


Rock slope temperature evolution and micrometer-scale deformation at a retreating glacier margin

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1 **Rock slope temperature evolution and micrometer-scale de-** 2 **formation at a retreating glacier margin**

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6 Key points

- 7 • We monitor and model the subsurface temperature evolution and deformation in a rock slope during
8 deglaciation.
- 9 • We observe a conduction dominated transient subsurface temperature regime adjusting to a new
10 surface temperature following ice retreat.
- 11 • We document reversible deformation and irreversible rock mass damage and identify important
12 drivers causing these deformations.

13 Abstract

14 In deglaciating environments, rock mass weakening and potential formation of rock slope instabilities is
 15 driven by long-term and seasonal changes in thermal- and hydraulic boundary conditions, combined with
 16 unloading due to ice melting. However, in-situ observations are rare. In this study, we present new monitor-
 17 ing data from three highly instrumented boreholes, and numerical simulations to investigate rock slope tem-
 18 perature evolution and micrometer-scale deformation during deglaciation. Our results show that the subsur-
 19 face temperatures are adjusting to a new, warmer surface temperature following ice retreat. Heat conduction
 20 is identified as the dominant heat transfer process at sites with intact rock. Observed nonconductive pro-
 21 cesses are related to groundwater exchange with cold subglacial water, snowmelt infiltration, or creek water
 22 infiltration. Our strain data shows that annual surface temperature cycles cause thermoelastic deformation
 23 that dominate the strain signals in the shallow thermally active layer at our stable rock slope locations. At
 24 deeper sensors, reversible strain signals correlating with pore pressure fluctuations dominate. Irreversible
 25 deformation, which we relate with progressive rock mass damage, occurs as short-term (hours to weeks)
 26 strain events and as slower, continuous strain trends. The majority of the short-term irreversible strain events
 27 coincides with precipitation events or pore pressure changes. Longer-term trends in the strain time series
 28 and a minority of short-term strain events cannot directly be related to any of the investigated drivers, but
 29 changes of glacial loading and fatigue processes are considered as potential causes.

30 Keywords

31 Subsurface temperature evolution; progressive rock mass damage; paraglacial rock slope evolution, bore-
 32 hole monitoring; rock slope deformation

33 1. Introduction

34 Retreating and advancing valley glaciers in alpine regions can induce both reversible and irreversible de-
35 formations in adjacent valley flanks. Several mechanisms can cause these deformations, however, glacial
36 influence on the thermal and hydraulic boundary conditions of adjacent slopes has been identified as a
37 major driver. In particular, as glaciers retreat, the stable thermal boundary conditions below glaciers are
38 perturbed, which results in a “thermal shock” and long term cyclic thermal loading, that can induce long term
39 rock mass damage (Grämiger et al., 2018; Grämiger et al., 2020). The present work is focused on field
40 investigations of the transient thermal regime induced by glacier retreat in adjacent valley slopes, and the
41 related subsurface deformations. This data set is key to the understanding of heat transfer mechanisms in
42 the subsurface, the timescales over which valley flanks adjust to the new thermal boundary conditions fol-
43 lowing deglaciation, as well as the magnitudes of thermo-mechanical stresses, which can lead to long term
44 rock mass damage. Investigating these phenomena requires high resolution measurements of subsurface
45 temperatures and deformation, which have never been recorded in alpine regions that feature temperate
46 valley glaciers. Additionally, thermo-mechanical deformation signals can be overprinted e.g. by hydrome-
47 chanical deformation (cf. Grämiger et al., 2020), and disentangling the two requires an understanding of
48 slope hydrology, as well as measurements of pore pressure and precipitation events.

49 Previous studies have used borehole temperature recordings to analyze the subsurface temperature regime
50 and reconstruct past climate conditions (Bodri and Čermák, 1995; Bodri and Čermák, 1999; Pollack, 1993;
51 Pollack et al., 1998; Pollack and Huang, 2000), as well as study permafrost processes (e.g., Gruber et al.,
52 2004; Haberkorn et al., 2015; Schneider et al., 2012) and drivers of active rock slope instabilities (e.g.,
53 Gischig et al., 2011a; b; Moore et al., 2011). However, to the authors knowledge, subsurface temperature
54 investigations in alpine environments undergoing glacial retreat below permafrost elevations have not been
55 monitored in the past.

56 When considering the thermal regime of a rock slope, the glacier – rock slope interaction is controlled by
57 both thermal and hydraulic conditions at the ice-rock interface. Bedrock in contact with temperate ice stays
58 at the pressure melting point of the ice, which is around 0° C (Wegmann et al., 1998). When ice retreats,

59 the previously covered bedrock surface is newly exposed to ambient temperature variations and solar radi-
 60 ation. These warmer temperatures then propagate to depth through a combination of conduction and ad-
 61 vection.

62 A conduction dominated subsurface temperature regime, which is expected in intact crystalline bedrock, is
 63 controlled by: 1) the surface temperature, which is subject to variability at daily to millennial timescales, 2)
 64 an outward flow of heat from the earth's interior, which can be considered constant at the relevant time-
 65 scales, 3) the thermal diffusivity of the medium, 4) latent heat effects from melting ice or freezing water in
 66 pores/cracks, and 5) the topography (Guéguen and Palciauskas, 1994; Pollack et al., 1998). In the case of
 67 homogenous subsurface thermal properties and flat topography, a steady-state temperature profile shows
 68 a constant rate of temperature increase with depth (dependent on the thermal conductivity and the heat flow
 69 from below). Surface temperature fluctuations then propagate as thermal waves at a pace governed by the
 70 thermal diffusivity, with amplitudes diminishing exponentially with depth (Pollack and Huang, 2000).

71 In reality, rock masses contain discontinuities that allow nonconductive heat transfer processes (e.g. advec-
 72 tion of air or water) and heterogeneities (e.g. in the lateral thermal conductivity) that potentially disturb the
 73 idealized heat conduction dominated subsurface temperature regime (Moore et al., 2011; Rybach and
 74 Pfister, 1994; Wegmann and Gudmundsson, 1999). Of particular importance to the present work are ad-
 75 vective heat transfer processes driven by seasonal changes in rock slope hydrology in a glaciated valley,
 76 which influence the thermal regime. Transient pore pressures in such rock slopes are controlled by direct
 77 recharge to the fractured bedrock slope from snow melt in late spring and rainstorms in summer and fall, as
 78 well as by the englacial water pressure variations. Temperate glaciers carry large amounts of liquid water,
 79 and a thin water film with pressures dependent on the glacial hydrological system is assumed to be present
 80 at greater depth at the glacier – rock interface (Lappegard et al., 2006).

81 Different types of temperature-coupled processes such as frost weathering, freeze-thaw cycles, and perma-
 82 frost degradation contribute to progressive rock mass weakening in alpine regions (Deprez et al., 2020;
 83 Draebing and Krautblatter, 2019; Draebing et al., 2017; Hales and Roering, 2007). Additionally, thermo-
 84 mechanical stresses themselves can drive deformation and progressive rock mass damage in an unstable

85 rock slope (Gischig et al., 2011a; b). In the context of post- and late-glacial ice retreat, numerical investiga-
 86 tions suggest that purely thermo-mechanical effects induced by long-term temperature variations related to
 87 glacial cycles and seasonal temperature variations induce significant thermo-mechanical stresses to the
 88 valley flanks which promote progressive rock mass damage (Baroni et al., 2014; Grämiger et al., 2018).
 89 These studies have shown that, during a glacial cycle, thermo-mechanically driven deformation and rock
 90 mass damage can be substantially higher than purely mechanical loading and unloading effects from glacier
 91 retreat or advance, and that the degree of fracturing (and the criticality of the fracture system in terms of its
 92 Mohr-Coulomb strength) is an important factor controlling the amount of induced thermo-mechanical dam-
 93 age (Grämiger et al., 2018).

94 Rock mass damage can also result from subcritical cyclic stress perturbations and stress corrosion
 95 (Cerfontaine and Collin, 2018; Eppes and Keanini, 2017). This has been observed in laboratory experiments
 96 and is described by several authors (e.g., Attewell and Farmer, 1973; Brown and Hudson, 1973; Cerfontaine
 97 and Collin, 2018). Thus, low stresses can also increase the criticality of a fracture to a level exceeding the
 98 fracture toughness, which enables the fracture to fail at an apparently random, low magnitude stress per-
 99 turbation, complicating correlations with environmental drivers (Eppes and Keanini, 2017).

100 In this study we use novel borehole monitoring data and numerical simulations to investigate the transient
 101 subsurface temperature regime and related micrometer-scale deformation in a rock slope adjacent to a
 102 temperate valley glacier undergoing rapid downwasting. We first analyze the borehole temperature and
 103 strain data and perform a statistical analysis of the strain data to highlight correlations between potential
 104 drivers and deformation events. We then use the borehole temperature data to estimate the depth-depend-
 105 ent apparent thermal diffusivity at our borehole locations, which is used to parameterize continuum 2D finite
 106 element models for heat conduction in the subsurface. Comparing these modeling results with temperature
 107 monitoring data provides insights into the transient subsurface temperature regime that follows glacier re-
 108 treat, as well as the dominant mechanisms governing heat transfer at our study site (conduction vs. advec-
 109 tion). Finally, we expand our finite element models to compute the expected thermoelastic deformation
 110 resulting from annual temperature cycles. These analyses provide crucial information regarding the relative
 111 importance of thermo-mechanical deformation on the total strain recorded at our instrumented rock slope,

112 and hence its relevance for the overall observed deformation and progressive rock mass damage.

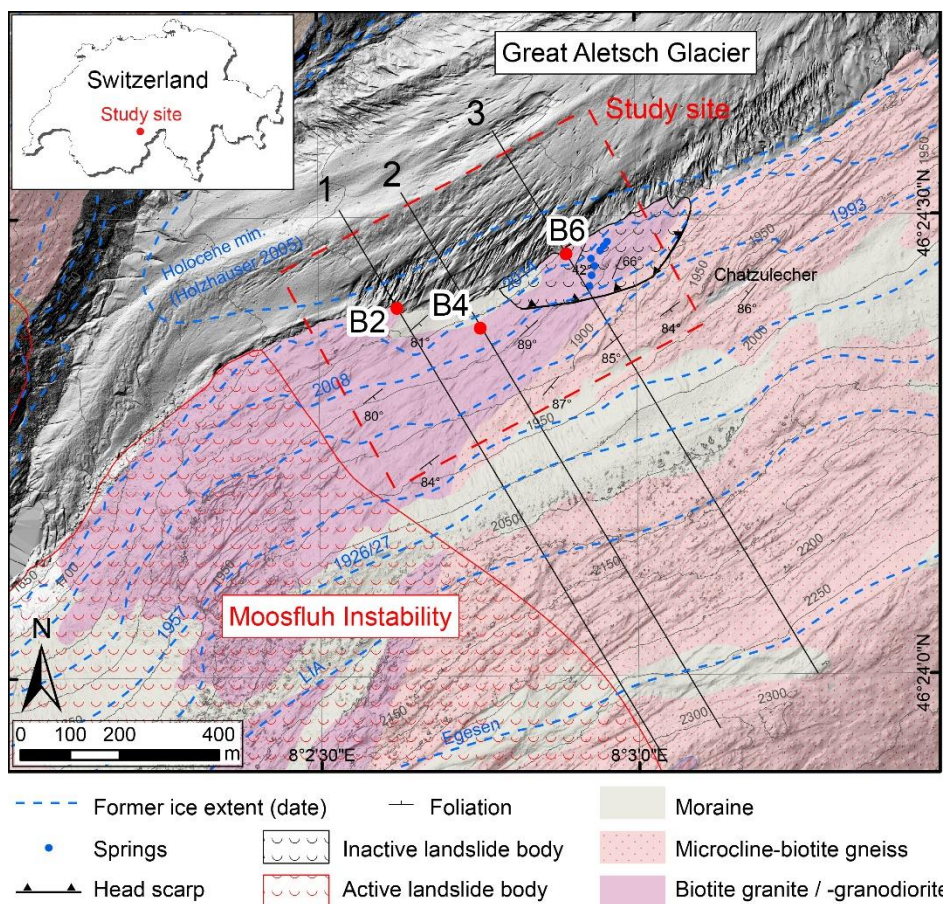
113 2. Site description and previous work

114 Our study site is located on the left (i.e. southern) valley slope adjacent to the retreating glacier tongue of
 115 the Great Aletsch Glacier (Canton Valais, Switzerland) (see Fig. 1). The bedrock consists of fractured crys-
 116 talline rocks (mainly gneisses and granites with thin schist layers) of the Aare-Massif (Steck, 2011). The rock
 117 slope is dipping NNW with an average slope angle of about 30° and the morphology is characterized by
 118 glacially smoothed bedrock surfaces with ridges and furrows. Furrows often follow weak schist layers or
 119 brittle-ductile shear zones normally oriented subparallel to the alpine foliation. This foliation strikes valley
 120 subparallel with a subvertical dip angle normally dipping into the slope. The rock mass is intersected by
 121 three persistent joint sets with a normal spacing around 1 to 3 m at the ground surface (for joint set orienta-
 122 tions refer to Fig. 4 in Hugentobler et al. (2020)).

123 A thin and patchy moraine layer, surrounding several rock outcrops, covers the lower study area. Due to the
 124 recent glacial retreat, no noteworthy vegetation is present in the study area. The Moosfluh instability, a well-
 125 documented deep seated gravitational slope deformation (DSGSD) with superimposed shallower rock-
 126 slides, is situated in the southwest of the study area x.

127 The glacial history of the Great Aletsch Glacier, the largest ice mass in the European Alps, is summarized
 128 in Holzhauser et al. (2005) and glacial extents in the vicinity of the study area have been investigated in
 129 detail by Grämiger et al. (2017). Additionally, glacial retreat from the last local maxima in 1860, the Little Ice
 130 Age (LIA), is well documented by GLAMOS (1881-2019). Since the LIA, the glacier tongue lost more than
 131 3000 m of length and over 300 m in ice thickness at the location of our study site. This leads to a mean ice
 132 height loss of about 2 m per year since LIA. However, annually acquired, high-resolution digital elevation
 133 models of the Great Aletsch Glacier area from the years 2012 until 2019 (source: Swiss Federal Office of
 134 Topography) show that the current melting rates are higher, with about 5 - 10 m ice height loss per year.

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Fig. 1. Overview map of the study area at the terminus of the Great Aletsch Glacier at the Autumn 2017 ice extent with the research borehole locations B2, B4, and B6 (modified after Hugentobler et al., 2020). The former glacial extents of years 1927 to 2014 were digitized from historical aerial photos by Glueer et al. (2020) and the Little Ice Age (LIA) and Egesen maximum extents are well constrained by lateral moraines in the map area. We extended the geological map from Steck (2011) in the area that had been glacier covered during their mapping campaign. The numbered solid black lines are the extents of the 2D profiles used for the numerical study.

