

1 **Simulating the effect of subsurface drainage on the thermal regime** 2 **and ground ice in blocky terrain, Norway**

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11 **Abstract.** Ground temperatures in coarse, blocky deposits such as mountain blockfields and rock glaciers have long been
12 observed to be lower in comparison with other (sub)surface material. One of the reasons for this negative temperature anomaly
13 is the lower soil moisture content in blocky terrain, which decreases the duration of the zero curtain in autumn. Here we used
14 the CryoGrid community model to simulate the effect of drainage on the ground thermal regime and ground ice in blocky
15 terrain permafrost at two sites in Norway. The model setup is based on a one-dimensional model domain and features a surface
16 energy balance, heat conduction and advection, as well as a bucket water scheme with adjustable lateral drainage. We used
17 three idealized subsurface stratigraphies, *blocks only*, *blocks with sediment* and *sediment only*, which can be either *drained*
18 (i.e. with strong lateral subsurface drainage), or *undrained* (i.e. without drainage), resulting in six scenarios. The main
19 difference between the three stratigraphies is their ability to retain water against drainage: while the *blocks only* stratigraphy
20 can only hold small amounts of water, much more water is retained within the sediment phase of the two other stratigraphies,
21 which critically modifies the freeze-thaw behaviour. The simulation results show markedly lower ground temperatures in the
22 *blocks only, drained* scenario compared to other scenarios, with a negative thermal anomaly of up to 2.2 °C. For this scenario,
23 the model can in particular simulate the time evolution of ground ice, with build-up during and after snow melt and spring and
24 gradual lowering of the ice table in the course of the summer season. The thermal anomaly increases with larger amounts of
25 snowfall, showing that well drained blocky deposits are less sensitive to insulation by snow than other soils. We simulate
26 stable permafrost conditions at the location of a rock glacier in northern Norway with a mean annual ground surface
27 temperature of 2.0–2.5 °C in the *blocks only, drained* simulations. Finally, transient simulations since 1951 at the rock glacier
28 site (starting with permafrost conditions for all stratigraphies) showed a complete loss of perennial ground ice in the upper 5
29 m of the ground in the *blocks with sediment, drained* run, a 1.6 m lowering of the ground ice table in the *sediment only, drained*
30 run and only 0.1 m lowering in the *blocks only, drained* run. The interplay between the subsurface water/ice balance and
31 ground freezing/thawing driven by heat conduction can at least partly explain the occurrence of permafrost in coarse blocky
32 terrain below the elevational limit of permafrost in non-blocky sediments. It is thus important to consider the subsurface

33 water/ice balance in blocky terrain in future efforts on permafrost distribution mapping in mountainous areas. Furthermore, an
34 accurate prediction of the evolution of the ground ice table in a future climate can have implications for slope stability, as well
35 as water resources in arid environments.

36 **1 Introduction**

37 Permafrost is defined as ground that remains at or below 0 °C for two or more consecutive years (Van Everdingen, 1998). It
38 is a common feature in the Arctic and high mountain environments, where permafrost occurs even in mid- and low latitudes
39 (Gorbunov, 1978). Different permafrost zones are classified based on the aerial extent of permafrost presence. These zones
40 are: continuous, discontinuous, sporadic and isolated, where the surface is underlain by permafrost in more than 90%, 50-
41 90%, 10-50% and less than 10% of the land area, respectively (Smith and Riseborough 2002). Snow is an important factor in
42 governing ground temperatures and permafrost distribution within an area (e.g. Zhang et al., 2001; Zhang 2005; Goodrich,
43 1982), especially in mountain areas where permafrost is often associated with a shallow snow cover (e.g. Gislén et al., 2014;
44 Luetsch et al., 2004). The influence of soil moisture is complicated as it has an impact on the surface energy balance (e.g.
45 Liljedahl et al., 2011), the thermal characteristics of the soil (e.g. Göckede et al., 2017), and freezing/thawing dynamics (e.g.
46 Hinkel et al., 2001; Hinkel and Outcalt, 1994), which can lead to both lower and higher ground temperatures. Finally, the
47 thermal and hydrological properties of the subsurface material can strongly influence permafrost distribution. In discontinuous
48 mountain permafrost terrain, the lowest-lying permafrost areas are frequently found in coarse, blocky terrain (Harris and
49 Pedersen, 1998). In particular, rock glaciers are frequently found below the general elevation limit of mountain permafrost
50 (Lilleøren and Etzelmüller 2011).

51 In Southern Norway, the lower limit of mountain permafrost is estimated between 1600 m a.s.l. in the west to 1000
52 m a.s.l. in the east (Etzelmüller et al., 2003), while a similar west-east decrease from 800–1000 m a.s.l. to ca. 300 m a.s.l. in
53 the east is observed in Northern Norway (Gislén et al., 2017). A first Norway-wide inventory of rock glaciers based on aerial
54 imagery was published in 2011 (Lilleøren and Etzelmüller, 2011). The density of rock glaciers is lower than in other mountain
55 permafrost areas which was attributed to a lack of bedrock competence and debris availability as well as to the relative lack of
56 steep topography above the permafrost limit. While this first inventory suggested that active rock glaciers occur only above
57 400 m a.s.l., Lilleøren et al. (2022) recently described rock glaciers near sea level in the area of Hopsfjorden, northern Norway,
58 which feature a limited ice body and are in transition from active to relict. Furthermore, Nesje et al. (2021) presented new
59 evidence for active rock glaciers in southern Norway well below the permafrost limits established in modelling studies
60 (Westermann et al., 2013; Gislén et al., 2017).

61 Rock glaciers play an important role in the hydrological cycle, especially in arid regions like the Andes, where in
62 some areas more water is stored in rock glaciers than in glaciers (Jones et al., 2019; Azócar and Brenning, 2010). The open
63 debris structure can act as a trap for snow and rock glaciers can store significant quantities of ice or liquid water. Rock glaciers
64 studied in Argentina are an important water resource as they release water mainly during periods of drought (Croce and Milana

65 2002). Sustained ground ice melt as a response to climate warming threatens this water source. Additionally, melting of ground
66 ice can lead to slope instability (e.g. Gruber and Haeberli, 2007; Saemundsson et al., 2018; Nelson et al., 2001) and damage
67 to infrastructure (e.g. Arenson et al., 2009).

68 The occurrence of a negative temperature anomaly in coarse, blocky deposits has long been recognized (e.g. Liestøl,
69 1966). Harris and Pedersen (1998) found a negative temperature anomaly of 4 to 7 °C in blocky terrain relative to adjacent
70 mineral sediment in mountains in Canada and China. They summarized four hypotheses that explain these anomalies: (a) The
71 Balch effect; (b) chimney effect; (c) continuous air exchange with the atmosphere when no continuous winter snow cover is
72 present; and (d) evaporation of water and sublimation of ice in the summer. The first three of these driving mechanisms relate
73 to air movement in the blocks, while the last hypothesis links characteristics of the water/ice balance to lower ground
74 temperatures in blocky terrain. In the Norwegian mountains, Juliussen and Humlum (2008) showed that blockfields featured
75 a negative temperature anomaly of 1.3 to 2.0 °C. They state that convection in the blockfields is of low importance in creating
76 the anomaly, while the effect was mainly attributed to rocks protruding into and through the snow cover which leads to an
77 increased heat transfer through the snow cover. Gruber and Hoetzle (2008) presented a simple model for the conductive effect
78 of blocks protruding through the snow cover and showed that the mean annual ground temperature is reduced as a result of a
79 lower thermal conductivity of a blocky layer. Additionally, Juliussen and Humlum (2008) argued that a low soil moisture
80 content in permeable blocky debris (due to subsurface drainage in permeable blocky debris) accelerates active layer refreezing
81 in autumn since less latent heat is liberated compared to soils with higher soil moisture content. Cold winter temperatures can
82 therefore penetrate to deeper layers already in early fall/winter, which may lead to decreased overall winter temperatures.
83 However, in spring, the opposite effect is observed when percolating meltwater refreezes at the bottom of the blocky surface
84 layer, leading to rapid ground warming to 0 °C even in deeper layers (e.g. Juliussen and Humlum, 2008; Hanson and Hoelzle,
85 2004; Humlum, 1997).

86 While many of the mechanisms and processes governing the ground thermal regime of blocky terrain are known, a
87 comprehensive quantitative understanding is still lacking. This is particularly relevant for conceptualization in numerical
88 models which generally do not account for the thermal anomaly of blocky terrain. One-dimensional heat flow models have
89 been used in studies to investigate the effect of climate change on permafrost (e.g. Etzelmüller et al., 2011; Hipp et al., 2012)
90 or to model specific processes in mountain permafrost (e.g. Gruber and Hoetzle, 2008). Since permafrost presence is generally
91 not visible at the surface, numerical models are often used to estimate the permafrost distribution (Harris et al., 2009). However,
92 as most models neither include a transient representation of the subsurface water and ground ice balance (e.g. Westermann et
93 al., 2013) nor reproduce the thermal anomaly in blockfields (e.g. Obu et al., 2019), the resulting permafrost maps likely show
94 biased ground temperatures and permafrost extent in mountain areas.

