 temperature drop along the fracture down to more than 100 m depth. In the meantime, the
 403 saturation propagates in the rock matrix from the fracture, which extends the cold area around the
 404 fracture.



405
 406 **Figure 4.** Thawing and freezing patterns under various saturation levels. **a.** Progressive fracture
 407 saturation and cooling under freezing conditions for the *uS-FI* case. **b.** Relation between
 408 saturation and thawing patterns (saturation for *Sa-FI* is not displayed as it is fully saturated
 409 everywhere).

410 In the same way, thawing is also highly constrained by the saturation level but also by the
 411 temperature and fracture connectivity. When saturated or frozen up to the surface, thawing occurs
 412 through heat conduction only and the bedrock thaws before ice-filled fractures due to enhanced
 413 latent heat effects in the latter. This explains that permafrost warming and thawing between 1850

414 and 2100 AD is less along the deepest fractures for the case *Sa-Fl* (Fig. 3). However, when the ice
415 content starts to melt in the fracture, and is further enhanced by water infiltration, water flow starts
416 and accelerates permafrost degradation (see red arrow on Fig. 3 for case *Sa-Fl* and Fig. 4b). At the
417 end of the thawing period (2100 AD), the isotherm 0°C has a square shape on the south-exposed
418 face of the Central Pillar, that is constrained by the thawing fractures geometries. Figure 4b shows
419 that interconnected thawing fractures provoke much faster thawing in the rock matrix also than
420 through heat conduction from the surface only. In this way, thawing is bottom-up (see Figure S3
421 for a more detailed chronicle). If the surrounding rock matrix is not saturated such as in the case
422 of *uS-Fl*, thawing corridors develop in the saturated fractures only. Under saturated conditions (*Sa-*
423 *Fl*), temperature gradients are exacerbated due to still frozen fractures on the north-exposed face
424 and enhanced thawing around interconnected thawed fractures.