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A subsurface monitoring system has been installed to investigate and quantify the effects of thermo-hydro-mechanical rock slope processes that occur along with downwasting and retreat of glacial ice (Hugentobler et al., 2020). The monitoring system is installed in three vertical, 44 to 50 m deep instrumented research boreholes of 10 cm diameter (B2, B4, and B6, see Fig. 1). The three borehole locations were selected to have varying lateral distance to the glacier margin, so that they monitor the rock slope conditions at different timings relative to deglaciation (**Fehler! Verweisquelle konnte nicht gefunden werden.**), and in slope sectors of different degree of stability and damage. Before the recent deglaciation, the borehole locations

150 were ice covered for approximately 2000 to 3000 years (Holzhauser et al., 2005).

151 The boreholes are fully grouted, and contain high resolution vertical and horizontal strain sensors, a tem-
 152 perature sensor chain, and a pore pressure sensor installed in a 1 – 2 m long sand filter at the borehole
 153 end. The setup and performance and early data of the monitoring system is described in detail in
 154 Hugentobler et al. (2020). Air temperature is measured at our study site with a 30 min interval and total
 155 precipitation data used in this study is provided by MeteoSchweiz and recorded at the weather station
 156 “Bruchji” (canton Valais) located about 6 km away from our study site.

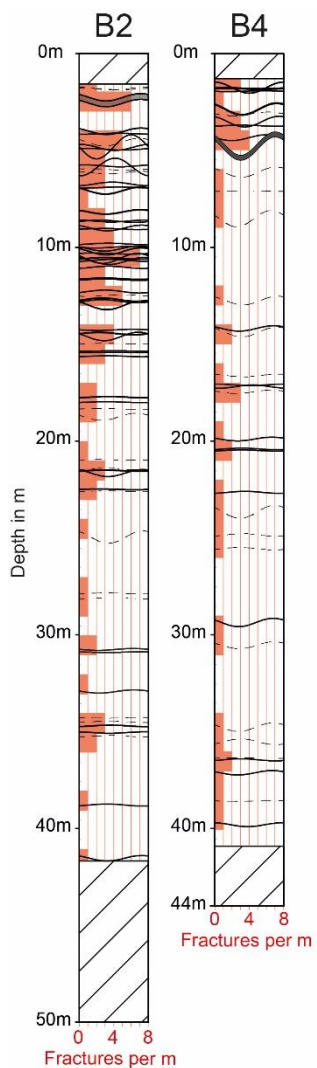
157 The rock mass and the hydrogeological properties at our study site have been characterized with surface
 158 structural mapping, rock mass classifications (at outcrops surrounding the borehole sites and from borehole
 159 optical televiewer images), borehole geophysical and hydro-geophysical investigations, as well as pumping
 160 and injection tests in the boreholes. These analyses have identified differences between the three research
 161 borehole locations (Hugentobler et al., 2020). The main findings for the individual boreholes related to rock
 162 mass quality are summarized below. As the boreholes were drilled in 2017, references to the glacier position
 163 refer to those present during that year.

164 • Borehole B2 was drilled directly at the ice margin at the end of a relatively flat plateau and only about
 165 200 m away from the active Moosfluh slope instability. Optical televiewer images of the borehole wall
 166 reveal a high fracture density with open joints often filled with fines in the uppermost 20 m (Fig. 2). Be-
 167 low this depth, fractures occur less frequently and with lower aperture. The rock mass at outcrops sur-
 168 rounding B2 was classified with a Geological Strength Index (GSI) between 55-65. According to the
 169 optical televiewer image, these values can be confirmed for the upper part of the borehole, but slightly
 170 increase with depth due to the decreasing fracture density (Fig. 2). From a borehole water injection
 171 test, the transmissivity in the lower part of the borehole (around 15 to 50 m depth) was estimated
 172 around $10^{-6} \text{ m}^2/\text{s}$ and in the uppermost, fractured part around $10^{-3} \text{ m}^2/\text{s}$.

173 • Borehole B4 is located about 50 m away from the glacier margin at a relatively evenly dipping slope.
 174 Borehole optical televiewer logs show that the rock mass is less fractured compared to B2 (Fig. 2),
 175 with mainly closed or low aperture structures. A surficial layer of about 5 m with a dense fracturing is
 176 present at this location (Fig. 2). GSI values at outcrops around B4 reveal a high-quality rock mass

177 with values between 75 and 85. Similar to B2, the rock mass quality appears to increase with depth
 178 with the observed decrease in fracture density. A pumping test in the borehole revealed a transmissiv-
 179 ity of about 10^{-6} m²/s between 9 and 44 m depth and higher assumed values in the uppermost section.
 180 Fluid electric conductivity logs show a major transmissive structure at the depth of 29.4 m and minor
 181 transmissive structures at 19.5 m, 25 m, and 39.8 m.

182 • Borehole B6 was drilled at a cliff position with about 5 m lateral distance to the glacier margin. The
 183 area around the borehole is characterized by a disturbed rock mass bounded by a prominent head
 184 scarp above (see Fig. 1). Surface structural mapping indicates that this area is likely an inactive top-
 185 pling rock slope instability, hosting important groundwater springs. We estimated GSI-values at out-
 186 crops around B6 to range between 50 and 60. Because of unstable borehole conditions, a perforated
 187 PVC casing was installed in the borehole (cf. Hugentobler et al., 2020). Therefore, only fluid electric
 188 conductivity logging and a pumping test could be conducted. These tests resulted in an estimated
 189 borehole transmissivity of around 10^{-5} m²/s with higher assumed values in the top layer and major
 190 transmissive structures at a depth of 35.5 m and 43 m. A surface creek that originates from springs
 191 located a few tens of meters above the borehole location passes a depression next to the borehole
 192 and was observed to infiltrate through fractures into the borehole during drilling works. The lower GSI-
 193 values, the higher transmissivity, and the borehole stability issues encountered while drilling reveal the
 194 lower rock mass quality at this borehole location compared to the other two.



195

196 *Fig. 2. Illustration of structures mapped on the optical televiewer image in borehole B2 and B4 (modified after*
 197 *Hugentobler et al., 2020). An optical televiewer image of B6 does not exist because no such survey could be con-*
 198 *ducted in this borehole. Solid black lines in the log indicate major discontinuities such as open joints, filled joints,*
 199 *sheared and fractured zones. Dashed lines indicate minor discontinuities (e.g., closed or healed joints). Red bars*
 200 *show the fracture count per m along the borehole.*

201 3. Methods

202 3.1. Analysis of monitoring data

203 In this paper, we present a detailed analysis of data from the subsurface temperature sensor chains, as well
 204 as from the high resolution vertical FBG strain sensors. Subsurface temperatures have been recorded since

205 October 2017 with an hourly measurement interval, and a downhole sensor spacing of 1 m. Some data
 206 gaps of a few weeks in total exist in the time series that are related to battery issues. One longer data gap
 207 of four months (April to July 2018) in the B2 time series is related to a temporary glacier re-advance that
 208 overran the borehole location and disrupted the loggers from the sensor chains.

209 We use the subsurface temperature data to empirically estimate the in-situ apparent thermal diffusivity D
 210 within the annual thermally active layer. D is defined as $D = \lambda/(\rho C_p)$, where λ is the thermal conductivity, ρ
 211 the density of the rock mass, and C_p the specific heat capacity of the rock. This was done by analytically
 212 solving the 1D heat conduction equation for measured periodic surface temperature forcing, and fitting our
 213 subsurface temperature data (cf. Koo and Song, 2008; Rajeev and Kodikara, 2016; Schneider et al., 2012).
 214 The analytical solution can then be inverted to estimate apparent thermal diffusivity based on measurements
 215 of either the amplitude decay or phase delay of the propagation of a thermal signal with depth. We apply
 216 both of these inversion procedures and use the results to interpret which borehole locations are conduction
 217 dominated, as well as to parametrize our 2-D continuum heat transfer models.

218 The FBG sensors are installed as a serial connection of ten individual sensors with base lengths of 4 to 5
 219 m and cover the full length of each borehole. The individual strain values were calculated by averaging over
 220 hourly dynamic measurements of approximately 30 s duration. In Hugentobler et al. (2020) we introduced
 221 the strain signals measured with the FBG-system and showed that the system is capable to reliably detect
 222 strains down to a magnitude below $1 \mu\epsilon$ (or $4 \mu m$). The individual FBG strain sensors show long term con-
 223 tinuous changes in strain (extension and contraction), distinct short-term strain events (within hours to days),
 224 and rapid strain steps (within one measurement cycle of 1 h) (Fig. 6, Fig. A 1, A 2, and A3). We use a
 225 principal component analysis (PCA) with the aim to investigate the continuous, longer-term strain changes
 226 and an event detection procedure for the short-term and rapid strain variations (hereafter called strain
 227 events). An explanation of these different strain signals is provided in the beginning of section 4.1.3.

228 PCA is an exploratory statistical method that can be used to synthesize information contained in our large
 229 number of datasets (e.g., Govaert, 2009). A PCA was run for all three boreholes separately based on cen-
 230 tered and standardized data matrices containing time series of the ten FBG-sensors. The resulting principal

231 components were then correlated with potential drivers, including measured surface temperature, pore pres-
 232 sure and precipitation, in order to understand the relationship between these potential drivers and the meas-
 233 ured strain.

234 To investigate potential drivers for strain events in the data and their distribution over the year, the timing of
 235 the most significant strain events in the time series were automatically retrieved with the MATLAB function
 236 `findchangepts`. This function finds the most significant changes in mean and slope of a time series based
 237 on a predefined maximum number of changing points, or a change threshold. All automatically detected
 238 strain events were then manually checked. A minority of detected events were corrected because they were
 239 artifacts related to data gaps, but the majority were confirmed. The timing of these strain events was then
 240 compared to the same potential drivers as used for the PCA. We expect that the drivers of strain events
 241 may be depth dependent, so we distinguish deep and shallow strain events. This allows us to analyze the
 242 different drivers of short term and rapid strain events, which are potentially related to rock mass damage.

243 3.2. Numerical simulation

244 We performed 2D, finite-element numerical modeling in order to understand long term (on the order of 200
 245 years) changes of the slope thermal regime caused by glacial retreat, as well as yearly changes in the
 246 thermal regime, and resulting thermo-elastic strain, driven by seasonal temperature cycles. 2D numerical
 247 simulations were conducted using the commercial finite element software COMSOL Multiphysics v 5.5. The
 248 model geometry was defined using 2D cross-sections that intersect the three borehole sites, and laterally
 249 extend from the glacier-filled valley bottom to the ridge of Moosfluh/Hohfluh (see Fig. 1). Surface topography
 250 above the current glacial ice level was defined using a high-resolution DEM from 2018 of the study area
 251 (source: Swiss Federal Office of Topography). Elevations of bedrock sections currently covered by glacial
 252 ice were interpolated using data from Rutishauser et al. (2016) in GlaThiDa_Consortium (2019).

253 Heat conduction is the only heat transfer process occurring in our continuum models. The transient subsur-
 254 face temperature field in the model was calculated by assigning boundary conditions (BC) of a constant
 255 geothermal heat flux at the bottom, a transient surface rock temperature at the top (described in more detail
 256 below), and no-flow lateral boundaries (Fig. 3). To achieve a realistic geothermal gradient of $\sim 23\text{ }^{\circ}\text{C km}^{-1}$ in
 257 depth (cf. Rybach and Pfister, 1994), the geothermal heat flux at the bottom boundary of the model was set

258 to 60 mW m⁻² which agrees to values used in literature (Wegmann et al., 1998).

259 Our models then solve the heat conduction equation defined as $\partial T / \partial t = D \nabla^2 T$ for the model domain. The
 260 thermal diffusivity D was mainly based on the empirical estimation described in the previous section, and
 261 the C_P and ρ values used in the 2D model are based on literature values, as summarized in Table 1.

262 We used a depth dependent thermal diffusivity in our models to partially account for the depth dependent
 263 structural heterogeneity present at our study site. Between 0 and 10 m depth we assigned the mean esti-
 264 mated in-situ value based on the inversion procedure detailed above. From 20 m downwards a constant
 265 literature value for intact rock thermal diffusivity was used (Table 1), and in between these two values the
 266 thermal diffusivity was linearly interpolated.

267 *Table 1. Literature parameters used for the numerical simulations.*

<i>Thermal properties</i>		
<i>Specific heat capacity C_P (at constant pressure)</i>	780 J kg ⁻¹ K ⁻¹	Wegmann (1998)
<i>Thermal conductivity λ (intact rock)</i>	2.9 W m ⁻¹ K ⁻¹	Wegmann et al. (1998)
<i>Mechanical properties</i>		
<i>Density ρ</i>	2700 kg m ⁻³	Wegmann et al. (1998)
<i>Poisson's ratio</i>	0.2	Grämiger et al. (2017)
<i>Young's modulus</i>	30 GPa	Grämiger et al. (2017)
<i>Thermal expansion α</i>	9.5 * 10 ⁻⁶ K ⁻¹	Keusen and Amiguet (1987)

268

269 The numerical modeling was performed in four steps, with the different boundary conditions visualized in
 270 Fig. 3 and the thermal boundary conditions detailed in Table 2 :

- 271 1. The steady-state subsurface temperature field was calculated for the LIA ice extent (see Fig. 9a) using
 272 0° C at the top boundary below ice elevation (BC1) and above the ice elevation an elevation depend-
 273 ent mean annual ground temperature (MAGT) function (BC2).
- 274 2. Starting from the steady-state temperature field of modeling step 1, a transient model was run to simu-
 275 late the thermal effect of the lateral ice retreat from the LIA (~1860) to the 2017 ice extent. Historical
 276 aerial photos show that the lateral ice retreat along our modeled profiles correlates linearly with the
 277 documented glacier length change of the Great Aletsch Glacier (GLAMOS, 1881-2019), which has

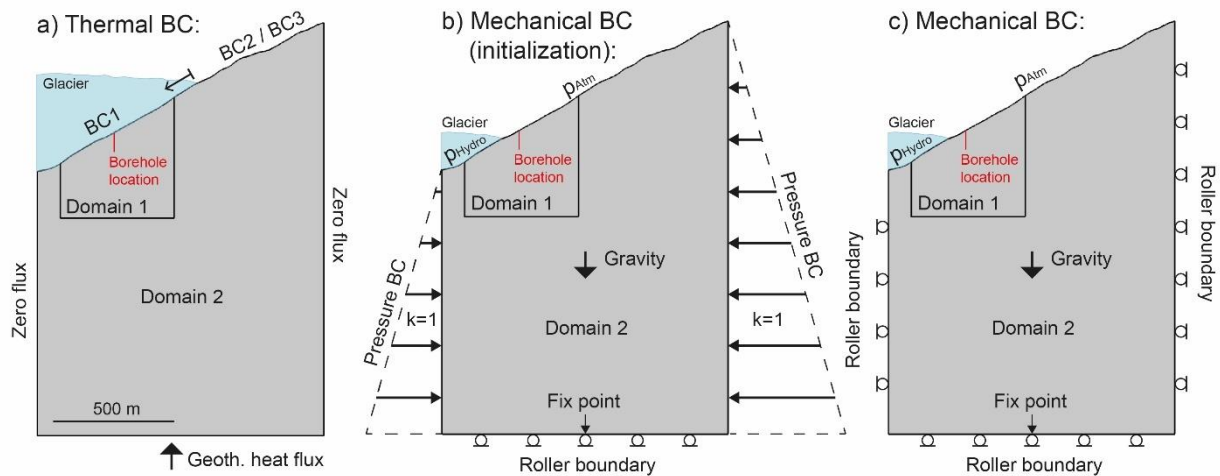
278 been monitored in detail. Therefore, the rate of glacial length change was used to calculate the lateral
 279 ice extent in each time step of the model. With this method, the modeled timing of deglaciation at the
 280 borehole locations deviate by a maximum of 2 years compared to the actual year of deglaciation. A
 281 constant temperature of 0° C (BC1) was assigned to the part of the top boundary below this lateral ice
 282 extent, and an elevation dependent MAGT function (BC2) was assigned above.