95 The CryoGrid community model (Westermann et al., 2022) is a simulation toolbox that can calculate ground
96 temperatures and water/ice contents in permafrost environments. It largely builds on the well-established CryoGrid 3 model
97 (Westermann et al., 2016) which has been used in e.g. peat plateaus and palsas (Martin et al., 2021), ice-wedge polygons
98 (Nitzbon et al., 2019) and boreal forests (Stuenzi et al., 2021) and has a broad range of applications, including the representation

99 of lateral drainage regimes (Martin et al., 2019), representation of steep rock walls (Schmidt et al., 2021) and massive ice
100 bodies. In the following, the CryoGrid community model is referred to as “CryoGrid” for simplicity.

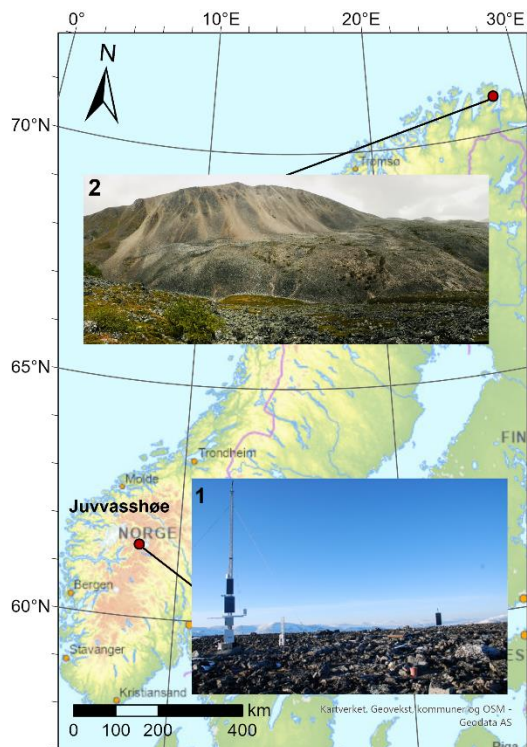
101 In this study, we present CryoGrid simulations of the coupled heat and water/ice balance for blocky terrain in Norway
102 and evaluate the impact of the ground stratigraphy and the drainage regime on ground temperatures. The model is set up with
103 forcing data for two Norwegian permafrost sites, namely a blockfield site in the high mountains in southern Norway and a
104 rock glacier site near sea level in northern Norway. The employed model scheme does not account for air movement and rocks
105 protruding the snow cover as the “classic” causes for the negative thermal anomaly of blocky terrain, but is capable of
106 simulating the seasonal dynamics of the ground ice table in blocky terrain. The goal of the study is to evaluate to what extent
107 the thermal anomaly in blocky terrain can be simulated by such a comparatively simple scheme which could in principle be
108 integrated in larger-scale permafrost modelling and mapping efforts. In particular, we investigate the interplay with the
109 seasonal snow cover and discuss the impact on the permafrost distribution in mountain environments.

110 **2 Study sites**

111 **2.1 Juvvasshøe, southern Norway**

112 Juvvasshøe (61°40 N, 08°22 E, 1894 m a.s.l.) (Fig. 1) is a site located in Jotunheimen in the southern Norwegian mountains,
113 well above the tree line. A 129 m deep borehole was drilled in August 1999 in the PACE (Permafrost and Climate in Europe)
114 project (Harris et al., 2001). Continuous data streams from this PACE borehole are available with the exception of a gap
115 between 21 December 2011 to 24 April 2014. The site is located in an extensive block field on a mountain plateau with sparse
116 vegetation cover. The bedrock (crystalline rocks, Farbrot et al., 2011) is located at approximately 5 m depth, the first meter
117 consists of large stones and boulders and the ground below mainly consists of cobbles (Isaksen et al., 2003). Between 2000
118 and 2004, Isaksen et al. (2007) measured a mean annual air temperature (MAAT) at 2 m height of -3.3 °C. The mean ground
119 temperature at 2.5 m below the surface during this period was -2.5 °C. The mean annual precipitation was estimated to be
120 between 800 and 1000 mm. The site is extremely wind-exposed, resulting in a low snow thickness due to wind drift. Hipp et
121 al. (2012) described a snow depth of less than 20 cm, while the snow thickness in surrounding, lower-lying and less exposed
122 sites can be up to 140 cm. Isaksen et al. (2007) measured the difference between the mean annual ground surface temperature
123 and MAAT, which is the surface offset, at exposed and less exposed sites in this area. At sites with a significant snow cover,
124 the surface offset was more than 2 °C, while at exposed (including Juvvasshøe) sites this offset is generally below 1 °C. The
125 permafrost thickness at the PACE borehole was estimated to be approximately 380 m (Isaksen et al., 2001), with the lower
126 permafrost limited at ca. 1450 m a.s.l. (Farbrot et al., 2011). The thickness of the active layer increased from 215 cm in 1999
127 (Isaksen et al., 2001) to ca. 250 cm in 2019 (Etzelmüller et al., 2020). A weak zero curtain effect suggests a low water content
128 in the active layer (Isaksen et al., 2007). A warming of 0.2 °C per decade and 0.7 °C per decade in surface air temperature and
129 ground surface temperature, respectively, occurred between 2000 and 2019 (Etzelmüller et al., 2020).

130



131

132 **Figure 1: Location of the two sites in Norway** (© Norwegian Mapping Authority). (1) **blockfield at Juvvasshøe (1894 m a.s.l.),** (2)
 133 **rock glacier at Ivarsfjorden (60–160 m a.s.l.).**

134 2.2 Ivarsfjorden rock glacier, northern Norway

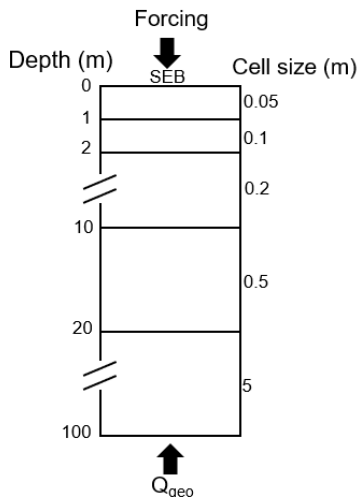
135 Ivarsfjorden is a small fjord arm of the larger Hopsfjorden, located on the Nordkinn peninsula in the Troms and Finnmark
 136 county in northern Norway (Fig. 1). Deglaciated around 14–15 cal kyr BP (calibrated kiloyears before present, Romundset et
 137 al., 2011), the peninsula is dominated by flat mountain plateaus of exposed bedrock, *in situ* weathered material and coarse
 138 grained till (Lilleøren et al., 2022), which feature steep slopes towards the sea. The coastal areas of Finnmark have a wet
 139 maritime climate, with mean annual precipitation around 1000 mm (Saloranta, 2012). Lilleøren et al. (2022) describe a MAAT
 140 of 1.6 °C between 2010 and 2019 in the area of the rock glacier, which lies in a southwest-northeast trending valley at an
 141 elevation extent of roughly 60 to 160 m a.s.l.. The mountain at its east (443 m a.s.l.) serves as the source area with rockfall
 142 debris and coarse talus slopes being common. The bedrock in Ivarsfjorden consists of sandstones and phyllites (NGU, 2008).
 143 Sandstones often generate coarse, bouldery material, which is favorable for the formation of rock glaciers (Haeberli et al.,
 144 2006). The rock glacier in Ivarsfjorden is northwest facing and has previously been interpreted as relict (Lilleøren and
 145 Etzelmüller, 2011), but a detailed analysis showed that a limited ice core might still be present (Lilleøren et al., 2022). A
 146 negative MAAT around 100 to 150 years ago is an indication that rock glaciers in this area were likely active at the end of the
 147 Little Ice Age (LIA). Refraction Seismic Tomography surveys indicate a porous air-filled stratigraphy such as blocky talus
 148 deposits at the near-surface at parts of the rock glacier. While observed mean annual ground surface temperatures between

149 2015 and 2020 are all positive, negative surface temperatures during summer have been observed by a thermal camera at the
150 front slope of the rock glacier. This is likely an indication of the chimney effect and thus of connected voids that support air
151 flow.