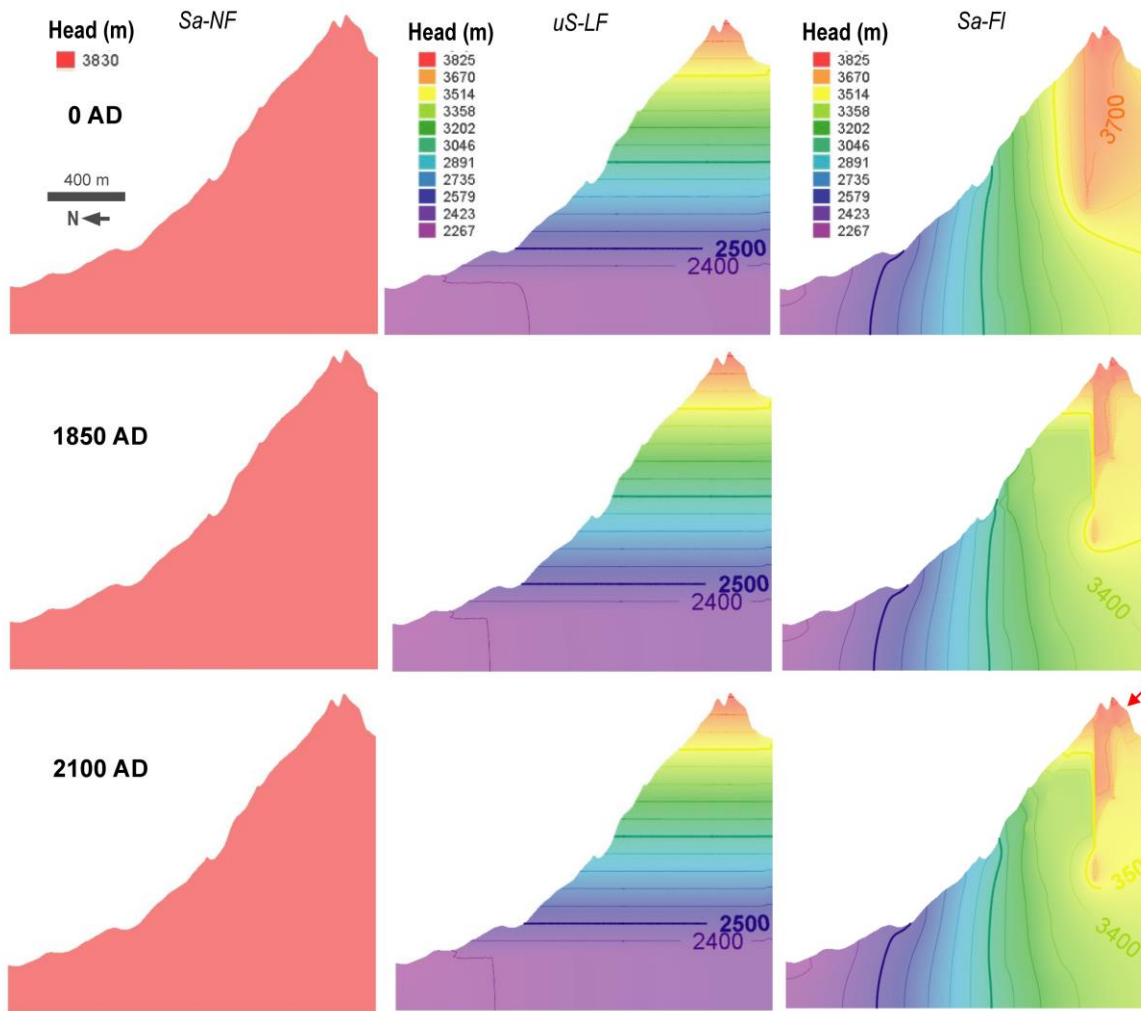
425 4.2. Hydraulic behavior

426 Figure 5 shows the hydraulic heads after initialization (0 AD), freezing (1550 AD) and
427 thawing (1850 AD) for the *Sa-NF*, *Sa-Fl* and *uS-LF*. While temperature fields are the same for *Sa-*
428 *NF* and *uS-LF* at every time step displayed in Figure 3, hydraulic head are fairly different between
429 the 2 cases and are merely the result of the chosen model parameterization to reduce water flow as
430 much as possible. However, conversely to temperature fields, they remain identical through time
431 for these 2 cases because water flows are not sufficient to provoke any perceptible head distortions.
432 Minor perturbations appear during freezing or thawing but they are not visible at the given scale.
433 For *uS-LF*, the hydraulic head roughly follows the elevation (with a 4-5 m negative offset related
434 to the unsaturation) and equipotential lines are thus horizontal. However, for *Sa-Fl* the
435 equipotential lines have a vertical to subvertical pattern that is related to the direction of the main
436 water flows along major vertical fractures and laterally from the fractures to the rock matrix.

437 Conversely to the 2 cases with no or little water flows, hydraulic heads are significantly affected
438 by freezing and thawing for *Sa-Fl*. Fractures freeze top-down and steadily form frozen barriers
439 that reduce water flows towards the rock matrix and thus lowers the hydraulic heads in the freezing
440 rock portions. Because a fraction of liquid water is maintained in ice, and because of the permanent
441 saturated conditions, water flows still occur in the main fractures, maintaining significant hydraulic
442 head values which equipotential lines are shaped by fracture geometry. When only the fracture is
443 saturated, head distortion is, similarly to temperature, only visible along the fracture (Fig. 6a).

444 .