283 3. Following the simulation of ice retreat from the LIA, seasonal temperature variations were modeled
 284 using the measured temperature recorded at the uppermost sensor of borehole B4 (BC3), superim-
 285 posed on the long-term deglaciation trend since LIA. In this modeling step, the temporal resolution
 286 was increased to 1 hour in order to investigate the effects of seasonal temperature cycles on the sub-
 287 surface temperature field. Similar to modeling step 1 and 2, BC1 was assigned at the top boundary
 288 below glacial ice. To achieve the best possible fit of the initial conditions with measured temperatures
 289 in borehole B4 we tweaked the model parameters for modeling step 2 (LIA retreat) by using a con-
 290 stant intact rock thermal diffusivity of $1.4 \cdot 10^{-6} \text{ m}^2/\text{s}$ instead of the 2-layer thermal diffusivity. Then mod-
 291 eling step 2 was run for 158 years (1860 (LIA) to 2018). Following this, we applied a spin-up period of
 292 one annual cycle using again the 2-layer thermal diffusivity and BC3 top condition (Table 2) to allow
 293 the models' initial conditions to get balanced with the boundary conditions. For the simulation of the
 294 seasonal temperature cycles we then used the 2-layer thermal diffusivity introduced above.

295 *Table 2. Used thermal boundary conditions (BC) to model 2D heat conduction in our glacier adjacent rock slopes.*

BC #	Description	Value	Notes
BC1	Temperature below ice elevation	0° C	Elevation dependent gradient adapted from Grämiger et al. (2018) with absolute temperature values fitted to surface temperature monitoring data at our borehole location
BC2	MAGT above ice elevation	$\text{MAGT}(z)=12.55-0.005 \cdot z \text{ [}^\circ\text{C]}$	
BC3	GT above ice elevation	$T(t,z)=8.975-0.005 \cdot z+T_{\text{TopB4}}(t)$	

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297

298 *Fig. 3. Visualization of the model geometry and thermal (a) and mechanical (b and c) boundary conditions (BC) used*
 299 *for the numerical simulations. BC1, BC2, and BC3 are described in Table 2. Domain 1 contains a fine mesh with*
 300 *mesh sizes between 0.03 m and 18 m. Domain 2 contains a coarser mesh with individual mesh sizes between 0.5 m*
 301 *and 95 m. The mesh size is increasing with normal distance to ground surface.*

302 4. Following the thermal modeling steps 1 to 3, we computed the expected thermoelastic strains result-
 303 ing from seasonal surface temperature cycles measured at our site. We used a simplified continuum
 304 modeling approach and homogenous and isotropic elastic rock mass properties, following Grämiger et
 305 al. (2017) (see Table 1). The thermo-mechanical modeling was conducted in two steps using the
 306 same model geometry as for the thermal models. First, we initialized the stresses in the rock slope
 307 with a steady-state simulation applying mechanical forces only. Far-field stresses that are of tectonic
 308 and exhumation-induced origin were modeled in a simplified way with a horizontal to vertical stress
 309 ratio of $k=1$ (Kastrup et al., 2004) at the lateral boundaries of the model (Fig. 3b). We applied a roller
 310 boundary at the bottom and a fix point constraint at the center of this boundary line to keep the model
 311 in place. In addition, we added gravity and modeled a constant glacier load at the 2017 ice extent as a
 312 hydrostatic stress boundary (cf. Grämiger et al., 2017). After initialization, we switched the lateral me-
 313 chanical boundaries to roller boundaries (see Fig. 3c) and calculated the thermal modeling step 3 in-
 314 cluding a thermo-mechanical coupling through the thermal expansion coefficient α . We used the same
 315 thermal initial and boundary conditions as described in modeling step 3.

316 4. Results

317 4.1. Subsurface monitoring data

318 4.1.1. Temperature data

319 Fig. 4 shows the subsurface temperature evolution of the glacier adjacent rock slope from 2018 to 2020 at
 320 the three borehole locations. The graphs in the three first columns of the figure (a, b, c, e, f, g, i, j, k) show
 321 the annual temperature variation in the form of monthly mean temperature profiles and the graphs in the
 322 column to the right of the figure (d, h, l) show the evolution of the annual mean temperatures of the three
 323 monitored years. Since we started to monitor the temperature in fall 2017, our provided annual temperature
 324 cycles start in November and end in October of the following year (e.g. 2018 contains data from Nov. 2017
 325 until Oct. 2018). Fig. 4 shows that the depth of the thermally active layer in B2 and B4, is around 17 m, and
 326 annual temperature cycles smoothly diffuse into the subsurface. In contrast, in borehole B6 the annual ther-
 327 mally active layer is around 28 m deep, and the temperature profile is highly irregular. All three boreholes
 328 show weak positive and negative temperature anomalies along the profile.

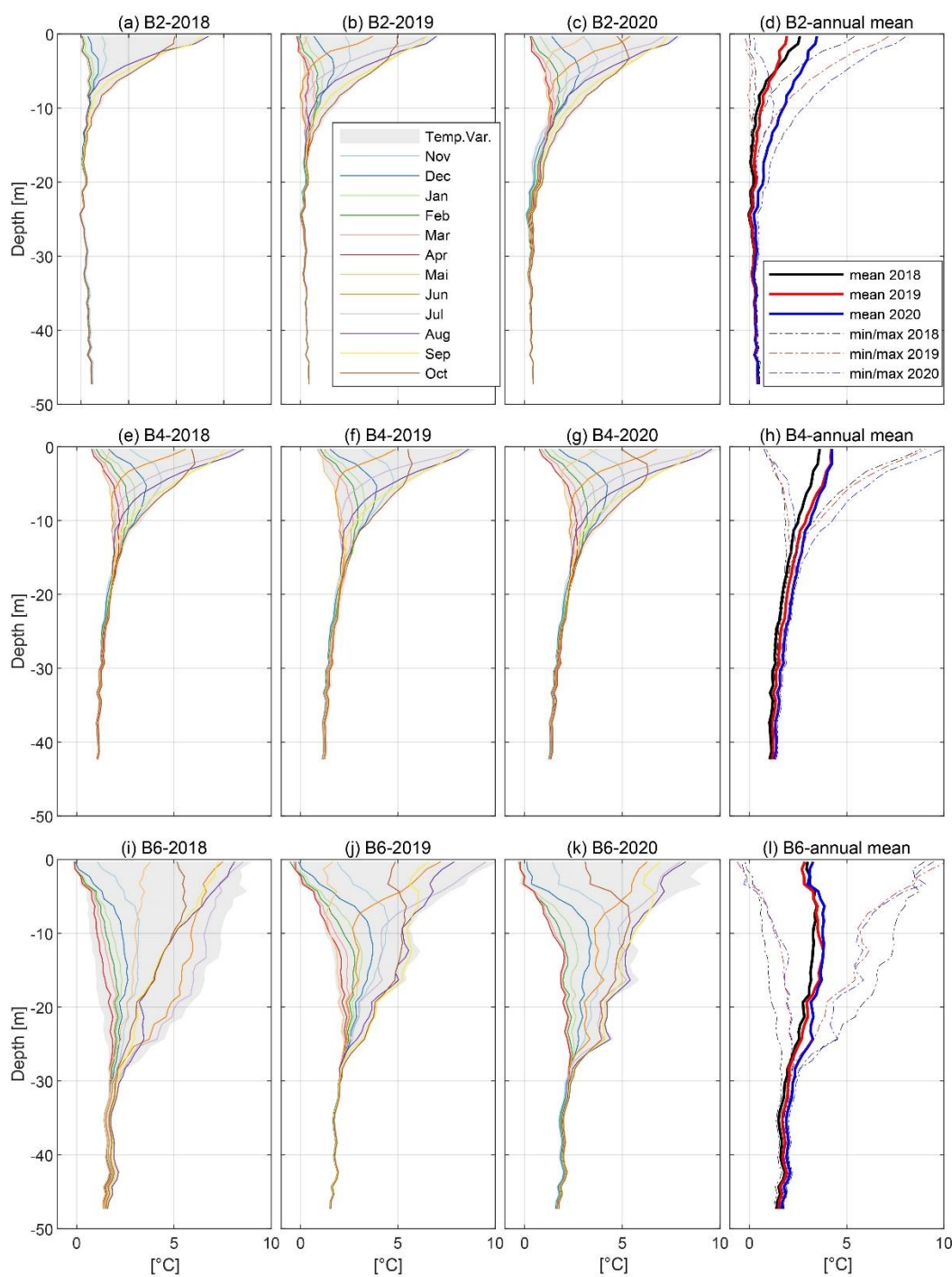
329 Borehole B2 is the closest of the three boreholes to the glacier margin, and showed the lowest mean surface
 330 temperature in 2018 and a clear warming trend in the two following years. This warming trend is likely related
 331 to the significant change of distance to the glacial ice at this borehole location (from 0 m in 2017 to about
 332 40 m in 2020). The annual variability in this borehole is around of 7 to 8 °C at the uppermost sensor. B4,
 333 which is the borehole located furthest from the glacier, shows a mean surface temperature of around 4 °C
 334 with an annual variability of around 8 to 9 °C. B6 has an intermediate distance to the glacier, and shows an
 335 annual mean surface temperature of around 3 °C and a variability of around 9 °C. All three boreholes display
 336 a general warming trend, which can be observed as a shift in positive direction of the mean annual temper-
 337 ature profiles between 2018 and 2020 (Fig. 4d, h, l). This warming trend can be overprinted by annual
 338 temperature variations in the uppermost borehole sections (e.g. boreholes B2 and B6).

339 The slopes of the mean annual temperature profiles in the three boreholes reveal that the warming front has
 340 propagated to different depths in each of the three boreholes. Starting with B2, Fig. 4d shows a negative
 341 temperature gradient (i.e., cooling downwards) to a depth of about 17 m (in 2018 and 2019). Between 17

342 m and 25 m, the temperature during these years were anomalously low, with measured values around 0 °C
 343 in years 2018 and 2019, and showing a clear increase in 2020. Below 25 m the temperature increases with
 344 depth with a gradient of about 0.2 °C / 10 m (i.e., similar to the assumed geothermal gradient of 23°/km in
 345 the area (Rybach and Pfister, 1994)).

346 In B4, the temperature profile (Fig. 4h) shows a cooling downward gradient along the whole length of the
 347 borehole, although this gradient reduces with depth. This reduction indicates that the trend may switch to
 348 warming downward below the borehole depth, similar to the trend observed in B2 below 25 m.

349 Borehole B6 shows a general temperature decrease with depth below a depth of 10 m, however, with a
 350 more irregular shape of the temperature profile showing deviations in positive and negative directions. It is
 351 important to mention that the upper 27 m of borehole B6 were not grouted until August 2018, because
 352 significant quantities of grout were lost through transmissive structures during the initial campaign in October
 353 2017 (cf. Hugentobler et al., 2020). The open borehole and the exothermal reaction during the second
 354 grouting campaign in August 2018 influenced the B6 temperature measurements in the upper 27 m in 2018.
 355 Nevertheless, 2019 and 2020 measurements of this borehole are considered to reflect the undisturbed in-
 356 situ subsurface temperatures at this location.



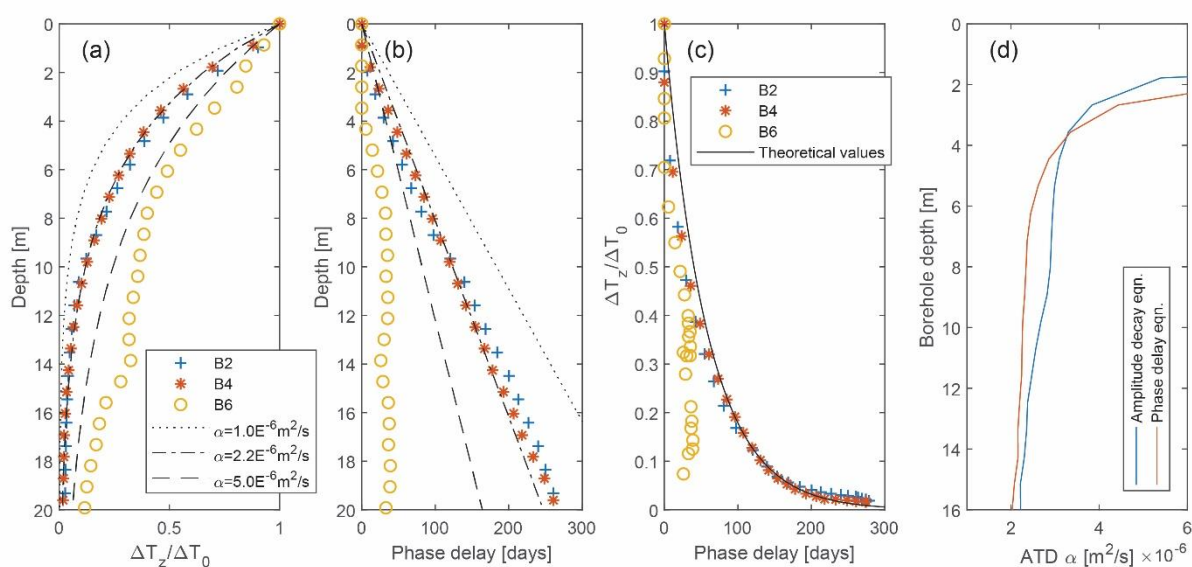
357

358 Fig. 4. Borehole temperature monitoring data of years 2018, 2019, and 2020 provided as monthly mean temperature
359 profiles (colored lines) for the three research boreholes (B2: plots a, b, c; B4: plots e, f, g; B6: plots i, j, k). The annual
360 temperature variability is illustrated in gray in the background. An annual temperature cycle presented in these plots
361 contains monthly mean temperatures starting from November of the previous year and end in October of the year

362 provided in the respective title. Plots d, h, l show the annual mean temperatures of years 2018, 2019, and 2020, as well
 363 as the minimum and maximum temperatures measured at each depth, for all the three boreholes.