152 3. Methods

153 3.1 The CryoGrid community model

154 CryoGrid is a simulation toolbox for ground thermal simulations that can be applied to a wide range of modelling tasks in the
155 terrestrial cryosphere thanks to its modular structure (see Westermann et al., 2022 for details). It is mainly applied in permafrost
156 environments, using the finite difference method to transiently simulate ground temperatures. We use a one-dimensional model
157 column with a domain depth of 100 m (e.g. Westermann et al., 2016; Schmidt et al., 2021) and grid cell sizes increasing with
158 depth (Fig. 2). The lower boundary condition is provided by a constant geothermal heat flux. The upper boundary results from
159 solving the surface energy balance, including both radiative and turbulent heat fluxes, as well as the heat flux in the ground.
160 In order to compute the surface energy balance, atmospheric forcing data are required (Sect. 3.2). To calculate ground
161 temperatures, both conductive heat transfer following Fourier's law and advection of heat with vertically moving water is
162 taken into account (Westermann et al., 2022). The freezing characteristic of subsurface water/ice depends on the soil type,
163 either following Painter and Karra (2014) for sediments, or set to free water (water changes state at 0 °C, Westermann et al.,
164 2022) for subsurface material with large pores/voids, such as blocky terrain. To define the properties of the subsurface material,
165 a stratigraphy of volumetric mineral, organic, water and ice contents and the field capacity (the ability to hold water against
166 gravity) must be provided (Westermann et al., 2022).



167

168 **Figure 2: Schematic of the model grid, indicating cell sizes at different depths and upper and lower boundary conditions. As upper**
169 **boundary condition, the surface energy balance (SEB) forced by near-surface meteorological data is used. The lower boundary**
170 **condition is provided by a constant geothermal heat flux, Q_{geo} .**

171 For soil hydrology, a gravity driven bucket scheme is used (Westermann et al 2022). Rainfall provided by the model
 172 forcing is added to the uppermost grid cell, while evaporation is determined by the surface energy balance calculations (note
 173 that we consider unvegetated surfaces and thus do not account for transpiration). Water that is in excess of the field capacity
 174 infiltrates downwards until either the water table or a non-permeable layer, such as a frozen grid cell is reached. If all grid cells
 175 are saturated, excess water is removed as surface runoff. We use a one-dimensional model setup, but simulate lateral drainage
 176 of water by introducing a seepage face, i.e. a lateral boundary condition for water fluxes representing flow between the
 177 saturated grid cells of the model domain and a stream channel (or the atmosphere) to which the water can freely flow out from
 178 the subsurface (e.g. Scudeler et al., 2017). Using the elevation of the water table, z_{wt} (computed as the elevation of the
 179 uppermost saturated grid cell), lateral water fluxes F_i^{lat} are derived for all saturated unfrozen grid cells i below the water table
 180 (i.e. at elevations $z_i < z_{wt}$) as

$$181 \quad F_i^{lat} = -K_H \frac{z_{wt} - z_i}{d^{lat}}, \quad (1)$$

182 where K_H is the saturated hydraulic conductivity, d^{lat} is the lateral distance to the seepage face and the flux is determined by
 183 the difference between the hydrostatic potential (proportional to z_{wt}) of the water column and the gravitational potential of
 184 free water at the elevation of each cell (proportional to z_i). Note that Eq. (1) is an approximation for small changes of the water
 185 table and small outflow fluxes for which the potential in the saturated zone can be approximated by the hydrostatic potential.
 186 The parameter d^{lat} is used to control the strength of the drainage, with small distances resulting in a well-drained column,
 187 while high values lead to suppressed drainage. In this study, we consider the two confining cases with a small and large value
 188 of d^{lat} , respectively (Sect. 3.3). In the former, water from rain or ground ice melt is removed rapidly, effectively preventing
 189 the soil water from pooling up, while drainage is negligible in the former, so that the setup corresponds to a classic one-
 190 dimensional model scheme.

191 The snow model used in this study was introduced by Zweigel et al. (2021) and is based on the Crocus snow scheme
 192 (Vionnet et al., 2012) which accounts for snow microphysics and is designed to reproduce a realistic snow pack structure (see
 193 Vionnet et al., 2012 for defining equations; Zweigel et al., 2021 for implementation in CryoGrid). Snowfall is added with
 194 density and microphysical properties derived from model forcing data, in particular air temperature and wind speed. The snow
 195 density evolves due to compaction by the overburden pressure of overlying snow layers, as well as wind compaction and
 196 refreezing of melt- and rainwater (Vionnet et al., 2012). The amount of snowfall from the forcing data can be adjusted by a so-
 197 called *snowfall factor*, sf , with which the snowfall rate from the model forcing is multiplied. With this, the effects of wind-
 198 induced snow redistribution on ground temperatures can be represented at least phenomenologically (Martin et al., 2019),
 199 using $sf < 1$ for areas with net snow ablation and $sf > 1$ for areas with net deposition.

200 **3.2 Downscaling of model forcing**

201 The meteorological data used to force the CryoGrid model were generated by applying TopoSCALE, a topography-based
202 downscaling routine (Fiddes and Gruber, 2014), to ERA5 reanalysis data (Hersbach et al., 2020). TopoSCALE is employed in
203 cryosphere applications in complex terrain, including estimating mountain permafrost distribution (Fiddes et al., 2015), snow
204 data assimilation (Aalstad et al., 2018; Fiddes et al., 2019), and downscaling regional climate model output (Fiddes et al.,
205 2022). ERA5 output is provided as interpolated point values on a regular latitude-longitude grid at a resolution of 0.25° at an
206 hourly frequency, both at the surface level, corresponding to Earth's surface as represented in the reanalysis, and at 37 pressure
207 levels in the atmosphere from 1000 to 1 hPa. We obtained data for the entire reanalysis period, from 1951 to 2019, and
208 converted this into a moving three-hourly average, which is the temporal resolution that the model is run at. As input to
209 TopoSCALE, we obtained from the surface level: 2 m air and dewpoint temperature, 10 m meridional (northward) and zonal
210 (eastward) wind velocity components, surface pressure, constant surface geopotential, incoming longwave radiation, incoming
211 shortwave radiation, and total precipitation. From the pressure levels we acquired: air temperature, specific humidity, zonal
212 and meridional wind velocity components, and dynamic geopotential. For Juvvasshøe we used all levels in the range 900 hPa
213 to 700 hPa, while for the lower elevation Ivarsfjorden rock glacier we used all levels between 900 hPa and 1000 hPa. To
214 account for terrain shading in the downscaling routine, a digital elevation model (DEM) is required, for which we use the
215 mosaic version of the ArcticDEM with a resolution of 32 m (Porter et al., 2018) at both sites. TopoSCALE delivered all
216 meteorological forcing data required to run CryoGrid: near surface air temperature, specific humidity, wind speed, incoming
217 longwave radiation, incoming shortwave radiation, as well as snowfall and rainfall.

218 **3.3 Model setup**

219 Three idealized ground stratigraphies are set up in order to investigate the effect of water drainage on the ground thermal
220 regime and ground ice dynamics in blocky terrain. These are referred to as the *blocks only*, *blocks with sediment* and *sediment*
221 *only* stratigraphies (Table 1) in the following. The *blocks only* stratigraphy consists of a coarse block layer with 50% porosity
222 and air-filled voids which is assigned low field capacity of 1% (Table 1), i.e. the surfaces of the coarse blocks retain only little
223 water. This idealized stratigraphy is designed to represent an active rock glacier where finer sediments resulting from
224 weathering and erosion processes are transported towards the tongue of the rock glacier. Furthermore, Dahl (1966) observed
225 that blockfields on slopes more often do not contain a fine sediment fraction between the blocks in northern Norway, so that
226 the *blocks only* stratigraphy can also represent active blockfields. The second stratigraphy, *blocks with sediment*, is designed
227 to represent blocky terrain where the voids are filled by finer sediments. This is often observed in blockfields on more flat
228 surfaces, which are more likely to retain finer sediment within their pores (as in Isaksen et al., 2003 and Dahl 1966). We again
229 consider coarse blocks with 50% porosity (as for the *blocks only* stratigraphy), but as the voids are filled with fine sediments
230 (which again are assumed to have 50% porosity), the overall porosity is only 25%. Furthermore, a significantly higher field
231 capacity than for the *blocks only* stratigraphy is assigned as more water can be held in the finer pores of the sediment fraction.

232 Finally, the *sediment only* stratigraphy serves as a control scenario for a soil without blocks. It contains sediment with 50%
 233 porosity and a high field capacity due to the water holding capacity of the fine-grained sediment material. For all stratigraphies,
 234 bedrock (3% porosity and saturated conditions, e.g. Hipp et al., 2012; Fabrot et al., 2011) is assumed below 5 m depth, which
 235 is in line with observations from Isaksen et al. (2003) at Juvvasshøe. Finally, none of the stratigraphies contain soil organic
 236 matter. We emphasize that the stratigraphies are in qualitative agreement with field observations of air and sediment-filled
 237 block layers in Norway, but the assumed porosities of 50% for both the block layer and the sediments represent idealized
 238 scenarios. However, we perform a sensitivity analysis for different porosity values (Table S1 in the Supplement) to investigate
 239 the impact of this parameter on the simulation results.