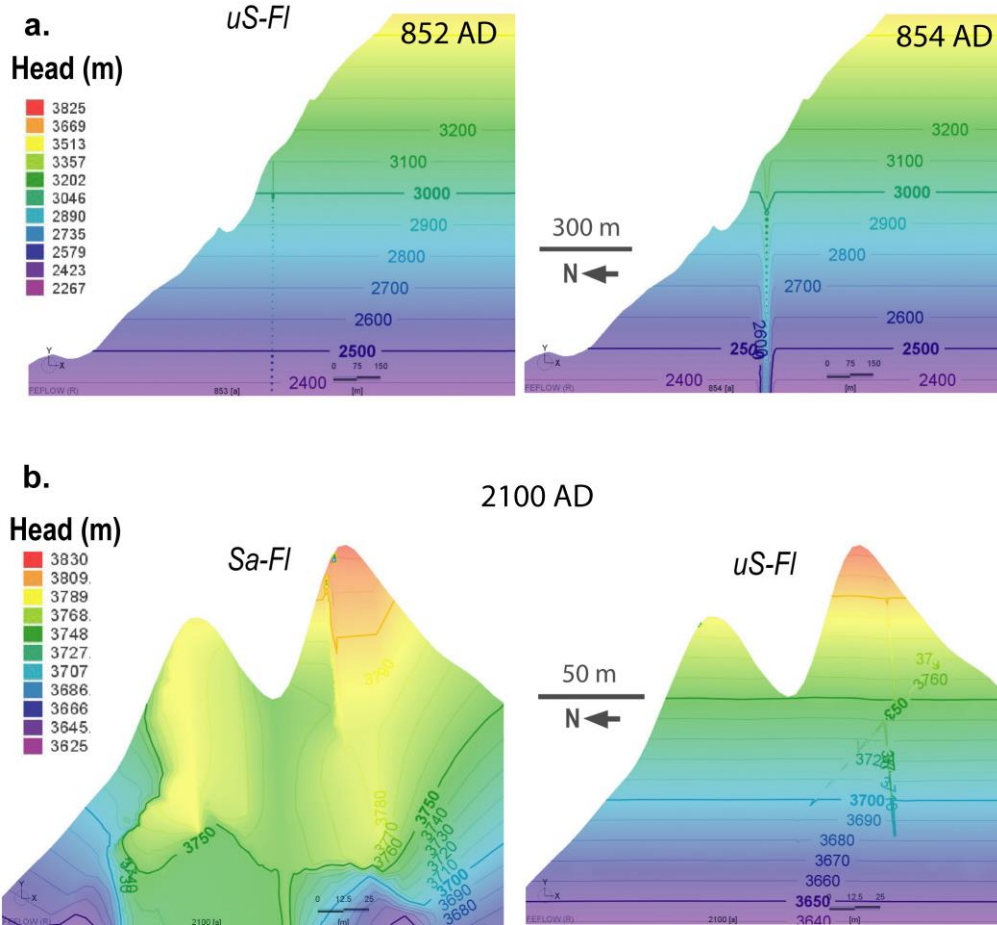
445



446
 447 **Figure 5.** Hydraulic heads distribution in the mountain flank for steady and transient states when
 448 freezing and thawing. Red arrows point out area of interest reported in Figure 6b.

449 When thawing occurs, the head increases again in thawing areas (red arrow on Fig. 5 and Fig. 6b).
 450 The equipotential lines shape moves from subhorizontal to subvertical as water flows are driven
 451 laterally from ice-blocked fractures to the thawed rock mass (see S3 for a more detailed chronicle).
 452 This observation reveals the behavior of a perched aquifer over an impermeable permafrost body.
 453 This lead to well higher hydraulic head values than in unsaturated or partially (fractures only)
 454 saturated conditions (Fig. 6b). Under unsaturated conditions, hydraulic head distortions are
 455 insignificant and become clearly perceptible along the fracture only and its surroundings when
 456 saturation of those occurs.

457 Therefore, the hydraulic behavior of rockwall permafrost follows similar principles as its thermal
 458 behavior: it is largely controlled by water flows in saturated areas. These local (fractures only) to
 459 more generalized head increase (propagation in the rock matrix) have strong implications for
 460 pressure distribution (see 5.2).



462

463 **Figure 6.** Hydraulic head patterns under various saturation levels. **a.** Progressive head increasing
 464 with saturation of the rock fracture and the rock mass when freezing (corresponding temperature
 465 and saturation are given in Fig. 4). **b.** Relation between saturation and hydraulic head patterns
 466 when thawing.

467 **5 Discussion**

468 **5.1 Limits and future developments**

469 Our study bears some conceptual and numerical limits which point out future research
 470 developments. Numerically, the large model domain discretized by a high amount of elements
 471 and nodes results in high CPU-time consumption, especially in unsaturated conditions (> 1
 472 month for some simulations). This domain size was chosen because we adopted a
 473 hydrogeological approach for which considering an entire mountain flank was necessary.
 474 However, the use of such hydrothermal model for geomorphological purposes (understanding
 475 bedrock failure for example) could be based on a reduced model domain. In addition, repeated
 476 numerical instabilities occurred when running simulations for unsaturated conditions. This is

477 most likely due to the non-linearity of the equations of the unsaturated flow that are usually not
478 used for so high unsaturated zones as well as in freezing and thawing bedrock.