364 4.1.2. Thermal diffusivity

365 Fig. 5 shows the results of the thermal diffusivity inversion based on periodic annual temperature cycles
 366 measured during 2019, which include separate estimates based on amplitude decay (Fig. 5a), phase delay
 367 (Fig. 5b), and their theoretical relationship (Fig. 5c). These results show a regular amplitude damping and
 368 phase delay with depth in boreholes B2 and B4, while B6 shows strong deviations from the one-dimensional
 369 heat diffusion model. Fig. 5c also shows that the annual temperature signal measured in B2 and B4 plots
 370 closely on the theoretical line, suggesting that conductive heat transfer dominates in these two boreholes.
 371 Temperature data measured in borehole B6 does not follow this theoretical relationship, hence, nonconduc-
 372 tive processes clearly dominate the subsurface thermal regime in the upper 20 m of this borehole.



373

374 Fig. 5. The plots a, b, and c show the annual variation of the subsurface temperature driven by periodic surface temper-
 375 ature variations. The black, dashed, dotted, and solid lines correspond to theoretical values calculated from an analytical
 376 solution of the one-dimensional heat conduction equation for a number of apparent thermal diffusivity values α . The
 377 colored markers in the plots indicate the annual temperature variations monitored in our research boreholes in year
 378 2019. (a) Amplitude decay of a periodic annual thermal signal vs. depth to ground surface. (b) Phase delay of a periodic
 379 annual thermal signal vs. depth to ground surface. (c) Relationship between phase delay and amplitude decay. (d)

380 *Estimated apparent thermal diffusivity over depth for borehole B4 data using the 1-D heat conduction equation. The*
 381 *blue line corresponds to the estimation using the amplitude decay in the measured temperature data. The red line*
 382 *corresponds to the estimation using the phase delay in the measured temperature data. The apparent thermal diffusivity*
 383 *values represent the estimates from the depth interval between the uppermost sensor and the given depth in the y-axis.*

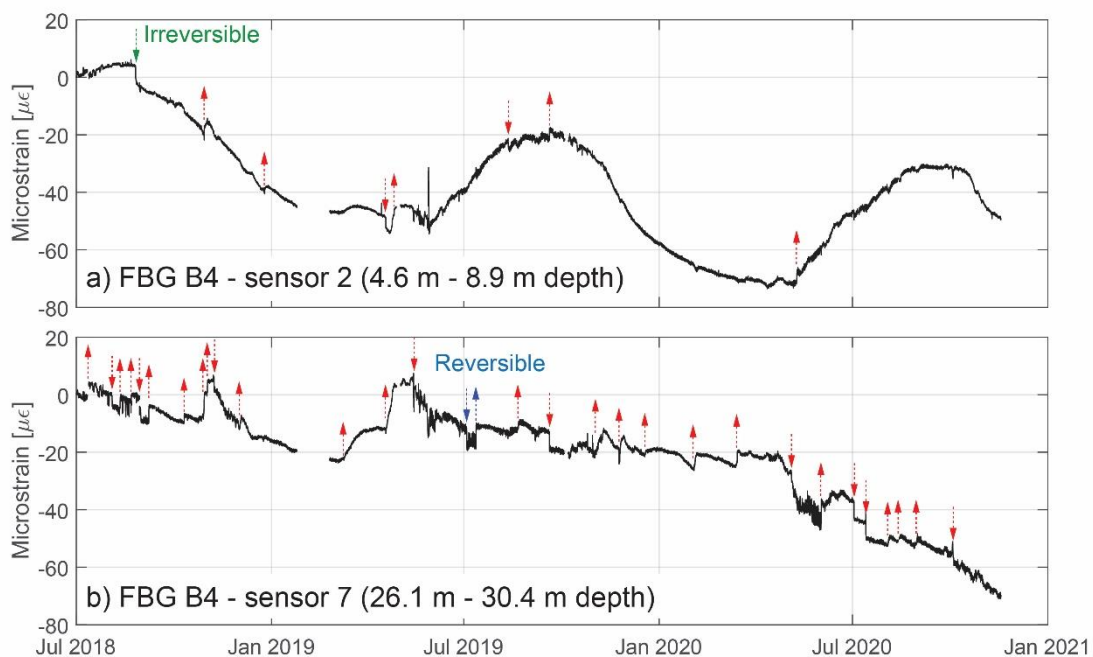
384 The best fit values for the mean apparent thermal diffusivity are $2.1 \cdot 10^{-6} \text{ m}^2/\text{s}$ for B2 and $2.2 \cdot 10^{-6} \text{ m}^2/\text{s}$ for
 385 B4, respectively. Around B4, the slope shows a relatively constant dip angle, so the 1-D approach is likely
 386 more reliable at this location compared to B2, where 2-D effects are assumed to play a more important role.
 387 Therefore, the apparent thermal diffusivity estimate of B4 was used for further investigations.

388 Fig. 5d shows the inverted depth dependent apparent thermal diffusivity for B4, based on both the amplitude
 389 decay and phase delay. The estimates from both methods show a similar trend, with decreasing apparent
 390 thermal diffusivity values with depth. The highest values occur in the uppermost 4 m and below a more
 391 gentle decrease with depth is observed. The estimated magnitudes of thermal diffusivity are higher than
 392 literature values for conductive heat transfer in intact gneissic rocks would suggest ($\sim 1.4 \cdot 10^{-6} \text{ m}^2/\text{s}$)
 393 (Wegmann et al., 1998; Wegmann, 1998). We expect the estimated apparent thermal diffusivity values
 394 within the thermally active layer to approach a more constant value close to the intact rock thermal diffusivity
 395 at greater depth. Hence, the higher estimated value is assumed to be an ‘apparent’ value that also com-
 396 prises effects related to the increased fracture density and water / air content in the surficial layer that allow
 397 minor nonconductive processes to occur. Such minor nonconductive effects should not be confused with
 398 stronger advective heat transfer processes, which are likely important in B6 and can result in significant
 399 disturbances of the diffusive profile (e.g., Ge, 1998).

400 4.1.3. Strain data

401 We measure both short- and long- term strain signals in the FBG monitoring data, which can be either
 402 reversible or irreversible. Two characteristic examples are provided in Fig. 6, and the complete FBG strain
 403 time series of all sensors in the three boreholes are provided in the Appendix (Fig. A 1, A 2, and A 3). In the
 404 present work, we interpret irreversible deformation as a proxy for rock mass damage (cf. Hugentobler et al.,
 405 2020 and references therein). Shallow sensors show annual cyclic strain components (e.g., Fig. 6a) and
 406 many sensors show a general, long term shortening trend in the 2.5 years’ time series (Fig. 6a and b). Short-

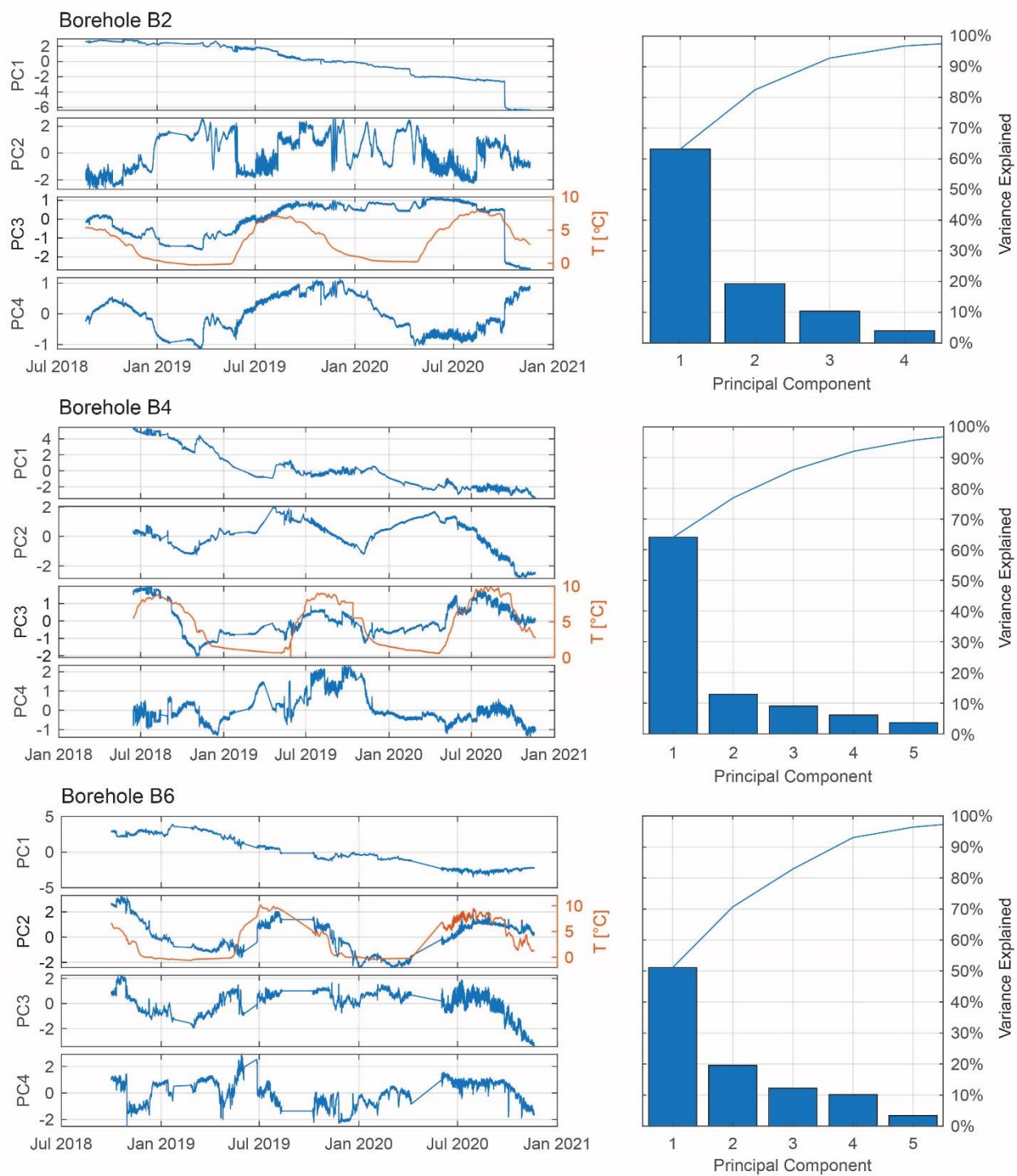
407 term and rapid strain events, indicated with arrows in Fig. 6, are superimposed on these longer-term signals.
 408 Some of these strain events (typically showing magnitudes of up to around $20 \mu\epsilon$) are irreversible (e.g. green
 409 arrow in Fig. 6a), whereas others are reversible at timescales from hours to several weeks (e.g. blue arrows
 410 in Fig. 6b). Reversible events often also contain a minor irreversible strain portion. The analyzed timeseries
 411 also contain a few larger irreversible strain events with magnitudes between 50 to a few $100 \mu\epsilon$. These large
 412 events occur most frequently in borehole B2 (Fig. A 1, Appendix), and could either be significant damage
 413 events, or an artefact of the FBG signal processing. Most of these steps could either be verified by simulta-
 414 neous deformation detected with the independent in-place inclinometer system in the borehole (cf. Hugen-
 415 tobler et al., 2020), occurred at several FBG sensors simultaneously, or coincided with potential triggers.
 416 Hence, they are interpreted as significant damage events. Only one particular large step in sensor FBG-4
 417 of borehole B2 (small form letter a in Fig. A 1) is likely a signal processing artefact and was removed in the
 418 analyses that follow.



419
 420 *Fig. 6. Time series of two FBG strain signals measured at different sensors in borehole B4. The complete strain data of*
 421 *all sensors is provided in the Appendix (Fig. A 1, A 2, and A 3). Arrows indicate the timing and direction of short-term*

422 *and rapid strain events referred to in the text. An example for a mainly irreversible strain event is shown in panel a*
 423 *(green arrow). An example for a mainly reversible strain event is indicated in panel b (blue arrows).*

424 The results of the PCA are shown in Fig. 7, and, similar to the example signals provided on Fig. 6, display
 425 both linear trends and cyclic signals that are contained in the strain time series. In all three boreholes the
 426 first principal component (PC1) shows a general shortening trend with no cyclic behavior that explains be-
 427 tween 50% to 70% of the total variance in the strain variables. The second or third principal components in
 428 B4 and B6 display annual cyclicity, and are moderately correlated with measured ground temperatures (the
 429 correlation values range from 0.70 (PC3) for B4 to 0.65 (PC2) for B6). These components explain 13% (B4)
 430 and 20% (B6) of the total variance of the measured strain. In B2 the third component (PC3) shows the
 431 relatively low maximum linear correlation coefficient of 0.3 with ground temperature, which explains only
 432 around 10% of the total variance in the strain signals. No clear correlation with other potential drivers for
 433 deformation (e.g. pore pressure variations or precipitation events) can be detected. In B2 no correlation with
 434 the pressure head can be investigated because no functioning pore pressure sensor is installed (cf.
 435 Hugentobler et al., 2020).