240

241 **Table 1: Mineral content, porosity, field capacity (all in vol. fraction) and soil freezing characteristic for the three idealized**
 242 **subsurface stratigraphies.**

Name	mineral	porosity	field capacity	soil freezing characteristic
<i>Blocks only</i>	0.5	0.5	0.01	Free water
<i>Blocks with sediment</i>	0.75	0.25	0.15	Free water
<i>Sediment only</i>	0.5	0.5	0.25	Sand

243

244 For the geothermal heat flux lower boundary condition, a value of 0.05 Wm^{-2} is used, which is a typical value for Norway used
 245 in previous modelling studies (Westermann et al., 2013).

246 To investigate the effect of subsurface drainage on ground temperatures and ground ice conditions, we distinguish *undrained*
 247 and *drained* scenarios by using two different values of d^{lat} (Eq. 1) for in the idealized stratigraphies. A d^{lat} value of 10^4 m is
 248 used for *undrained* cases, which emulates conditions at a flat surface, resulting in a to a good approximation one-dimensional
 249 water balance, where only surface water is removed. For the *drained* cases, a d^{lat} value of 1 m is used, which results in well-
 250 drained conditions which are typical in sloping terrain. For the saturated hydraulic conductivity K_H , a fixed value of 10^{-5} m s^{-1}
 251 is used for all stratigraphies, although the true hydraulic conductivities almost certainly differ between stratigraphies.
 252 However, the key parameter controlling lateral water fluxes in Eq. 1 is in reality the “drainage timescale” $K_H/d^{lat} [\text{s}^{-1}]$, which
 253 is varied by four orders of magnitude between $K_H/d^{lat} = 10^{-5} \text{ s}^{-1}$ ($d^{lat} = 1 \text{ m}$, well-drained conditions) and $K_H/d^{lat} = 10^{-9} \text{ s}^{-1}$
 254 ($d^{lat} = 10^4 \text{ m}$ undrained conditions). As the study setup is designed to analyze these two “confining cases”, it is sufficient to
 255 only vary d^{lat} and leave K_H constant for simplicity. Further sensitivity tests for d^{lat} and K_H are provided in Table S2 in the
 256 Supplement. With the exception of the *snowfall factor* (see Sect. 3.3.1 to 3.3.3), the parameters in the snow model are kept
 257 constant in all model runs, using a surface emissivity of 0.99, a roughness length of 10^{-3} m , a saturated hydraulic conductivity
 258 of 10^{-4} m s^{-1} and a field capacity of 0.05 (Westermann et al., 2022). For the ground surface, we used an albedo of 0.15,
 259 emissivity of 0.99, and a roughness length of 10^{-3} m .

260 We perform three types of model simulations which differ in their overall purpose. For *validation* runs (Sect. 3.3.1),
 261 we adjust subsurface stratigraphy and *snowfall factor* in order to compare model results with the available field measurements
 262 from the two study sites. *Equilibrium* runs (Sect. 3.3.2) and *transient* runs (Sect. 3.3.3) are designed to explore the sensitivity
 263 of the simulated ground thermal regime towards the three idealized stratigraphies (Table 1) and the two drainage cases. An
 264 overview of the basic settings of the different simulation types is provided in Table 2.

265

266 **Table 2: Overview of basic model settings for the different simulation types. A spin-up of subsurface temperatures is achieved by**
 267 **repeated simulations for the spin-up period (until a stable temperature profile is reached), before the actual model run for the**
 268 **simulation period is conducted. “Idealized” stratigraphy and drainage refers to three subsurface stratigraphies (Table 1) combined**
 269 **with two types of drainage conditions. See Sect. 3.3.1 to Sect. 3.3.3 for details.**

Simulation type	Site	Spin up period	Simulation period	Stratigraphy and drainage	Snowfall factor
<i>Validation</i>	Juvvasshøe	1951-2010	2010-2019	Best-fit	0.25
	Ivarsfjorden	1951-2016	2016-2019	Best-fit	1
<i>Equilibrium</i>	Juvvasshøe	2000-2010	2000-2010	Idealized	0.0, 0.25, 0.5, 0.75, 1.0, 1.5
	Ivarsfjorden	1962-1971	1962-1971	Idealized	0.0, 0.25, 0.5, 0.75, 1.0, 1.5
<i>Transient</i>	Juvvasshøe	1962-1971	1951-2019	Idealized	0.25
	Ivarsfjorden	1962-1971	1951-2019	Idealized	1

270

271 3.3.1 Validation runs

272 As a prerequisite for conducting model experiments on ground stratigraphy and drainage (Sects. 3.3.2, 3.3.3), validation runs
 273 are set up to show that the model can reproduce key characteristics of the thermal regime at the two sites in a satisfactory
 274 manner (based on available observations). Furthermore, we use the observations to determine the best-fitting *snowfall factor*
 275 for the two sites which is subsequently used in the transient runs (Sect. 3.3.3). At Juvvasshøe, temperature measurements in a
 276 borehole are available from 2000 to 2019, allowing a comparison at different depths. At the Ivarsfjorden rock glacier site,
 277 observations of ground temperature at deeper depths are lacking, but measurements of near-surface ground temperatures are
 278 available from July 2016 to July 2019 (Lilleøren et al., 2022). These are compared to simulation results to ensure that the
 279 model reproduces the observed surface offset between air and ground surface, largely caused by the winter snow cover (e.g.
 280 Martin et al., 2019; Schmidt et al., 2021). At both sites, the model is ran for the entire period of available forcing data, leaving
 281 at least 60 years for the model spin-up which is sufficient to analyze ground temperatures in uppermost meters of the ground
 282 column.

283 Manual adjustment of the ground stratigraphy (porosity and thus mineral content) and snowfall factor are performed
284 until a good fit with daily measurements is achieved. At Juvvasshøe, based on observations of blocks and smaller cobbles with
285 finer sediments down to the onset of bedrock at a depth of 5 m (Isaksen et al., 2003), the *blocks with sediment* stratigraphy is
286 used as a starting point to vary porosities until a good fit is achieved. As this site is extremely exposed to wind and most snow
287 is blown away (Isaksen et al., 2003; Westermann et al., 2013), the snowfall factor is stepwise decreased to values below one
288 to improve the model performance. At Ivarsfjorden, we considered 11 temperature loggers within the rock glacier outline (Fig.
289 1d in Lilleøren et al., 2022), of which all except for one are placed on the relict surface of the rock glacier (Fig. 2a in Lilleøren
290 et al., 2022). On the relict surface, deposition of finer sediment in between blocks is more likely than on the active surface,
291 due to the lack of movement. Here, the *blocks with sediment* stratigraphy is considered appropriate and used as starting point
292 for the calibration. At both sites, the root-mean-square-error (RMSE) and bias are calculated in order to provide an objective
293 measure of the model fit. At Juvvasshøe this was accomplished for daily values at 0.4 m and 2 m depth, while at Ivarsfjorden
294 the mean daily ground surface temperature of the loggers within the rock glacier outline is used.

295 **3.3.2 Equilibrium runs**

296 The goal of equilibrium runs is to investigate the sensitivity of the ground thermal regime towards ground properties and
297 drainage conditions, using both the *undrained* and *drained* setup for the three idealized stratigraphies (Table 1) which results
298 in six scenarios. As the heavily wind-affected snow cover is a key source of spatial variability in ground temperatures in the
299 Norwegian mountains (Gisnås et al., 2014; Gisnås et al., 2016), the model is run for a range of snowfall factors between 0.0
300 and 1.5 (Table 2) for each scenario. This analysis allows us to identify the magnitude of the thermal anomaly that the subsurface
301 drainage induces at various amounts of snow, as well as estimate the threshold snow amount for permafrost existence in the
302 six scenarios. This analysis is performed for equilibrium conditions for 10 year periods of roughly stable climate, which is
303 iterated three times until a steady state temperature profile of the uppermost 5 meters is established. For Juvvasshøe, the period
304 2000 to 2010 is selected as the model can be initialized with real-time borehole data. For Ivarsfjorden, the comparatively cold
305 period 1962 to 1971 is selected as this relatively stable period is the coldest period in the available forcing data and thus the
306 closest to Little Ice Age climate conditions, when the Ivarsfjorden rock glacier was very likely active (Lilleøren et al., 2022).