479 Such numerical limitations could challenge the consideration of shorter-term freeze and thaw
480 cycles such as the seasonal ones which are relevant for geomorphological concerns. A reduction
481 of the model domain should be considered for such purposes. This would involve to setup a
482 perched water table. The scientific literature mentioning water table level in crystalline massif
483 like in our study is really poor. Most knowledge comes from random measurements and
484 assumptions. The study from Masset & Loew (2010) reports a height of up to 400 m of
485 unsaturated zone in a crystalline massif of the European Alps. But its value cannot be
486 generalized and should be apprehended in light of local topographical settings. The Aiguille du
487 Midi, that is the western margin of the Aiguilles de Chamonix, is located in a unique
488 topographical context characterized by > 1000 m high granitic rockwalls overhanging the valley
489 shoulder. Permanent streams reflecting a natural water table are only found on this shoulder, and
490 our water table settings does not seem irrelevant. However, the absence of permanent streams in
491 the rockwalls may rather reflect a permafrost-related aquiclude than unsaturated rockwalls.
492 During the past hot summers such as 2015 or 2019, numerous springs intermittently appeared in
493 the rockwalls of the Mont Blanc massif and are assumed to be linked to temporary drainage of
494 bedrock fractures above perched aquiclude. Observation and investigation of these temporary
495 offsprings should help in conceptualizing perched water tables in rockwall permafrost that our
496 simplified model represented on the thawing sunny faces (red arrows in Fig. 3 and 5, Fig. 4 and
497 6). These observations do not allow to understand the detailed functioning of perched aquiclude
498 but at least showed that our modeling approach allows to simulate them which might then be
499 relevant to understand rockwall permafrost mechanics (Fischer et al., 2010; Stoll et al., 2020).
500 The water table defined in the frame of our study, was however relevant to address long-term
501 hydraulic and thermal behavior of high-mountain rockwalls with implications for questioning
502 hydrothermal regime in alpine environments.

503 In the same way, water inputs were determined according to our research objectives which were
504 to investigate the effect of saturated versus unsaturated conditions. We thus adapted the
505 hydraulic head in order to represent those different saturation conditions, but not to account for
506 realistic water input. Realistic water input could be considered in future developments to address
507 the role of snow melt infiltration or rain events on the permafrost dynamics (see 5.2).

508 Another limit is related to the chosen fracture set that was composed of few but widely open
509 fractures overlooking the role of fracture density. We thus had to ensure bedrock permeability by
510 accounting for a rather high porosity indirectly representing fracture density and weathered zones
511 such as in former thermal modeling approach (Magnin et al., 2017; Noetzli et al., 2007). At a
512 reduced spatial scale, it would be relevant to implement a more realistic rock matrix (less porous)
513 combined with more realistic set of fractures (a dense network of thin fractures in addition to the
514 most important ones) as granitic bedrock permeability is mostly dependent on the fracture
515 network characteristics (Hsieh, 1998; Renshaw, 1996; 1997).

516 It is also noteworthy to point out that Feflow simulates saturation from top to bottom in the
517 fractures, which is not necessarily the case in natural environments. Improvements in the
518 modeling approach in order to setup more realistic hydrological processes representing water

519 percolation along the fracture and its accumulation at the ice cement as described by Hasler et al.
520 (2011a) should be considered for more detailed investigations.

521 Finally, other improvements could be performed by considering air convection in unsaturated
522 fractures which likely cools the bedrock (Hasler et al., 2011b), a 3D geometry which would
523 better account for high hydraulic and thermal gradients, but would also challenge the calculation
524 capacity, or turbulent heat fluxes that may locally affect cleft ice erosion and heat transfers
525 (Hasler et al., 2011a).