436

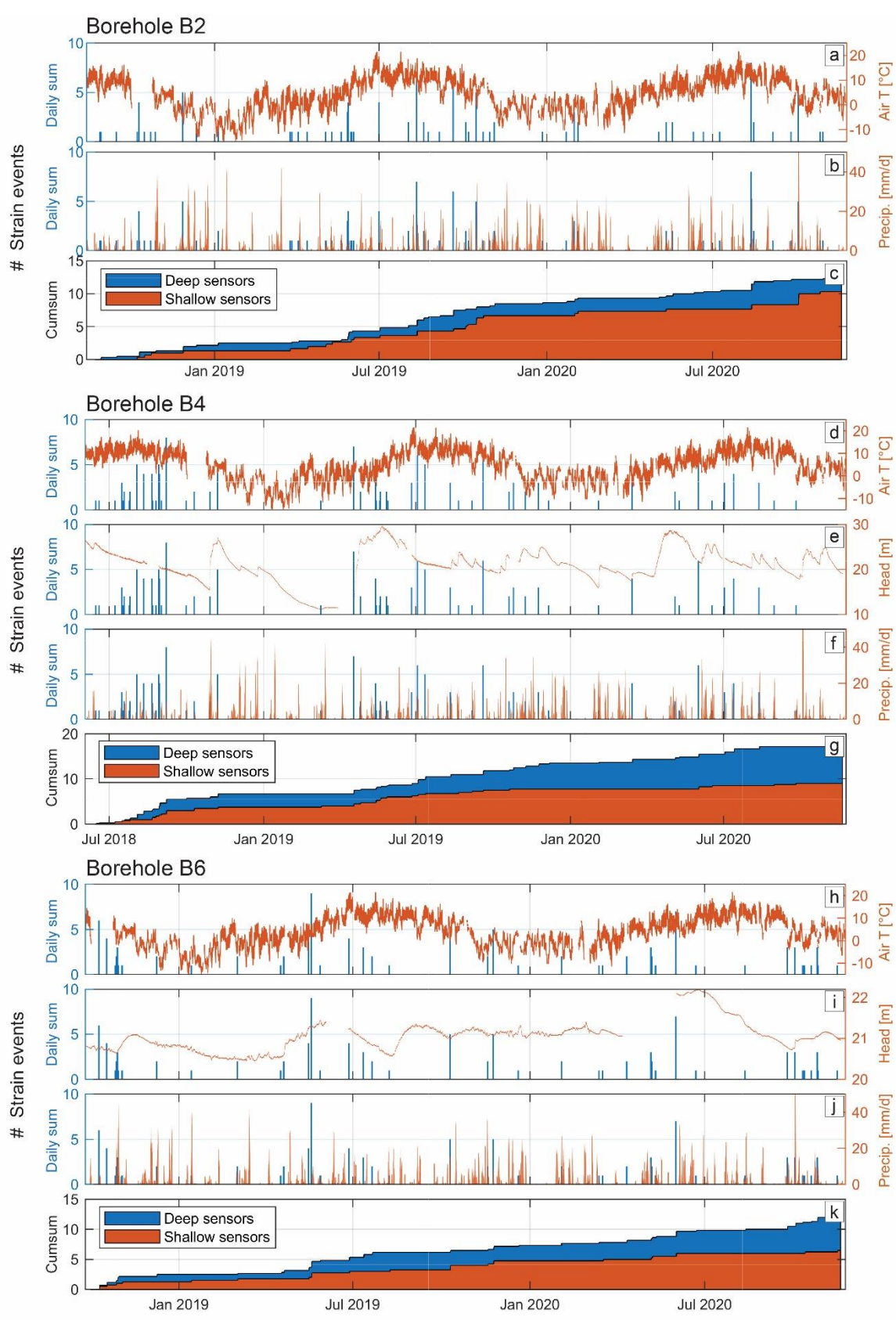
437 *Fig. 7. Plots of scores containing the principal components 1 to 4 for the strain measured in the three boreholes B2, B4,*
 438 *and B6 (left column). For each borehole, a histogram (including the cumulative sum; blue line) is provided that shows*
 439 *how much of the total signal variance can be explained by the individual principal components (right column).*

440 Fig. 8 presents a comparison of detected strain events measured with the FBG-sensors to potential drivers
 441 for deformation, which include surface temperature, pressure head, and precipitation. The data shows that,
 442 depending on the borehole, between 60 and 75 % of the strain events coincide with precipitation events
 443 (Fig. 8b, f, and j). Only about half of these precipitation events caused a clear pressure head increase (Fig.
 444 8e and i). On the other hand, 5 to 10 % of the strain events temporally correlate with pressure head increases
 445 not related to direct rainfall infiltration but to snowmelt infiltration. Very few strain events (< 2 %) coincide
 446 with extreme surface temperatures (Fig. 8a, d, and h). These results emphasize the importance of hydro-
 447 mechanical and potentially moisture-related processes for deformation, however the interactions are likely
 448 complex, as rainfall would also modify the mechanical and hydraulic glacial boundary conditions. Analyzing
 449 these complex interactions is beyond the scope of the present paper, but will be explored in a future contri-
 450 bution.

451 Between 20 to 30 % of the strain events do not coincide with any of the three potential triggers. To investi-
 452 gate if these events could be related to earthquakes, which are also frequently discussed as destabilizing
 453 factors during deglaciation (e.g., McColl and Draebing, 2019), we compared the timing of the strain events
 454 with measured earthquakes in the time period (data provided by the Swiss Seismological Service, SED).
 455 Earthquakes that occurred at epicentral distances less than 50 km from the study site and having magni-
 456 tudes greater than M1 and an earthquake depth of less than 15 km were used for the comparison. However,
 457 we did not find any correlation between the around 70 recorded earthquakes during the monitoring period
 458 (maximum magnitude M3) and the measured strain events in our boreholes. A characterization of all earth-
 459 quake events used for the comparison (i.e., magnitude, epicentral distance to study site, and earthquake
 460 depth) can be found in the Appendix (Fig. B 1).

461 The cumulative numbers of detected strain events (Fig. 8c, g, and k) allow us to better visualize the variation
 462 of the number of events occurring over the year. Data in these plots is provided as mean number of strain
 463 events per sensor interval for shallow sensors (within the annual thermally active layer, i.e. the uppermost
 464 ~ 18 m) and for deep sensors (~18 m down to the borehole end). Fig. 8 shows that fewer strain events occur
 465 during wintertime compared to the spring and summer season and that these strain events occur more

466 frequently in deeper sensors. This difference between shallow and deeper sensors is more strongly pro-
467 nounced in B4 and B6 compared to B2. In B2, some large magnitude strain events are measured in the
468 shallow layer (see Fig. A 1, Appendix).



470 *Fig. 8. Number of detected strain events during the monitoring period in the three boreholes compared to potential*
 471 *drivers for deformation. Panels a, d, h: Daily sum of events vs. air temperature at the study site. Panels e, i: Daily sum*
 472 *of events vs. pressure head measured at the borehole end. Panels b, f, j: Daily sum of events vs. cumulative total*
 473 *precipitation per 24 h measured around 5 km away from the borehole location at the weather station “Bruchji”, canton*
 474 *VS (provided by MeteoSchweiz). Panels c, g, k: Mean cumulative sum of events per sensor provided for “shallow sen-*
 475 *sors” (i.e., down to 18 m depth) and “deep sensors” (i.e., from 18 m down to the borehole end).*

476 4.2. Modelling temperature evolution and deformation during deglaciation

477 4.2.1. Drivers of temperature evolution since LIA

478 Fig. 9 shows simulated subsurface temperatures during LIA deglaciation compared to our measured mean
 479 subsurface temperatures at the borehole locations. Fig. 9a shows the temperature field of the steady-state
 480 solution at the LIA ice extent, which, as mentioned previously, is used as an initial condition for the following
 481 transient ice retreat simulation. The modeled temperature field in 2019 is shown in Fig. 9b. Fig. 9c illustrates
 482 the change of the subsurface temperature profile at the specific indicated mid-slope location during glacial
 483 retreat.

484 As can be seen on Fig. 9c, the subsurface temperature profile before deglaciation shows 0° C at the ground
 485 surface (BC1 below ice) with a nearly linear temperature increase with depth. As soon as the ice retreats
 486 below position c, a new mean annual ground temperature is applied to the model (BC2 in Table 2) which
 487 results in warmer surface temperatures that penetrate downward with time. The adaption of the subsurface
 488 temperature to a new, warmer MAGT causes a transient inverse gradient in the upper subsurface temper-
 489 ature profile (i.e. cooling downwards), which transitions to a positive geothermal gradient at depth. Hence,
 490 an inverse thermal profile can be an indication for a transient subsurface temperature adapting to a new,
 491 warmer surface temperature (e.g., Pollack et al., 1998). The dashed line in Fig. 9c illustrates a newly ap-
 492 proached steady-state temperature profile assuming the new MAGT of around 2.6° C to be constant over a
 493 very long time at the elevation of the profile. The time required to reach this new steady-state temperature
 494 profile is dependent on the thermal diffusivity of the medium.

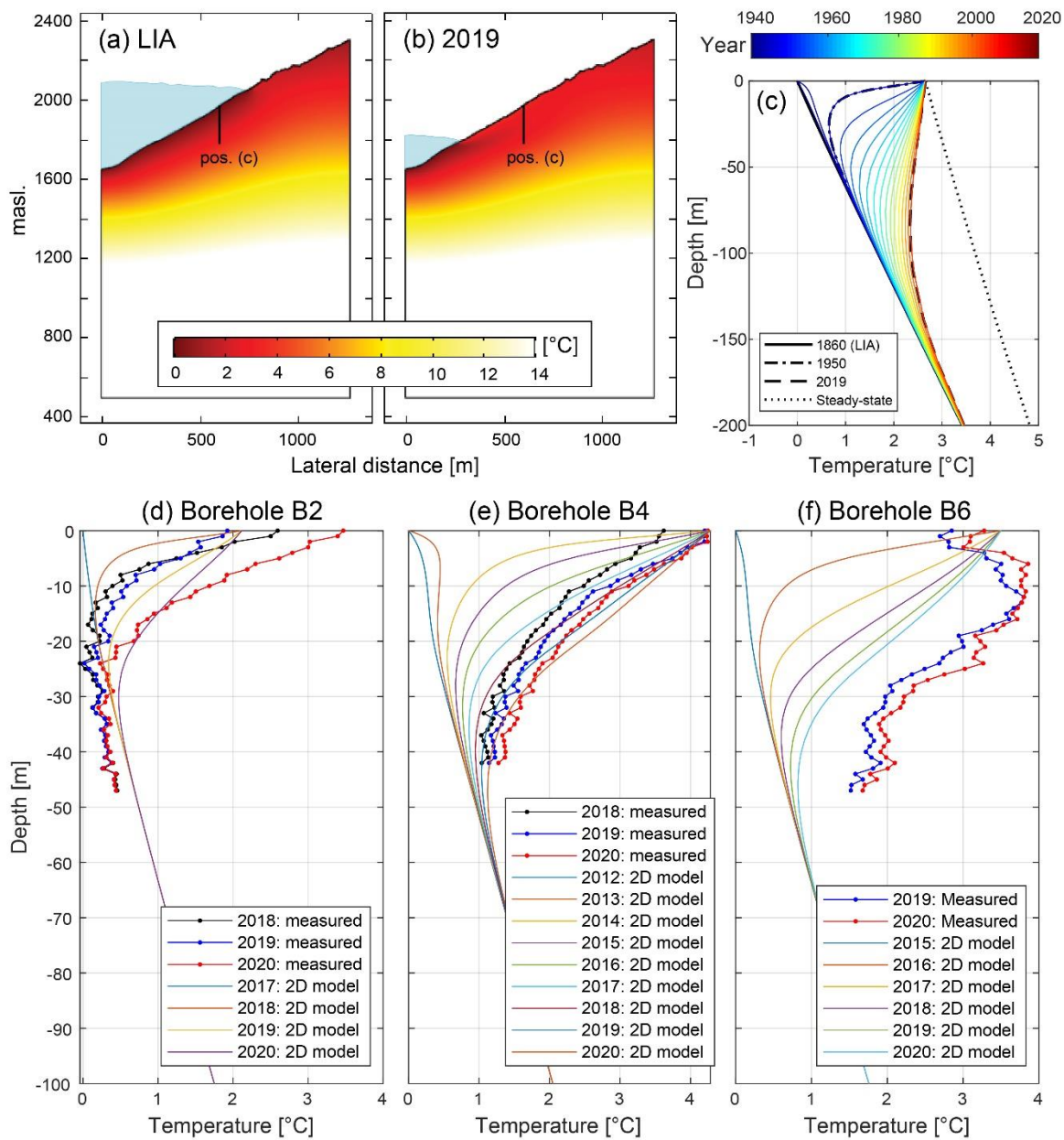
495 In order to allow a meaningful comparison of modeled temperatures with temperatures measured in the
 496 boreholes, the uncertainty related to the timing of deglaciation resulting from the lateral ice retreat function

497 (max. 2 years, cf. section 3.2) was corrected to get the true year of deglaciation in the models. Consistent
 498 with the results shown on Fig. 9c, all the measured temperature profiles show an inverse temperature gra-
 499 dient at least in the upper part of the borehole, which reflects the preserved cold temperature from the glacial
 500 occupation at depth and a more recent warming trend.

501 Fig. 9d shows that the measured temperature profile of B2 is not exactly reproduced with our conduction
 502 model and the applied boundary conditions. However, the modelled shape of the profile and depth of the
 503 warming front is similar to that measured at this borehole location (Fig. 9d). The measured temperature
 504 profile is shifted by about 0.4° C to colder temperatures compared to the modeled values and has temper-
 505 atures of around 0° C at a depth of 20 to 25 m.

506 Annual mean temperatures measured in borehole B4 can be well reproduced by the conduction model (see
 507 Fig. 9e). Nevertheless, the model shows a slightly slower warming at a depth below the annual thermally
 508 active layer and the borehole end. The measured temperature profiles within the thermally active layer (i.e.
 509 the uppermost ~ 20 m) deviate from the model because temperatures at these depths are subject to annual
 510 temperature variability that can differ from the used MAGT in the model.

511 Conversely, the measured temperature profile of B6 cannot be reproduced with our thermal conduction
 512 models. The measured temperature profile in this borehole shows a very different shape and quicker warm-
 513 ing of the subsurface compared to the simulated ground warming after ice retreat by only heat conduction
 514 (Fig. 9f). The warming front, evident by the turning point to a positive geothermal gradient, is below 50 m in
 515 the measured temperature profile and significantly shallower in the results of the conduction model.