307 **3.3.3 Transient runs**

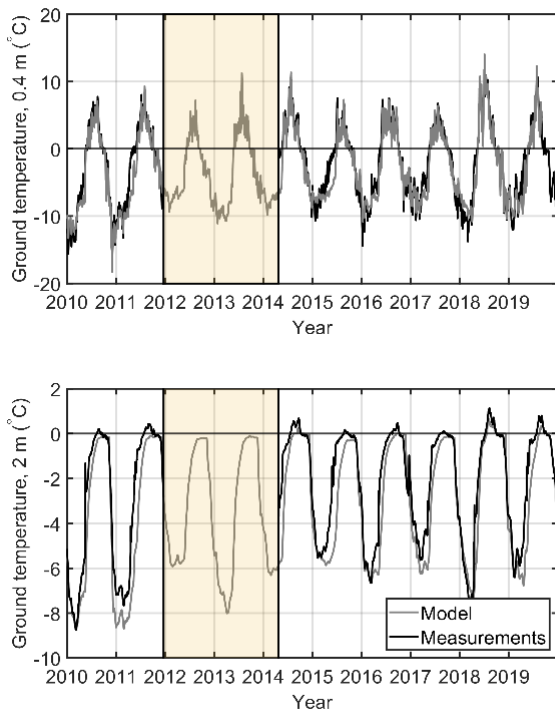
308 The goal of the transient runs is to analyze the effect of ground stratigraphies and drainage conditions on the transient response
309 of ground temperatures and ice tables to climate warming. For this purpose, we perform model simulations from 1951 to 2019,
310 when air temperatures have increased by more than 1 °C in Norway. To initialize simulations, we perform a model spin-up by
311 iterating three times over the coldest 10 year period in the forcing data (1962 – 1971) which is sufficient to achieve a stable
312 ice table. This is the same period as in the equilibrium runs (see Sect. 4), for which it was selected to capture permafrost
313 conditions at the Ivarsfjorden rock glacier site (see Sect. 4). Thus, the transient runs allow us to analyze the evolution of the
314 permafrost towards the warming of the recent decades. We only use the best fitting snowfall factor (Table 2), as derived from

315 the validation runs (Sect. 3.3.1), but again perform simulations for the three idealized stratigraphies and *undrained* and *drained*
316 conditions. This way, we can evaluate whether different ground stratigraphies or drainage conditions lead to different warming
317 rates of ground temperatures, as well as different thresholds for permafrost thaw.

318 4. Results

319 4.1 Comparison to in-situ measurements

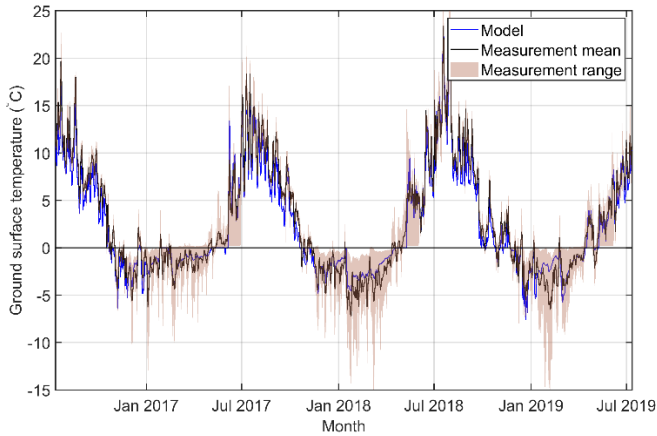
320 The results of the *validation runs* at Juvvasshøe are compared with measured daily ground temperatures at the PACE borehole
321 (Etzelmüller et al, 2020). Figure 3 shows the comparison of measured ground temperatures with modelled temperatures at 0.4
322 and 2.0 m depth for the best fitting model configuration. The snowfall factor for this model setup is 0.25, i.e. incoming snowfall
323 is reduced by 75% in order to capture the effect of snow ablation due to wind drift. This resulted in mean annual maximum
324 snow depths of 34 cm, in broad agreement with observations from the site (Iskasen et al., 2003) and earlier modeling studies
325 at the site (Westermann et al., 2013). The subsurface stratigraphy for this model configuration is highly similar to the *blocks*
326 *with sediment* stratigraphy, but with a slightly lower porosity of 0.2 (i.e. a volumetric mineral content of 0.8). This would for
327 example correspond to blocks and cobbles with a porosity of 0.4 (0.5 for *blocks with sediment*), filled with fine sediments with
328 a porosity of 0.5 (and field capacity 0.25), which is plausible given the broad characteristics of the observed borehole
329 stratigraphy (Isaksen et al., 2003). This configuration used *drained* conditions, although differences with *undrained* conditions
330 are minimal for this stratigraphy. For daily temperatures at 0.4 m depth, the RMSE and bias are 2.1 °C and -0.6 °C, respectively,
331 while they are 1.2 °C and -0.7 °C at 2 m depth. There is a mismatch in the timing of spring temperatures at 2 m depth in several
332 years, for which modelled temperatures increase later than measured values. This is likely a result of differences in the snow
333 melt, as the snowpack dynamics resulting from wind redistribution is not completely captured by the snowfall scaling with a
334 constant snowfall factor (e.g. Martin et al., 2019). Furthermore, the uppermost 1 m contain large stones and boulders, while
335 the layer below is characterized by smaller stones and cobbles (Isaksen et al., 2003), so that a ground stratigraphy with two
336 layers in the uppermost 5 m may further improve the performance of the simulations.



337

338 **Figure 3: Modelled and measured ground temperature at the PACE borehole in Juvvasshøe at 0.4 m (upper) and 2.0 m (lower)**
 339 **depth. The shaded area indicates a period when no borehole data are available.**

340 At the rock glacier in Ivarsfjorden, a comparison between modelled and measured temperature is performed for
 341 average daily ground surface temperatures, using the mean of the measurements at 11 sites within the rock glacier as target for
 342 the comparison (Fig. 4). The best-fitting model configuration was found to be the *blocks with sediment* stratigraphy and a
 343 snowfall factor of 1.0, resulting in an RMSE of 1.3 °C and a bias of -0.4 °C. As in Juvvasshøe, the configuration used *drained*
 344 conditions, while differences with *undrained* conditions are small. Figure 4 also shows the significant spatial variability of
 345 ground surface temperatures, which is particularly large in winter. Also in periods, when the simulation results and the mean
 346 of the measurements visibly deviate, the simulations remain within the range of the measurements. While there are some
 347 deviations between the observations and simulation results at both Juvvasshøe and Ivarsfjorden, we conclude that the model
 348 setup, including the model forcing, can capture the general ground surface temperature regime at both sites which is a
 349 prerequisite for obtaining meaningful results from the *equilibrium* and *transient runs*.

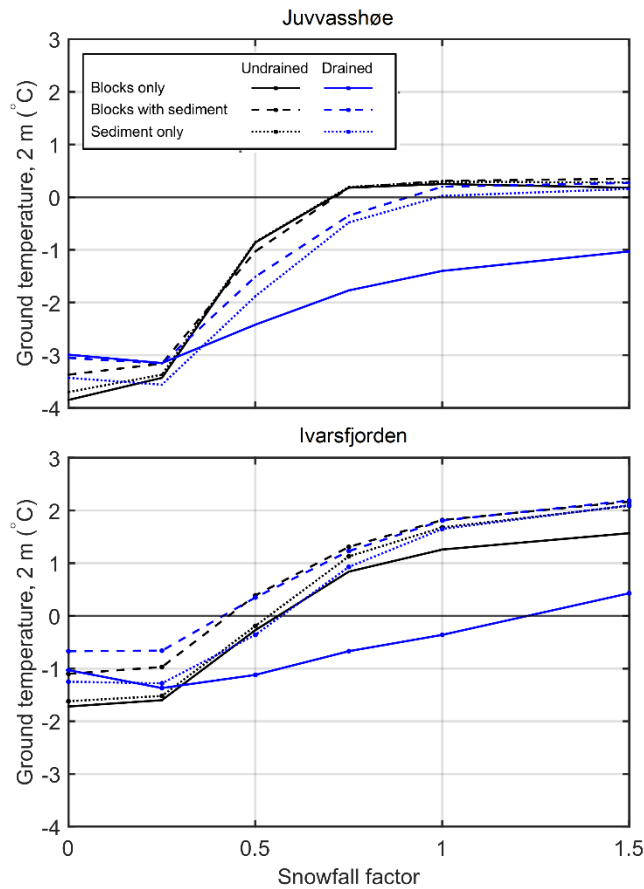


350

351 **Figure 4: Daily modelled and measured ground surface temperatures in Ivarsfjorden from July 2016 to July 2019. The shaded area**
 352 **indicates the minimum to maximum range of measured daily values from 11 loggers (based on Lilleøren et al., 2022), while the black**
 353 **line represents the mean value of all loggers.**

354 4.2 Equilibrium ground temperatures and sensitivity to snow

355 Figure 5 shows the average ground temperature at 2 m depth for the three stratigraphies, the *drained* and the *undrained*
 356 scenario, and different snowfall factors at both sites. At both sites there is a clear pattern of lower temperatures in the *blocks*
 357 *only, drained* scenario (solid blue line) compared to all five other scenarios. For snowfall factors of 0.75 and larger, the
 358 difference in ground temperature between *blocks only, drained* and the other scenarios is in the range of 1.1 °C and 1.8 °C at
 359 Juvvasshøe and in the range of 1.1 °C and 2.2 °C at Ivarsfjorden. This shows that the magnitude of the negative thermal
 360 anomaly increases with a larger amount of snowfall. Results of the sensitivity study to porosity of the soil (Table S1 in the
 361 Supplement) show that mean ground temperatures are within 0.4 °C between the highest and lowest porosity tested.