526 5.2 Results implications

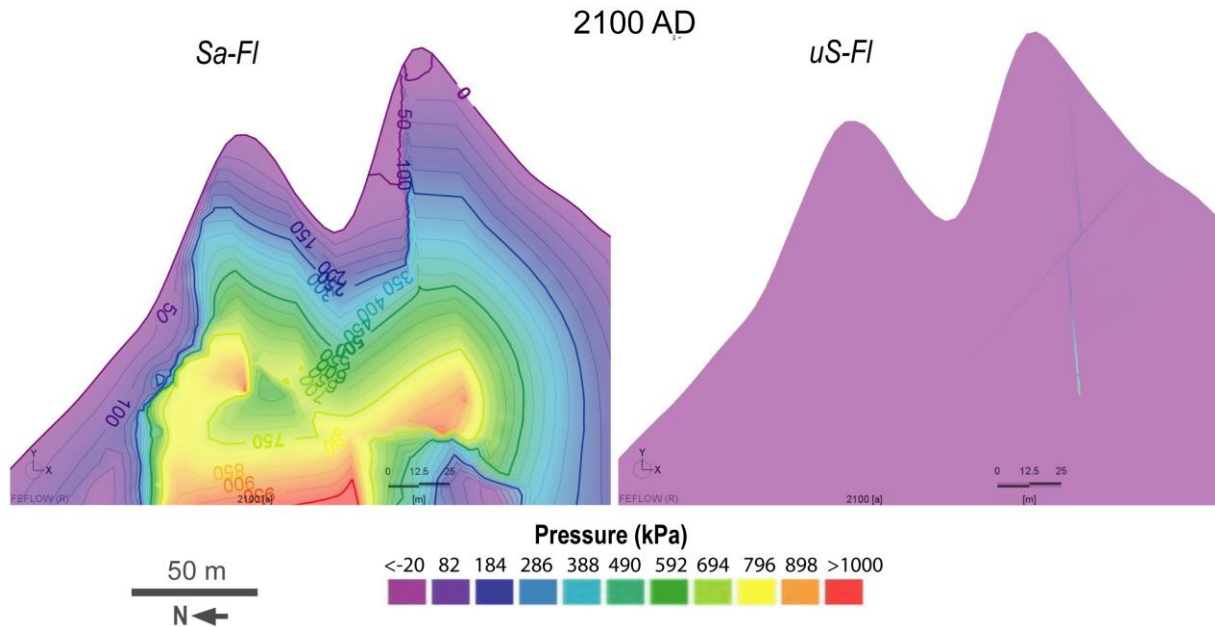
527 The results of this study bear strong implications for understanding permafrost dynamics
528 and its response to climate change, alpine morphodynamics, alpine hydrogeology and hazards
529 assessment.

530 Permafrost temperature and active layer thickness are recognized as Essential Climate Variables
531 by the Global Climate Observing System. Understanding their response to climate change is thus
532 of global concern to apprehend the environmental impact of atmospheric warming. Simulations
533 run with water flows show that rockwall permafrost dynamics are more complex than previously
534 thought. Water stretches the permafrost body deeper than under purely conductive transfer,
535 meaning that it may be present in areas not predicted by thermal models accounting for heat
536 conduction only. Because of ice bodies occupying large fractures, thawing may be either delayed
537 or enhanced, while in the nearby rock mass, thawing corridors and area may develop in and from
538 fractures. Interconnected fractures may enhance thawing at depth, in areas delineated by those
539 fractures and promote bottom-up permafrost degradation. Thawing corridors along fractures
540 were already suggested by experimental or geophysical investigations (Hasler et al., 2011a;
541 Keuschnig et al., 2017; Krautblatter & Hauck, 2007), and the numerical approach proposed in
542 this study could be essential to conduct systematic investigations on such processes in order to
543 quantitatively scale the effects of fracture aperture, dipping, density, connectivity and
544 temperature, as well as the effects of water inflow characteristics (amount, temperature, timing of
545 infiltration) on permafrost dynamics. In addition, coupled thermal and hydrological models could
546 help investigating the effect of snow or glaciers melting water and rainfalls infiltrations in the
547 rock mass, some processes that have been so far overlooked in the conceptualization of rockwall
548 permafrost evolution.

549 Besides understanding thermal dynamics, our results may provide relevant knowledge to
550 understand cold water anomalies found during deep geotechnical work such as tunneling
551 (Maréchal et al., 1999; Maréchal, 1998; Mommessin, 2015). Such anomalies are attributed to
552 cold water infiltrating from mountain top which are in some cases glaciated and our findings
553 further question the role of permafrost in alpine hydrothermal processes.