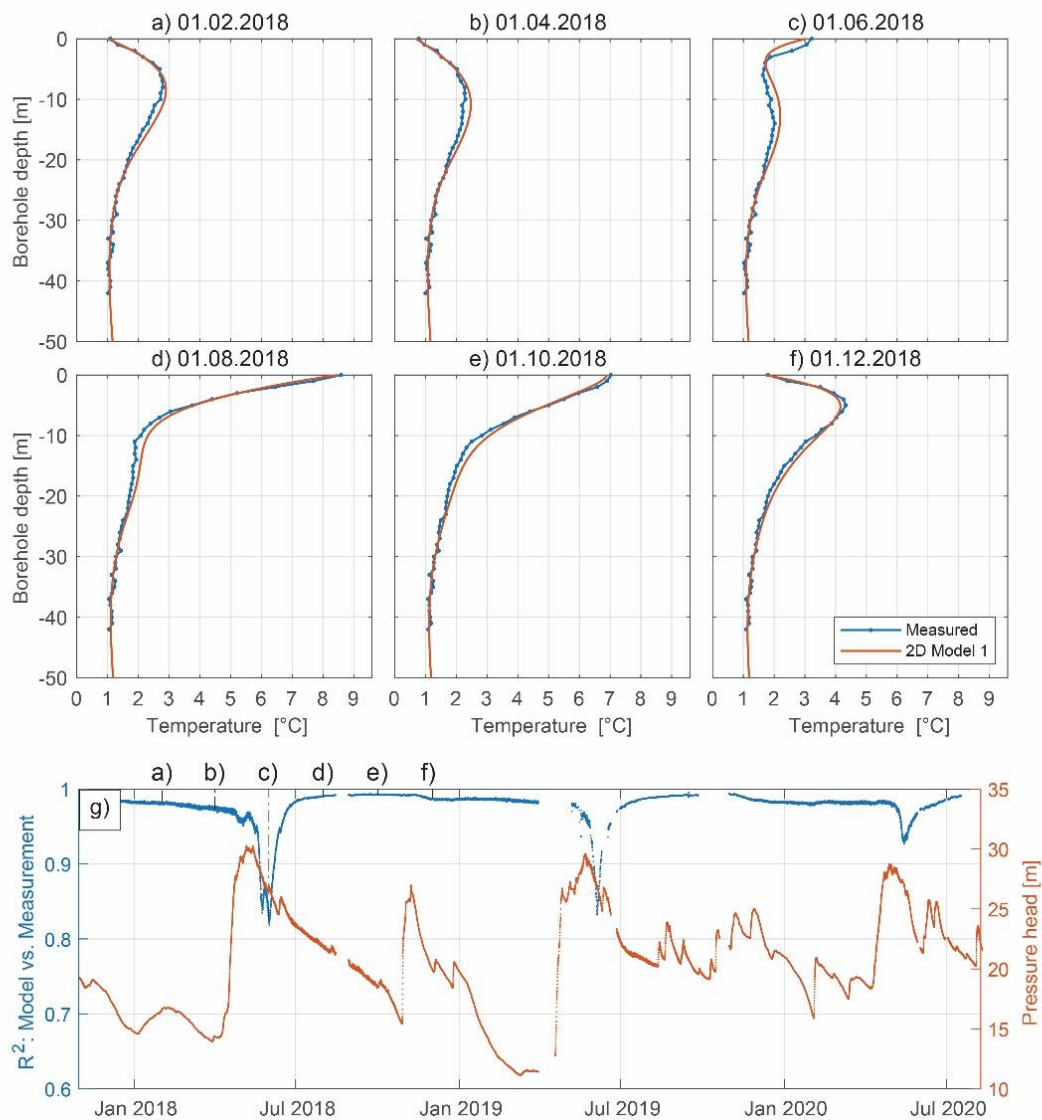
516

517 *Fig. 9. Results of the transient 2-D numerical model showing the annual mean subsurface temperature field for different*
 518 *ice elevations: (a) LIA ice extent; (b) 2019 ice extent. (c) Evolution of the annual mean subsurface temperature profile*
 519 *from 1860 to 2019 at the indicated mid-slope position c. Graphs d, e, and f show the comparison of the annual mean*
 520 *subsurface temperatures measured at the three borehole locations and the results from the 2-D numerical conduction*
 521 *model.*

522 4.2.2. Drivers of seasonal temperature variations

523 As described above, the results from the annual mean subsurface temperature simulation show good agree-
 524 ment with the monitoring data in borehole B4 (section 4.2). Therefore, we investigated recent seasonal
 525 temperature variations (modelling step 3, section 3.2) at this location, and compared results to monitoring
 526 data. Fig. 10 shows a comparison of modelled and measured temperature profiles at different months during
 527 the year 2018. The results for other years are similar. The annual surface temperature variability, both in
 528 the model and the monitoring data, affects a depth range of approximately 20 m. Measurements and model
 529 show a good accordance.

530 To investigate the mechanisms of seasonal temperature variations in detail, the correlation coefficient be-
 531 tween the modeled and measured temperature profile is calculated for each hourly time step and the R-
 532 squared value is plotted against time (see Fig. 10g). The small form letters in Fig. 10g indicate the timing of
 533 the different panels (Fig. 10a to f). The correlation coefficient time series shows a cyclic behavior with gen-
 534 erally very high values (above 0.95) from midsummer to winter. Lower correlation values are observed dur-
 535 ing spring and temporally correlate with strong pressure head rise in borehole B4, caused by snowmelt
 536 infiltration (see Fig. 10g). This negative peak in the R-squared value, however, shows a delay compared to
 537 the pressure peak. Daily acquired photos from a time lapse camera installed at the opposite valley flank
 538 with view to the borehole locations show that this negative R-squared peak coincides with the time when
 539 the rock surface surrounding the borehole becomes snow free.



540

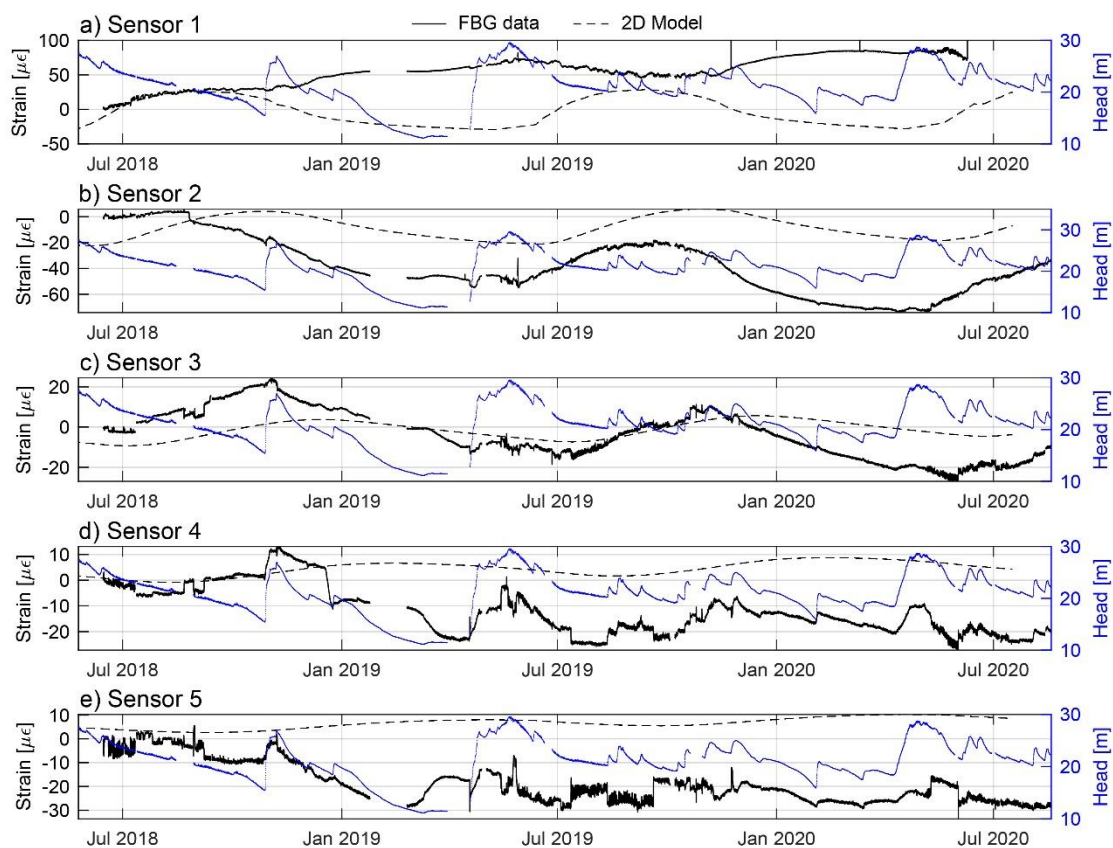
541 *Fig. 10. Panels a to f: Modeled temperature profiles together with the measured subsurface temperature profile at*
 542 *borehole location B4 at specific time steps over the time span of 1 years. Panel g: Correlation (R-squared value) be-*
 543 *tween the measured subsurface temperatures in borehole B4 and the 2D heat conduction model results over time.*
 544 *Additionally, the pressure head measured at the borehole end of B4 is provided. The small form letters indicate the*
 545 *timing of the plots a to f.*

546 4.2.3. Thermoelastic strain

547 Finally, we investigate the magnitude of thermoelastic strains in the subsurface resulting from seasonal
 548 temperature cycles (Fig. 11). This is based on a comparison between monitored and modelled thermoelastic

549 strains in borehole B4 which is dominated by conductive heat transport. In addition, the pressure head
 550 measured at the borehole end is provided in the Fig. 11.

551 The results show that the strain signal of the uppermost sensor cannot be reproduced by the simplified
 552 thermoelastic continuum model (Fig. 11a). However, strain readings of FBG-sensors number 2 and 3 show
 553 a clear annual cyclicity with similar amplitude and phase as the modeled strain (Fig. 11b and c). In the strain
 554 signal of sensor 4 (Fig. 11d), a weak cyclic signal with an annual period and an amplitude similar to the
 555 modeled one might be present superimposed to another signal strongly correlating with the pressure head.
 556 In the deeper FBG-sensors, the modeled thermoelastic strain is an order of magnitude (or more) lower than
 557 the actual strain measured in the borehole. Further, the strain variations in these depths often coincide with
 558 pressure head fluctuations.



559

560 *Fig. 11. Comparison of the modeled thermo-elastic strain (dashed line), the measured strain in the borehole for the*
 561 *uppermost 5 FBG strain sensors (solid black line), and the pressure head (blue line) measured at the borehole end in*
 562 *borehole B4. Sensor depth intervals: 1: 0 – 4.3 m; 2: 4.3 – 8.6 m; 3: 8.6 – 12.9 m; 4: 12.9 – 17.2 m, 5: 17.2 – 21.5 m.*

563 5. Discussion and interpretation

564 5.1. Controls on subsurface temperature at a retreating glacier margin

565 The results in section 4 show that heat transfer by conduction through the rock mass is an important process
 566 to explain large parts of the observed subsurface temperature field. Measured temperatures in borehole B4
 567 can be well reproduced with the 2-D heat conduction models and the results for B2 suggest that conduction
 568 plays an important role. Conversely, the temperature profile measured in borehole B6 strongly deviates from
 569 an expected conduction dominated regime and hence cannot be explained with this process. Additionally,
 570 all boreholes display a warming trend, which is likely caused by glacier retreat and subsequent exposure of
 571 the slope to annual temperature cycles. The propagation depth of this warming trend is observed to increase
 572 in depth the longer a borehole has been ice free.

573 Non-conductive heat transport processes at our monitoring boreholes are potentially driven by groundwater
 574 interaction with the adjacent glacier and yearly recharge cycles on the ice free slope. Typical patterns of
 575 such advective temperature disturbances have been discussed by Ge (1998) who distinguishes between
 576 (1) distributed vertical groundwater flow in semiconfined porous aquifers, (2) horizontal flow in confined
 577 aquifers, and (3) localized groundwater flow in preferential groundwater pathways such as fractures in
 578 gneiss and granite, all causing substantially different signatures (Bredehoeft and Papaopulos, 1965; Ge,
 579 1998). Such localized groundwater flow through fractures is assumed to be the cause for the observed
 580 positive and negative temperatures anomalies along the profiles (cf. Hugentobler et al., 2020). In order to
 581 develop localized thermal anomalies, the water flowing in transmissive fractures has to be in thermal dise-
 582 quilibrium with the surrounding rock and must be forced to flow by a natural head gradient, e.g. resulting
 583 from a stationary or transient groundwater flow field in an inclined slope.

584 In borehole B2, Fig. 9d shows a temperature offset of the modeled profile from the measured data of about
 585 0.4° C between depths of 25 m and 50 m. This deviation can be explained by cooling the subsurface with
 586 cold water directly connected to glacial ice, which could have caused temperatures of around 0° C down to

587 a borehole depth of 25 m during the time of ice occupation. We suggest that the high fracture density in this
 588 shallow layer (see Fig. 2), as well as the relatively low normal distance (~ 18 m at borehole meter 25) to the
 589 rock surface because of the slope steepening below the borehole, enabled efficient advective heat transfer
 590 during time of ice occupation and during the years when the glacial ice was still close (i.e., 2018 and 2019).
 591 The measured temperature increase at around 25 m that is observed between 2019 and 2020 (*Fig. 4d*) can
 592 be related to a decoupling from the advective glacial hydraulic system due to the ongoing ice retreat and an
 593 enabling of the penetration of the conductive warming front from the surface.

594 Relatively minor non-conductive heat transport processes are likely occurring at the B4 borehole location.
 595 This is evident when comparing the seasonal variability of the temperature profile measured in the borehole
 596 with the 1-D analytical solution and the 2-D numerical modeling results (Figs. 8 and 10). The correlation
 597 values reveal an annual repeating decrease during the time of snowmelt in spring, indicating a secondary
 598 heat transport process superimposed on the purely conductive transport model (Fig. 10g). Lower correla-
 599 tions are observed in the upper 20 to 30 m of borehole B4. The peak of the reduced correlation values
 600 occurs a few weeks after the pore pressure peak (recorded at 43.75 m depth), at a time when the borehole
 601 location becomes snow-free. We attribute the observed low correlation to advective heat transfer processes
 602 related to water infiltration happening directly at the borehole location.

603 Borehole B6 temperature measurements cannot be explained using a conduction model, the thermally ac-
 604 tive layer is 10 m deeper as compared to B2 and B4, and the measurements display larger temperature
 605 variations. This borehole is located within an inactive (presumably dormant) landslide body with substantially
 606 more disturbed and damaged rock mass compared to the other two borehole locations (cf. section 2). Ad-
 607 ditionally, a creek fed by nearby regional springs passes through a morphological depression next to the
 608 borehole. During drilling in fall 2017, bubbles from the pressurized air used for drilling extruded all around
 609 this depression, revealing a high transmissivity in the shallow subsurface. It is assumed that the availability
 610 of creek water to infiltrate into the subsurface at this location and higher transmissivity due to the disturbed
 611 rock mass enables strong advective heat transfer processes to occur in the shallow subsurface, as docu-
 612 mented in Anderson (2005).