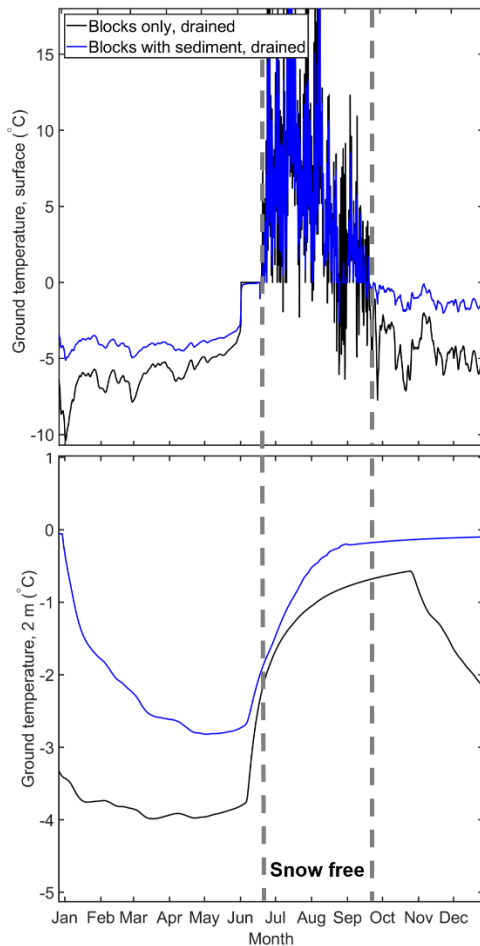
362

363 **Figure 5: Equilibrium ground temperature at 2 m depth for three idealized stratigraphies (Table 1) and different snowfall factors.**
 364 **Each data point represents one model run of one of the six scenarios at a certain snowfall factor.**

365 Annual maximum snow depths at a snowfall factor of 1.0 are between 1.5 m and 2.4 m at Juvvasshøe and between
 366 0.4 m and 1.0 m at Ivarsfjorden. At Juvvasshøe, all three *undrained* scenarios feature positive ground temperatures at snowfall
 367 factors of 0.75 and above, which corresponds to permafrost-free conditions. Temperatures in the *blocks with sediment, drained*
 368 and *sediment only, drained* runs are positive for a snowfall factor of 1.0 and above. The ground temperature in the *blocks only,*
 369 *drained* runs remains below -1.0 °C for all snowfall scenarios. A similar pattern is seen in Ivarsfjorden, although a snowfall
 370 factor of 1.5 results in positive temperatures for the scenario *blocks only, drained* which is clear evidence of the overall warmer
 371 ground temperatures. Temperatures for the *blocks with sediment* stratigraphy are positive for snowfall factors exceeding 0.5,
 372 and exceeding 0.75 for the other scenarios (with the exception of the *blocks only, drained* scenario, see above). For snowfall
 373 factors above 0.25, ground temperature at 2 m depth increase with snow depth as a result of increased insulation of the ground
 374 during winter. However, the increase from a snowfall factor of 0 to 0.25 leads to a slight cooling for the *drained* scenarios as
 375 opposed to a slight warming in the *undrained* scenarios. The reason for this cooling is likely the higher winter albedo of the

376 completely snow-free ground (for snowfall factor zero), which outweighs the insulating of the shallow snow cover for snowfall
377 factor 0.25.

378 Figure 6 shows simulated temperatures for one year at the ground surface and 2 m depth for drained conditions for
379 both *blocks only* and the *blocks with sediment* scenarios for Juvvasshøe (snowfall factor 0.75). While ground surface
380 temperatures are largely similar during the snow-free summer season, they decrease much faster in fall for the *blocks only*
381 compared to the *blocks with sediment* scenario, for which the slow refreezing of the active layer leads to a prolonged warming
382 of the ground surface from below. In the *blocks only* scenario, on the other hand, the active layer contains only little water, so
383 that refreezing occurs within only a short time period. The rapid cooling in the *blocks only* scenario is also visible within the
384 permafrost table at 2 m depth, resulting in lower winter temperatures compared to the *blocks with sediment* curve and thus
385 explaining the simulated differences in mean ground temperature.

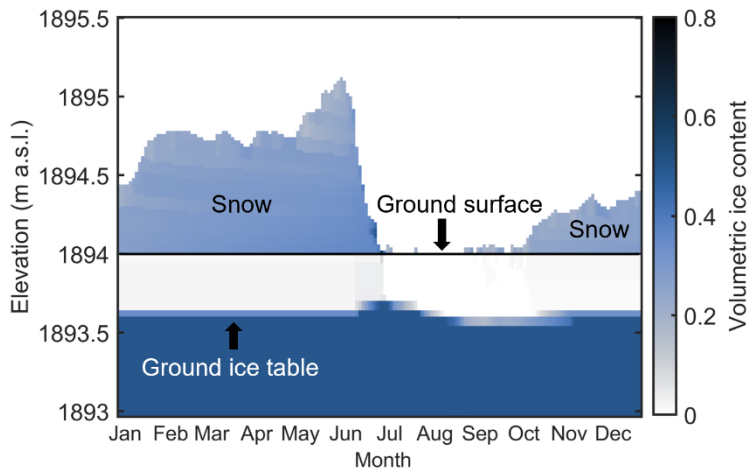


386

387 **Figure 6: Modelled ground temperature at 0.05 m (top) and 2 m (bottom) depth for the *blocks only, drained* and *blocks with sediment,***
388 ***drained* scenarios during a year of an equilibrium run at Juvvasshøe for $sf = 0.75$. The snow-free summer season is highlighted. Note**
389 **that the upper plot is truncated at 17 °C, maximum summer temperatures are 26 °C in both scenarios.**

390

391 Figure 7 shows the corresponding snow cover and volumetric ground ice content in the upper meter of the ground for
392 the *blocks only, drained* scenario. A largely stable ground ice table forms already at a depth of about 0.5 m, while the active
393 layer is almost free of ground ice in winter, corresponding to the low water contents for thawed condition, enabling rapid
394 refreezing and thus strong cooling during winter. During and after snow melt, meltwater infiltrates in the blocky layer and
395 refreezes at the then very cold ice table, resulting in the formation of new ground ice which slowly melts during the course of
396 summer. The slight increase of the ground ice table in early winter is due to refreezing of residual water above the ice table
397 from rain and snow melt events in October which has not fully drained before refreezing.