554 Thawing corridors and strongly variable hydraulic heads further bear implication for
555 understanding rock failure hazards in periglacial environments. The warming or thawing of ice-
556 filled fractures altering their shear resistance or provoking a loss of rock-ice contact are currently
557 recognized as the main permafrost-related triggering factors of rockfalls (Davies et al., 2001;
558 Krautblatter et al., 2013; Mamot et al., 2018; Matsuoka & Murton, 2008). But high hydrostatic
559 pressures related to perched aquifer is also thought to play a role in rockwall destabilization

560 (Fischer et al., 2010; Walter et al., 2020). However, those fluid pressures are the less understood
 561 and recent developments in mechanical modeling have shown that hydrostatic pressure of 0.1
 562 MPa over an impermeable permafrost body could trigger rock slope failure (Stoll et al., 2020).
 563 Our results show that such critical level of hydrostatic pressure may be reached with hundreds of
 564 kPa associated to thawed bedrock, either locally in water saturated fractures or in the entire
 565 thawed rock matrix (Fig. 7), hinting at favorable conditions for rockwall destabilization.



566

567 **Figure 7.** Pressure fields in 2100 AD for different saturation levels.

568 Finally, as already concluded by Hasler et al. (2011a), our study confirms that thawing corridors
 569 but also thawing areas delineated by thawing fractures may explain rockfall triggering in cold
 570 permafrost areas or before the maximum active layer depth (Gruber et al., 2004; Luethi et al.,
 571 2015). It also shows that frozen fractures may be found in thawed bedrock.

572 **5 Conclusions and outlooks**

573 In this study, we propose the very first fully coupled hydrological and thermal models applied to
 574 rockwall permafrost. As a preliminary approach, we adopted a hydrogeological
 575 conceptualization and considered a mountain flank as model domain. Our model set up resulted
 576 in very high CPU-time but this study could serve as a baseline for future modeling developments
 577 focusing on a reduced model domain. We run four simulations accounting for variable saturation
 578 levels and water flows. We draw the following conclusions:

- 579 - Water flows have major effects on the aggradation and degradation of alpine rockwall
 580 permafrost. They drive the cold from top to bottom when permafrost forms resulting in a
 581 deeper permafrost body than under heat conduction only. Ice-cement in bedrock fractures

582 first delays permafrost degradation compared to surrounding bedrock but then accelerates
583 it as soon as ice starts to melt.

584 - Thawing fractures act as thawing corridors accelerating permafrost degradation at depth.
585 When surrounding bedrock is saturated, interconnected thawing fractures create thawing
586 wedges provoking bottom-up permafrost degradation.

587 - Fractures exacerbate thermal gradients due to enhanced frost in still ice-cemented
588 fractures that can be in the vicinity of thawing corridors or wedges (exacerbation of the
589 north-south contrasts for example). In this respect, frozen fractures may subsist in thawed
590 bedrock and thawed areas may form in frozen bedrock.

591 - Bedrock permeability is dependent on fracturing as water flows mostly occur along
592 fractures and remain minimal in the bedrock matrix, notably in unsaturated conditions. In
593 saturated conditions, water first flows in the fracture and then towards the bedrock
594 matrix.

595 - Water flows provoke very unequal hydraulic head distribution and related fluid pressures.
596 In a saturated and fractured medium, equipotential lines follow fracture shapes.

597 - Thawing results in high hydraulic head and fluid pressure than can reach several
598 hundreds of kPa over or within the permeable permafrost body.

599 - Thermal and hydraulic patterns of rockwall permafrost are strongly affected by fluid
600 flows under saturated conditions only. Unsaturated conditions substantially minor their
601 effect. Knowing the level of saturation of bedrock fractures is thus highly relevant to
602 understand permafrost evolution and rockwall destabilization patterns.

603 This study bears strong implications for understanding permafrost response to climate change,
604 rock slope failure, related hazards and geomorphic processes, as well as alpine hydrogeology.
605 Future developments would benefit of a reduced model domain accounting for a perched water
606 table, realistic fracture sets and water inflows to better characterize water flow effects.

607 **Acknowledgments and Data availability**

608 This study is funded by the ANR-19-CE01-0018 WISPER project. Model input, forcing and
609 output data are available on the *Zenodo* data repository: <https://doi.org/10.5281/zenodo.3958847>

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611

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