613 5.2. Deformation and implications for rock mass weakening

614 5.2.1. Strains in the thermally active layer

615 The analyses of chapter 4.1.3 and 4.2.3 have revealed differences in the drivers of the FBG strain signals
 616 as a function of depth, as well as between the borehole locations. All boreholes feature a principal compo-
 617 nent that displays an annual cyclic strain signal that correlates with ground temperature, and explains be-
 618 tween 10 and 20% of the variance in the strain data. Given the correlation to ground temperature, as well
 619 as the results of the numerical modeling (Fig. 11, discussed below), it is likely that this component corre-
 620 sponds to thermoelastic deformation. The thermoelastic component is less pronounced in the PCA result of
 621 B2, likely because the overall deformation in this borehole is larger than the other two boreholes, and few
 622 rapid deformation events dominate the signal (Fig. A 1, Appendix).

623 The numerical modeling shows that thermoelastic deformation occurs in the thermally active layer, and
 624 decreases with depth. This can be clearly observed in our monitoring data of borehole B4 with the most
 625 intact rock mass conditions. The FBG data from sensor 2 and 3 (Fig. 11), which correspond to depth inter-
 626 vals of 4.6 to 8.9 m and 8.9 to 13.2 m, show that here the cyclic, thermoelastic deformation signal dominates
 627 over other signals. Deeper in the borehole (i.e. from ~ 13.2 m downwards) other signals, that show minor or
 628 no correlation with the modeled thermoelastic deformation, dominate the strain time series. In borehole B2
 629 and B6, with less intact rock mass conditions, the uppermost few sensors also show signals with an annual
 630 cyclicity that is attributed to thermoelastic effects. However, this continuous cyclic deformation is interrupted
 631 by abrupt strain events of similar magnitude or superimposed to sometimes even stronger deformation
 632 trends (see Fig. A 1 and A3, Appendix).

633 An exception is the strain signal measured in the uppermost sensor of all boreholes, which shows an inverse
 634 correlation with the model data and hence must be related to a different driving mechanism. Expansion in
 635 wintertime and contraction in summertime could be caused by the effect of hygroscopic expansion or mois-
 636 ture induced strain (e.g., Gor et al., 2017) and related to potentially higher relative humidity in the shallow
 637 rock layer below the snowpack compared to dry summer conditions. Another possibility to generate exten-
 638 sional strain in wintertime and contraction in summer are frost weathering processes such as volumetric
 639 expansion or ice segregation (Draebing et al., 2017). This, however, is unlikely to be the reason for the

640 measured strain at the uppermost sensor as no negative subsurface temperatures were measured in win-
 641 tertime.

642 5.2.2. Strains below the thermally active layer

643 Some strain signals measured in deeper sensors clearly correlate with pressure head changes in the bore-
 644 hole (e.g. sensor 4 and 5 in Fig. 11) and hence are probably strongly driven by hydromechanical effects.
 645 Other sensors show rapid or short-term strain events often temporally coinciding with pressure head
 646 changes or precipitation events but overall show a low accordance to the pressure head signal. No individual
 647 PCA mode clearly correlates with the measured pressure head, however deformation signals that temporally
 648 coincide with pressure head changes are present in several PCA components. We explain this with the
 649 more complex nature of hydromechanically coupled deformation in a fractured rock slope and an assumed
 650 effect of fluctuating mechanical and hydraulic glacial boundary conditions. Hydromechanically forced strain
 651 signals can be substantially different depending on whether the measurement interval is intersecting a trans-
 652 missive structure, and on the mode of fracture deformation (Krietsch et al., 2020). Effective glacial ice load
 653 changes, due to variations in the subglacial water pressures related to daily meltwater cycles and rainfall
 654 infiltration into the glacier during summertime, might be strong enough to cause direct strain and additional
 655 hydromechanically coupled effects. Hence, all these processes together are not expected to generate a
 656 uniform response that could show a linear correlation with pore pressure. We relate the observed strain
 657 events either to poroelastic reactions (showing a strong reversible component and coinciding with pore
 658 pressure fluctuations) or to irreversible fracture slip and fracture propagation (damage) driven by pore pres-
 659 sure fluctuations or presumably also glacial load changes from varying englacial water pressures. A detailed
 660 discussion of hydromechanical drivers for deformation is beyond the scope of the present work but will be
 661 explored in a future contribution.

662 The higher number of reversible and irreversible short-term strain events detected in deeper, pore pressure
 663 controlled sensors compared to the sensors within the thermally active layer (Fig. 8), can be related to
 664 hydromechanical effects. During summer and fall a higher activity in the hydraulic systems of the slope and
 665 glacier must be expected, i.e. groundwater recharge by snowmelt and rainfall infiltration induce recharge
 666 variations and higher hydraulic gradients in the rock slope, and daily ice melt in summer strongly increases

667 the dynamics of the subglacial hydrological system (Harper et al., 2005). We interpret these effects to be
 668 the reason for the observed increased number of strain events occurring during the summer season.

669 5.2.3. Implications for rock mass damage

670 As discussed in the previous section, strain signals during our 2.5-year measurement period are partly
 671 driven by thermomechanical and hydromechanical processes. The majority of our detected short-term and
 672 rapid strain events coincide with precipitation events, of which about 50 % were strong enough to cause a
 673 clear reaction in the pore pressure signal due to infiltration. As discussed above, many of these strain events
 674 contain at least a minor irreversible strain contribution, which means that the irreversible deformation is also
 675 strongly controlled by hydromechanical effects. This observation agrees with the numerical modeling results
 676 of Grämiger et al. (2020), who conclude that hydromechanical effects are more effective in generating long
 677 term damage in paraglacial rock slopes compared to thermomechanical effects.

678 The strain time series also contain longer-term irreversible strain trends (predominantly vertical contraction),
 679 and 20 to 30 % of the short-term strain events cannot directly be related to any of the investigated drivers.
 680 We suggest that these signals may be related to fatigue processes occurring in the studied rock slope. As
 681 summarized in the introduction, cyclic stresses can promote fatigue processes and drive damage. In our
 682 studied slopes, cyclic loading is mainly driven by annual temperature variations, pore pressure fluctuations
 683 and fluctuations of the mechanical load placed on the slope by the Aletsch glacier. In the shallow subsurface,
 684 cyclic loading from annual thermoelastic deformation dominates the reversible signal causing largest strain
 685 amplitudes of $\sim 60 \mu\epsilon$ at an annual cyclicity within the uppermost 4 m and strain amplitudes decreasing with
 686 depth (to $\sim 5 \mu\epsilon$ at around 20 m depth). Below around 12 m depth, reversible strain cycles that are not
 687 related to surface temperature fluctuations and show amplitudes up to $\sim 20 \mu\epsilon$ dominate the signal. These
 688 cyclic strains are driven by pressure head fluctuations due to snowmelt infiltration or heavy rainfalls and
 689 normally occur at frequencies greater than 1/yr. The fact that irreversible strain events are often triggered
 690 by similar pressure head magnitudes as the ones causing cyclic reversible deformation, shows that many
 691 fractures must be close to a critical stress state. Therefore, we propose that these loading cycles are strong
 692 enough to promote progressive damage by cyclic fatigue. This process could also explain the detected
 693 short-term strain events that do not temporally coincide with any of the investigated triggers (about 20 - 30

694 % of all strain events).

695 The PCA component 1 showing an axial (vertical) overall shortening trend explains the majority of the vari-
 696 ance in all three boreholes. Because this signal is mainly irreversible (while the other components often
 697 show reversible features) the variance explained by PC1 (50 to 70 %) provides a good estimate for the
 698 relative portion of irreversible deformation during our monitoring period. Further, we suggest that the short-
 699 ening trend in PC1 is composed of (1) a slow, continuous long-term strain signal that can be observed in
 700 several sensors (Fig. A 1, A 2, and A 3, Appendix) and (2) the accumulation of major compressive short
 701 term strain events (e.g., steps in PC1 of B2, Fig. 7). We relate the irreversible vertical strains in our relatively
 702 limited time series to long-term progressive rock mass damage in our deglaciating environment. Compres-
 703 sive strain can originate either from slip along toppling fractures steeply dipping into the slope or from slip
 704 along sliding fractures dipping in downslope direction (cf. Fig. 15 in Hugentobler et al., 2020). Fractures of
 705 both orientations frequently occur at our study site (Hugentobler et al., 2020). Besides the annual thermo-
 706 mechanical and hydromechanical coupled drivers for progressive rock mass damage, longer term stress
 707 changes due to unloading from glacial ice downwasting and changes in the thermal and hydraulic boundary
 708 conditions connected to the ice retreat might also contribute to this observed deformation (cf. Grämiger et
 709 al., 2017; Grämiger et al., 2018; Grämiger et al., 2020).

710 Finally, we find that the magnitude of measured strain responses depends on the local rock mass quality at
 711 the given borehole location. This is observed in our research boreholes, as strong pore pressure variations
 712 (e.g. due to the heavy rain storm in early Oct. 2020) cause minor irreversible responses in the more intact
 713 rock mass of B4, but very clear irreversible deformation in the more strongly disturbed rock mass around
 714 B2. This observation is consistent with the numerical investigations of Gischig et al. (2011a) and Grämiger
 715 et al. (2020), who show that the efficiency of thermo-mechanical and hydromechanical loading cycles for
 716 fracture propagation (damage) in fractured crystalline rocks strongly depend on the magnitude of the drivers,
 717 as well as the initial damage state or criticality of the rock mass.

718 6. Summary and conclusion

719 We use three full years of hourly measured borehole temperature and high resolution FBG strain monitoring
 720 data to investigate the transient subsurface temperature regime and micrometer-scale deformation in a rock

721 slope adjacent to a retreating, temperate valley glacier. The presented data give unique new insights into
 722 the thermo-mechanical and hydromechanical drivers of rock slope deformation adjacent to retreating glaci-
 723 ers over decadal to seasonal time scales. Detailed analysis and decomposition of the subsurface defor-
 724 mation signals allow to identify various potential drivers for reversible (elastic) and irreversible deformation.
 725 This knowledge is important for understanding the main processes contributing to short- and long-term pro-
 726 gressive rock mass damage that potentially lead to the formation of paraglacial rock slope instabilities. The
 727 most important findings of the present work can be summarized as follows:

- 728 1. The thermal investigation reveals that the subsurface temperatures at our monitored slope are cur-
 729 rently in a transient state adjusting to new surface temperature conditions after glacial ice retreat.
 730 We show that heat conduction is a dominant process and explains most of the observed transient
 731 subsurface temperatures at sites with intact rock, however, locally strong deviations from a conduc-
 732 tion dominated regime can occur. These deviations can be explained by nonconductive processes
 733 (e.g., heat transfer by water) that become important when the rock mass is fractured (i.e., has an
 734 increased transmissivity), and are efficient when water and a hydraulic head gradient is available.
 735 The observed nonconductive temperature deviations in our study area are related to groundwater
 736 exchange with cold subglacial water, snowmelt infiltration, or creek water infiltration.
- 737 2. Our strain monitoring data analysis shows that annual surface temperature changes, precipitation
 738 events and pore pressure fluctuations are important drivers for deformation at our study site. How-
 739 ever, longer-term trends in the strain time series and a minority of short-term strain events cannot
 740 directly be related to any of the investigated drivers. We find that thermo-elastic effects due to an-
 741 nual temperature cycles dominate the strain signal in the uppermost 12 m of our stable fractured
 742 crystalline rock slope. Below this depth the hydromechanical effects dominate the deformation. Fur-
 743 ther, we show that both reversible and irreversible short-term deformation more frequently occur
 744 during summertime when the hydrological system in the rock slope and in the temperate glacier are
 745 more dynamic.
- 746 3. Irreversible deformation in our time series, which we relate with progressive rock mass damage,
 747 occurs as short-term or rapid strain events within hours or days and as slower, continuous strain

748 trends (mainly in vertical contraction direction). We show that the majority of short-term strain events
 749 coincide with precipitation events or pore pressure changes. Other potential drivers such as extreme
 750 surface temperatures or earthquakes occurring in the area do not show temporal correlation with
 751 strain events. We propose that cyclic fatigue from above described thermomechanical and hydro-
 752 mechanical loading cycles, as well as longer term stress changes related to deglaciation (e.g., gla-
 753 cial ice downwasting and connected changes in thermal, hydraulic and mechanical boundary con-
 754 ditions) can explain both the short-term strain events that do not coincide with any investigated
 755 driver, as well as the continuous irreversible strain trends observed in our boreholes.

756 4. We show that the magnitude of irreversible deformation depends on the magnitude of the driver but
 757 also on the spatially variable pre-existing damage at our three borehole locations. At our most stable
 758 borehole location, single events with irreversible strain magnitudes are normally on the order of a
 759 few $\mu\epsilon$, whereas at the more damaged borehole location irreversible strain magnitudes up to tens
 760 or few hundreds of $\mu\epsilon$ have been recorded.

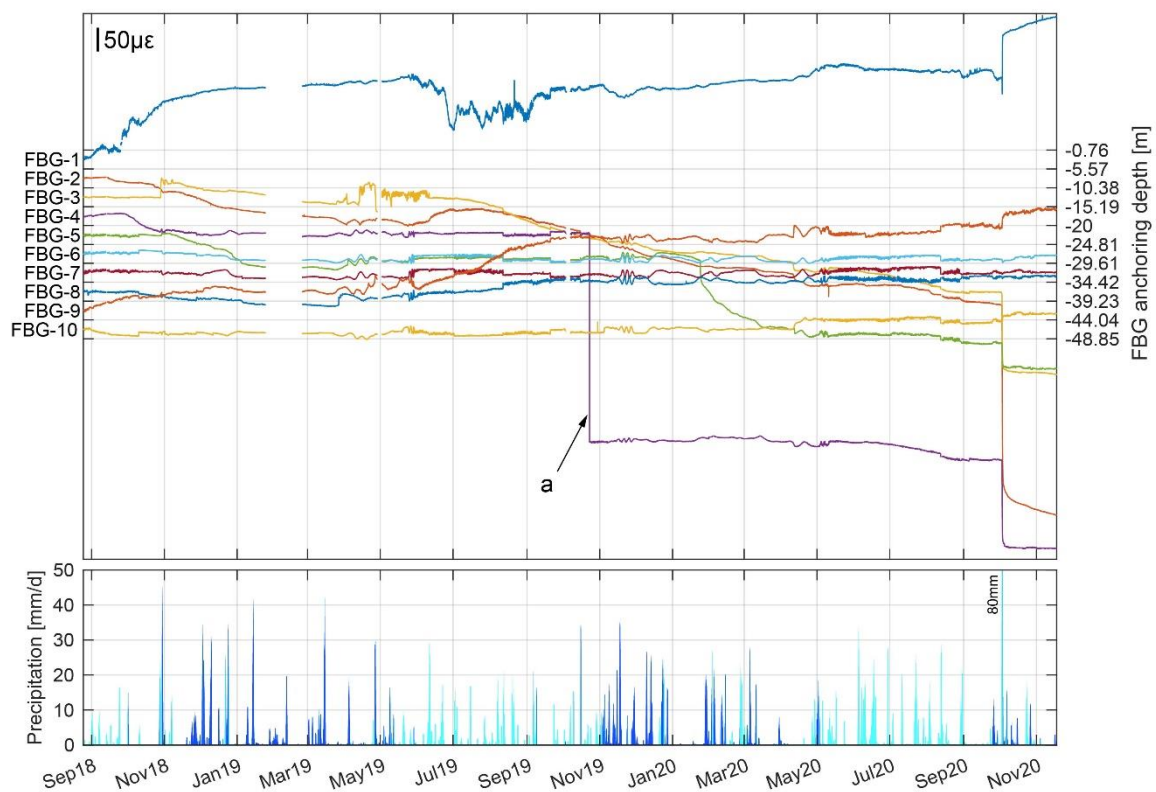
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 764 lapse images from the study site to constrain the timing of snow coverage at the borehole locations. Further,
 765 we are grateful for the inputs on the PCA analysis from Nicolas Oestreicher and Clément Roques. Also, we
 766 want to thank Reto Seifert and the numerous field helpers who joined our maintenance visits and helped to
 767 keep the monitoring system running. Data will be made accessible by acceptance on the ETH Zurich online
 768 repository (<https://www.research-collection.ethz.ch/>).