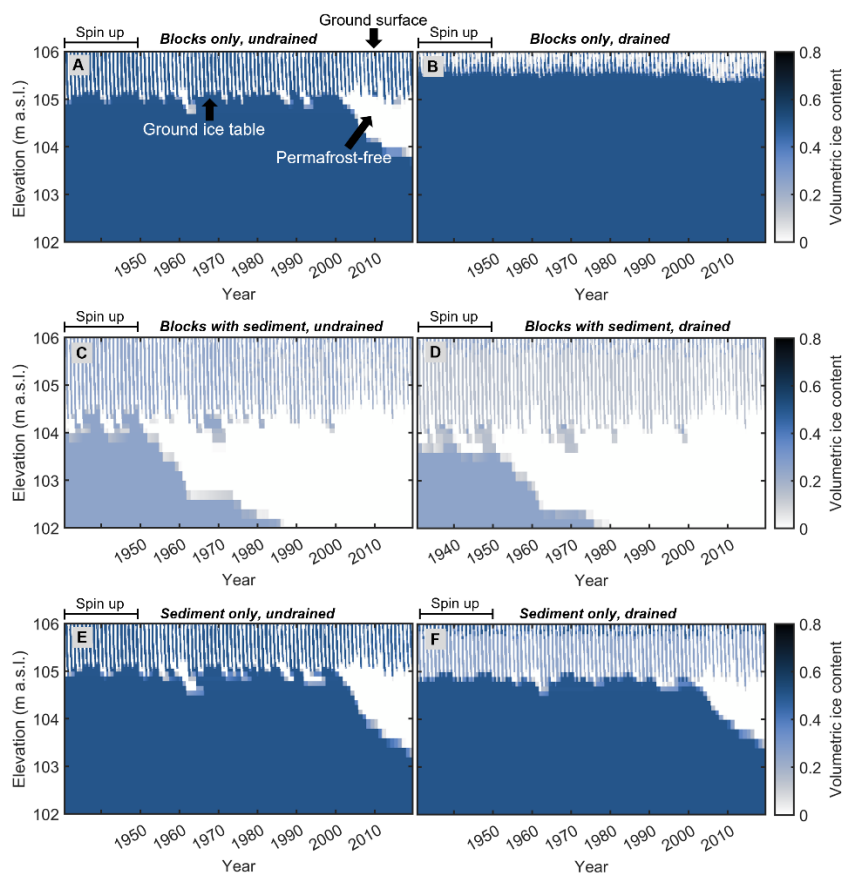
398

399 **Figure 7: Modelled volumetric ground ice content in the upper 1 m of the ground (below 1894 m a.s.l.) and the snow cover (above**
400 **1894 m a.s.l.) for the *blocks only, drained* scenario, during one year of an equilibrium run at Juvvasshøe for $sf = 0.75$. Note the rise**
401 **of the ground ice table, here defined as the uppermost cell where ground ice persists for two or more years, in June after infiltrated**
402 **snow melt water refreezes.**

403 4.3 Transient response to climate warming

404 The ERA5 reanalysis dataset allows us to simulate the evolution of the ground thermal regime and ground ice content from
405 1951 to 2019, during which mean air temperatures increased from $-4.5\text{ }^{\circ}\text{C}$ (1951-1960) to $-3.8\text{ }^{\circ}\text{C}$ (2010-2019) for Juvvasshøe
406 and from $0.5\text{ }^{\circ}\text{C}$ (1951-1960) to $1.2\text{ }^{\circ}\text{C}$ (2010-2019) at Ivarsfjorden. Figure 8 shows the ground ice content for different
407 scenarios in Ivarsfjorden. In all simulations, a stable ice table and permafrost conditions form during the spin up period (using
408 model forcing for the cold period 1962-1971, Table 2), with volumetric ice contents of 0.5 (*blocks only, sediment only*) and of
409 0.25 (*blocks with sediment*) according to the applied stratigraphy (Table 1). In the period 1951 to 2019, ground ice content
410 evolves as a response to the applied climate forcing, showing different responses of the ground ice table. In the *blocks only*,
411 *drained* scenario, the ice table in the upper 5 m (so between the active layer and the bedrock) does not lower by a significant
412 amount (0.1 m lowering), while the ice table lowers by 1.2 m in the *blocks only, undrained* scenario. The ice table in the *blocks*

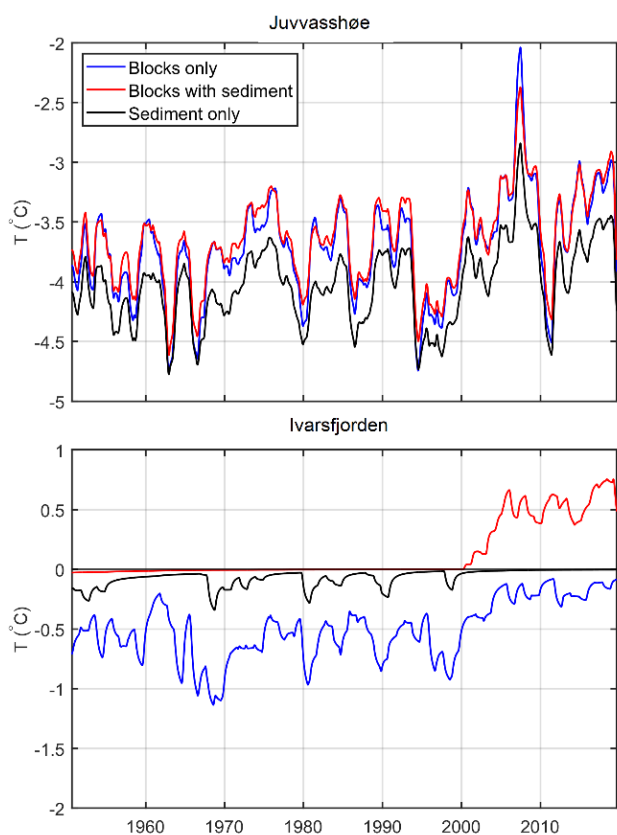
413 with sediment stratigraphy disappears by 1985 and 1975 in the *undrained* and *drained* scenarios, respectively. Finally, the
 414 *sediment only* simulations show an intermediate effect where the ice table has lowered by 1.7 m and 1.6 m for *undrained* and
 415 *drained* conditions respectively by 2019. The complete degradation in the *blocks with sediment* runs compared to partial
 416 degradation in all other scenarios (except *blocks only, drained*) is not unexpected since this stratigraphy has a 25% porosity
 417 (and thus ice content), compared to 50% in the others. We conclude that the ground stratigraphy and drainage conditions
 418 strongly control the response of the ground towards warming, with full degradation near-surface permafrost in both *blocks*
 419 *with sediment* runs, partial degradation in the *blocks only, undrained* run and in both *sediment only* runs and finally continued
 420 stable permafrost conditions in the *blocks only, drained* simulation. At the Juvvasshøe site, the ice table remain stable in all
 421 simulations, but a slight lowering occurs in the *blocks with sediment* scenarios.



422

423 **Figure 8: Modelled volumetric ground ice content at Ivarsfjorden between 1951 and 2019 for the idealized stratigraphies in**
 424 ***undrained* and *drained* conditions and $sf = 1.0$. Only the subsurface domain is shown, with the ground surface elevation at 106 m**
 425 **a.s.l.. In the active layer, ice contents increase and decrease annually, corresponding to the active layer refreezing and thawing. The**
 426 **ground ice table is defined as the uppermost cell where ground ice persists for more than two years and follows the permafrost table.**

427 Figure 9 shows the change in temperatures at 5 m depth for the *drained* scenarios, which at the Ivarsfjorden rock glacier
428 (snowfall factor 1.0) correspond to a full (*blocks with sediment*) and partial (*sediment only*) lowering of the ice table, as well
429 as a relatively stable (*blocks only*) ice table. The *blocks only* simulation shows an increase from $-0.6\text{ }^{\circ}\text{C}$ to $-0.2\text{ }^{\circ}\text{C}$ between the
430 1951–1960 and 2010–2019 means, not being strongly influenced by latent heat effects due to the relative stable ice table. The
431 *sediment only* case experiences only minimal warming, as it is strongly influenced by the ongoing ground ice melt which
432 confines ground temperatures to close to $0\text{ }^{\circ}\text{C}$. Finally, the complete disappearance of ground ice in *blocks with sediment* run
433 coincided with in a warming to positive temperatures, from $0.0\text{ }^{\circ}\text{C}$ to $0.6\text{ }^{\circ}\text{C}$. At Juvvasshøe, permafrost degradation and thus
434 strong ground ice melt does not occur for any of the scenarios for snowfall factor 0.25, and ground temperatures only increased
435 by $0.2\text{ }^{\circ}\text{C}$ (*blocks only*) to $0.4\text{ }^{\circ}\text{C}$ (*blocks with sediment* and *sediment only*) between the 1951–1960 and 2010–2019 means
436 (Fig. 8). The results of the transient runs indicate that the subsurface stratigraphy and drainage conditions strongly affect the
437 timing of permafrost degradation in blocky terrain. While ground ice melt controls the warming rates at Ivarsfjorden, only
438 small differences in warming rates are simulated for the still stable permafrost in Juvvasshøe. However, we emphasize that the
439 simulations at Juvvasshøe were performed for a shallow snow cover ($sf = 0.25$) for which differences in modeled ground
440 temperatures are small (Fig. 5).



441

442 **Figure 9: Ground temperature at 5 m depth for the idealized stratigraphies under *drained* conditions; $sf = 1.0$ for Ivarsfjorden and**
443 **$sf = 0.25$ for Juvvasshøe.**

444 **5. Discussion**

445 **5.1 Limitations of the model setup**

446 In this study, CryoGrid has been applied at two permafrost sites in Norwegian mountain environments. At both sites, we set
447 up validation runs to benchmark the performance of model system against measurements of ground (Juvvasshøe) and ground
448 surface (Ivarsfjorden) temperatures. At Juvvasshøe, the model can largely reproduce the annual cycle of measured ground
449 temperatures at the PACE borehole, when the snowfall is reduced to account for the generally shallow snow cover at the site.
450 At Ivarsfjorden, simulations with full snowfall yielded a similar performance for the ground surface temperature,
451 approximately reproducing the mean of measurements at 11 sites. A statistical evaluation at both sites indicated a cold bias of
452 the model of approximately -0.5 °C which we considered acceptable, considering the spatial variability of the ground thermal
453 regime at both sites (see Gissnås et al., 2014 for Juvvasshøe). At Ivarsfjorden, the transient simulations are in broad agreement
454 with observations at the rock glacier which indicate that permafrost has been present in the recent past (Lilleøren et al., 2022).
455 Permafrost conditions are simulated for all stratigraphies during model spin using the cold period 1962-1971 for which
456 temperatures are closest to Little Ice Age conditions when the rock glacier was likely active.