769 Appendix A: FBG strain monitoring data

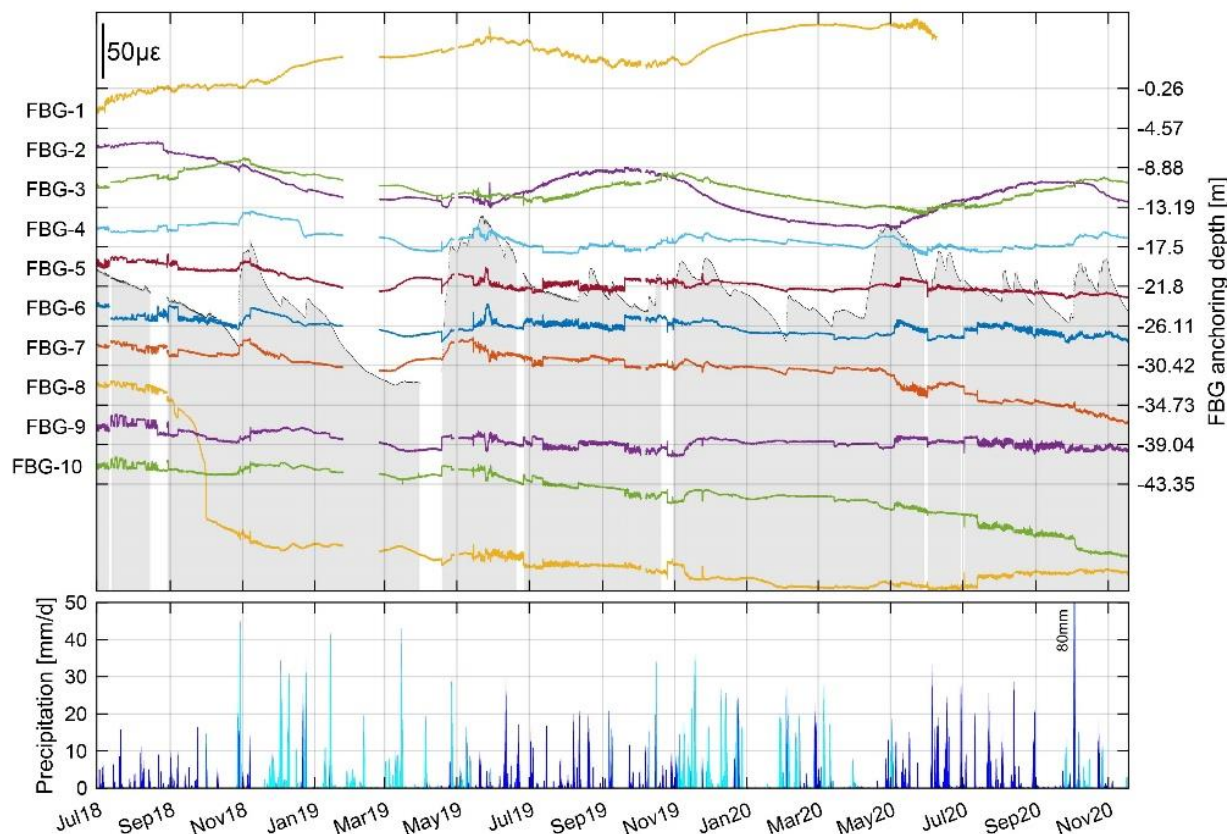
770 Fig. A 1, A 2, and A 3 contain the temperature corrected hourly mean FBG-strain measurements for each
 771 sensor in the three vertical research boreholes. The setup and detailed description of the monitoring system
 772 is provided in Hugentobler et al. (2020). All FBG sensors of the three boreholes contain a data gap of about
 773 one month in January 2019 which is related to a winter storm that broke the solar panel used for the power
 774 supply of the system. A second data gap of around two month duration is existing in FBG monitoring data

775 of all sensors in borehole B6 in summer 2019 (Fig. A 3). This gap is related to the breakage of the reinforced
 776 fiberoptical cable connecting the borehole location to the FBG data acquisition system that happened during
 777 a debris flow event. Furthermore, sensor 1 of boreholes B4 and B6 contain incomplete data recordings
 778 because of malfunctioning FBG temperature sensors causing traveling reflection peaks that sometimes in-
 779 terfere with the reflected light peaks from the strain sensors because of similar wavelength.

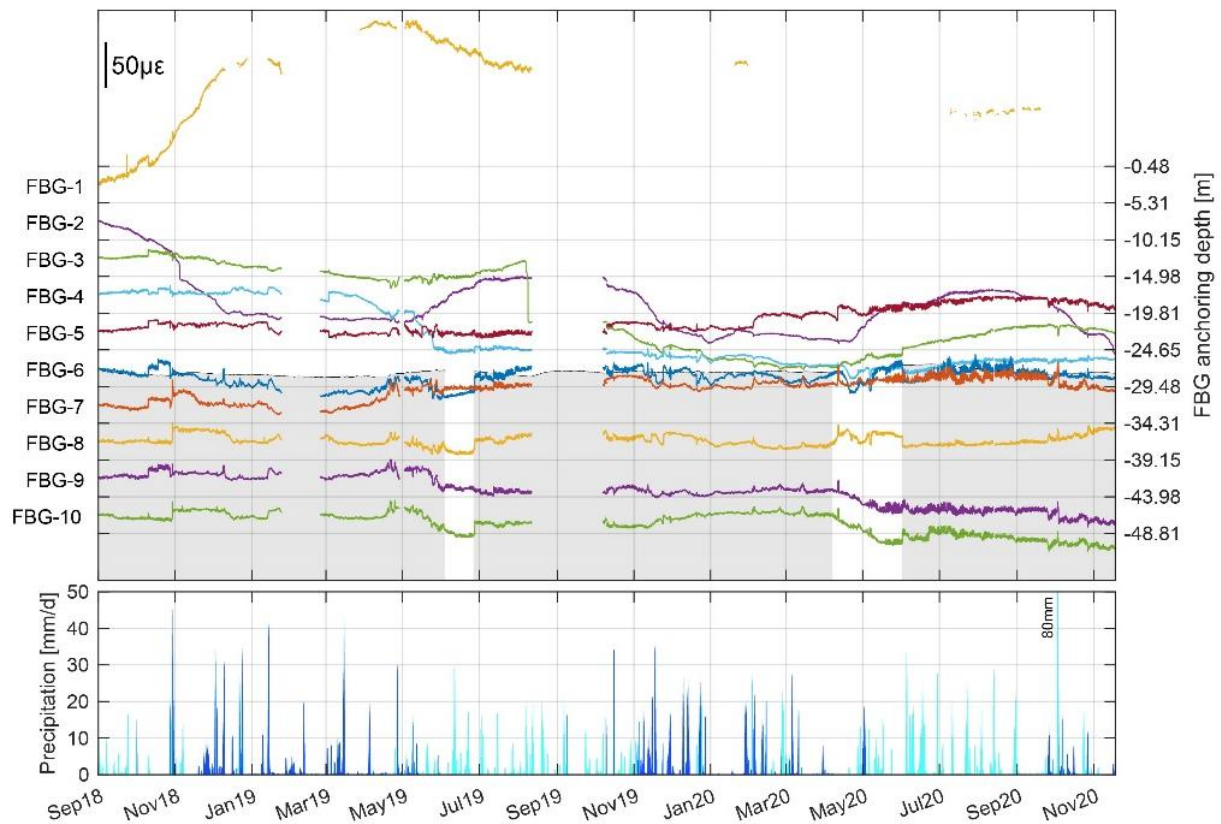


780
 781 *Fig. A 1. Upper plot: The colored lines show a visualization of the temperature corrected strain data of the individual*
 782 *4.81 m base-length FBG sensors installed in borehole B2. The strain data of each sensor is plotted in the center of the*
 783 *specific depth interval, and the vertical axis ticks correspond to anchoring depth of the FBG sensors. Positive strain*
 784 *means elongation and negative strain compaction of the sensor, and the strain scale is provided on the top left in*
 785 *microstrain ($\mu\epsilon$). (a) Indicated strain step in FBG-4 was identified as a measurement artifact and corrected before the*
 786 *analysis. Lower plot: The vertical bars show the cumulative total precipitation data (per 24 hours) provided by Mete-*
 787 *oSchweiz from the weather station "Bruchji" (Valais) located approximately 6 km away from our study site. The blue*

788 colored bars indicate precipitation that presumably occurred as snow (i.e., at surface temperatures below 1°C at our
 789 study site (cf. Jennings et al., 2018)).



790
 791 Fig. A 2. Upper plot: The colored lines show a visualization of the temperature corrected strain data of the individual
 792 4.31 m base-length FBG sensors installed in borehole B4. The strain data of each sensor is plotted in the center of the
 793 specific depth interval, and the vertical axis ticks correspond to anchoring depth of the FBG sensors. Positive strain
 794 means elongation and negative strain compaction of the sensor, and the strain scale is provided on the top left in
 795 microstrain ($\mu\epsilon$). Additionally, the elevation of the pressure head in the borehole is provided as a gray area in the back-
 796 ground of the plot. Lower plot: The vertical bars show the cumulative total precipitation data (per 24 hours) provided by
 797 MeteoSchweiz from the weather station “Bruchji” (Valais) located approximately 6 km away from our study site. The
 798 blue colored bars indicate precipitation that presumably occurred as snow (i.e., at surface temperatures below 1°C at
 799 our study site (cf. Jennings et al., 2018)).



800

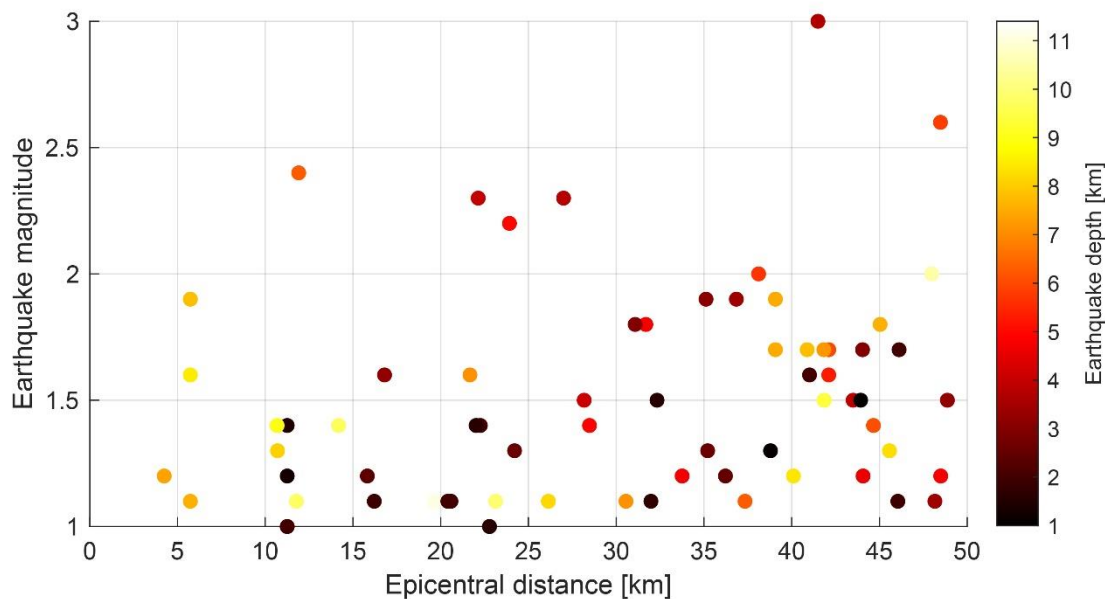
801 *Fig. A 3. Upper plot: The colored lines show a visualization of the temperature corrected strain data of the individual*
 802 *4.83 m base-length FBG sensors installed in borehole B6. The strain data of each sensor is plotted in the center of the*
 803 *specific depth interval, and the vertical axis ticks correspond to anchoring depth of the FBG sensors. Positive strain*
 804 *means elongation and negative strain compaction of the sensor, and the strain scale is provided on the top left in*
 805 *microstrain ($\mu\epsilon$). Additionally, the elevation of the pressure head in the borehole is provided as a gray area in the back-*
 806 *ground of the plot. Lower plot: The vertical bars show the cumulative total precipitation data (per 24 hours) provided by*
 807 *MeteoSchweiz from the weather station "Bruchji" (Valais) located approximately 6 km away from our study site. The*
 808 *blue colored bars indicate precipitation that presumably occurred as snow (i.e., at surface temperatures below 1°C at*
 809 *our study site (cf. Jennings et al., 2018)).*

810 Appendix B: Earthquake events

811 In Fig. B 1 all earthquake events used for the comparison with the strain events measured in our research
 812 boreholes are characterized by means of earthquake magnitude, epicentral distance to the study site, and

813 earthquake depth. We used all recorded earthquakes that occurred during our monitoring period with mag-
 814 nitudes greater than M1, epicentral distances of less than 50 km to our study site, and with depth between
 815 0 and 15 km.

816



817

818 *Fig. B 1. Earthquake catalog used for the comparison with strain events detected in our research boreholes (data pro-*
 819 *vided by the Swiss Seismological Service (SED)). The earthquake magnitude is provided vs. the epicentral distance of*
 820 *the individual earthquake event from our study site. Depths information for the individual earthquake events are shown*
 821 *with the color scale.*

822 7. References

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