457 Within the model setup, in particular the exact ground stratigraphy and other poorly constrained parameters, such as
458 the albedo, give rise to uncertainties. While the real porosity of the ground is unknown, sensitivity tests show a maximum of
459 0.4 °C differences in simulated ground temperatures between the highest and lowest porosity values tested (Table S1 in the
460 Supplement). Only at Juvvasshøe, the stratigraphy has been described from the borehole (Isaksen et al., 2003), while no
461 thorough evaluation of the subsurface stratigraphy is available for Ivarsfjorden. Lilleøren et al. (2022) described the site as a
462 complex creeping system with inhomogeneous subsurface properties. Most of the rock glacier surface is described as ‘relict’
463 (Lilleøren et al., 2022) with sand and gravel in between blocks. For these ‘relict’ areas, the simulations for the *blocks with*
464 *sediment* stratigraphy, in which near-surface permafrost fully or partially degrades, could indeed represent the thermal state
465 adequately. This is supported by the validation run with the *blocks with sediment* stratigraphy which yielded a good fit with
466 ground surface temperature measurements at sites largely located on this ‘relict’ surface (Lilleøren et al., 2022). Two areas are
467 described as ‘fresh’ which could indicate lateral movements due to the presence of ground ice. These contain larger blocks and
468 could thus be better described by the *blocks only* stratigraphy for which permafrost and ground ice still persist at the end of the
469 simulations. However, also in these ‘fresh’ areas, the amount of finer sediment is unclear, in particular in deeper layers. In our
470 simulations, we have only considered a single, homogeneous layer in the uppermost 5 m in order to compare the thermal
471 regime and ground ice dynamics for idealized stratigraphies. In reality, ground stratigraphies in blocky terrain can feature
472 aspects of all scenarios, for example a blocky layer with air-filled voids on top, followed by blocks filled with sediments and
473 a sediment only layer in the bottom. For the cooling effect described in this study, it is critical that the blocky top layer is deep

474 enough so that a ground ice table from which water can drain can form within. Therefore, it is likely that also shallower blocky
475 layers with air-filled voids can lead to lower ground temperatures, depending on the climatic conditions which determine the
476 depth of the ground ice table.

477 We emphasize that a consistent model setup was selected for all scenarios, so that uncertainties caused by other parts
478 of the model system influence them all in a similar, consistent way. In particular, none of the convective processes summarized
479 by Harris and Pedersen (1998) that cause a negative thermal anomaly in blocky terrain are considered in the model setup. The
480 same applies to the effect of rocks protruding into and through the snow cover as was described by Juliussen and Humlum
481 (2008) which could potentially be included in CryoGrid by laterally coupled simulations (e.g. Zweigel et al., 2021) with snow
482 redistribution between tiles representing blocks of different heights. Considering air convection in future simulations (as e.g.
483 in Wicky and Hauck, 2017) should become a priority for model development as this is likely to interact with the ground ice
484 mass balance for the blocky drained scenario and could thus exacerbate the thermal anomaly.

485 Further uncertainties are related to the model forcing data. The ERA5 reanalysis data is a global product with coarse
486 horizontal resolution, so that the TopoSCALE downscaling routine (Fiddes and Gruber 2014) is applied to obtain more
487 representative meteorological forcing. Nonetheless, as mentioned in Fiddes and Gruber (2014) and Fiddes et al. (2019; 2022),
488 there are limitations to this scheme, in particular the primitive downscaling scheme for precipitation, which only interpolates
489 between ERA5 grid points and thus misses the effects of local orography. The same is true for the effects of local cloud build-
490 up around slopes and mountains, which affects the radiation budget. While these uncertainties could affect the comparison of
491 model results to field measurements (Sect. 4.1), the model forcing data can certainly capture the regional-scale climate
492 characteristics of the two study sites, e.g. the significant differences in MAAT between the two sites. The thermal anomaly of
493 the *blocks only, drained* scenario consistently occurs for both sites and thus over a significant range of climate conditions, so
494 that the effect is likely robust despite the uncertainties in the model forcing data. The same is true for the uncertainty caused
495 by the Crocus-based snow scheme (Vionnet et al., 2012; Zweigel et al., 2021). In this study, we have performed a sensitivity
496 study with respect to the amount of snow (by modifying the snowfall factor, Sect. 4.2), but simply scaling snowfall cannot
497 represent the true time evolution of the snow cover due to wind redistribution (e.g. Liston & Sturm, 1998; Martin et al., 2019),
498 possibly resulting in differences between observed and simulated temperatures. Nevertheless, it seems unlikely that the exact
499 time dynamics of snow ablation and/or deposition events strongly affects the dependence of the thermal anomaly in the *blocks*
500 *only, drained* scenario on overall winter snow depths. We therefore conclude that the significant negative thermal anomaly for
501 the *blocks only, drained* scenario is likely robust in the light of the model uncertainty.

502 **5.2 The effect of the ground ice dynamics on ground temperatures**

503 Despite the uncertainties of the model setup, our results show a clear negative thermal anomaly for the *blocks only, drained*
504 scenario. If the winter snow depth is sufficiently high, a surface cover of coarse blocks with air-filled voids (i.e. high porosity
505 and low water holding capacity) results in 2 m ground temperatures 1 to 2 °C lower than for the other stratigraphies. In the
506 Ivarsfjorden simulations, the *blocks only, drained* scenario is the only one where near-surface permafrost conditions persist

507 even today, while near-surface permafrost degrades for the *blocks with sediment* and *sediment only* scenarios. This is
508 accompanied by a strong thermal offset, with a mean ground surface temperature of more than 2 °C for the *blocks only, drained*
509 scenario, while the mean ground temperatures at 2 m were below 0 °C. Interestingly, the temperature anomaly appears largely
510 constant over time in the transient simulations, except for periods when permafrost disappears in one of the scenarios and
511 confines ground temperatures to 0 °C, which delays further ground warming. For lower snow depths, the temperature anomaly
512 becomes smaller and eventually vanishes for the (largely irrelevant) case of permanently snow-free conditions.

513 The negative temperature anomaly largely accumulates during fall and winter (Fig. 6). The active layer contains very
514 little water in the *blocks only, drained* scenario. Dry soils have a lower thermal conductivity compared to wet soils, but the
515 lack of latent heat release allows for rapid refreezing during fall which enables fast cooling of the deeper soil layers and thus
516 leads to overall lower winter temperatures. In spring, this “cold content” (i.e. sensible heat) of the ground is partly transformed
517 into the build-up of new ground ice (i.e. latent heat, Fig. 7) which only melts slowly during summer due to the insulation of
518 the overlying blocky layer. This timing of the ground ice formation is strongly different from all other scenarios, for which
519 ground ice mostly forms in fall/early winter due to refreezing of the water contained in the active layer (e.g. Hinkel et al.,
520 2001). A somewhat similar effect has been described for peat plateaus in northern Norway where simulations yielded 2 °C
521 lower temperatures for well-drained peat compared to water-saturated peat (Martin et al., 2019). This refreezing of the active
522 layer can take several months and is further delayed if a significant snow cover forms during this period, which leads to overall
523 higher winter temperatures in the permafrost due to the insulation (Zhang, 2005). It is exactly for these “high-snow situations”
524 (corresponding to higher snowfall factors in our sensitivity analyses) that the temperature anomaly of the *blocks only, drained*
525 scenario is largest. Our results for example suggest that permafrost can occur for blocky ground on slopes around Juvvasshøe,
526 even if the winter snow cover exceeds 2 m thickness.

527 We note that the thermal anomaly caused by the ground ice dynamics in blocky ground is not related to convective
528 processes (Harris and Pedersen, 1998) or the effect of blocks protruding through the snow cover (Juliussen and Humlum, 2008;
529 Gruber Hoelzle, 2008). The simulated temperature anomaly is similar to the 1.3–2.0 °C lower temperatures that Juliussen and
530 Humlum (2008) found in blockfields compared to till and bedrock in Central-eastern Norway. While a complete process model
531 for blocky ground and rock glaciers will certainly have to take air convection and the interplay between surface blocks and the
532 snow cover into account, it is encouraging that the relatively simple model approach presented in this work offers prospects to
533 improve our estimates of permafrost occurrence in mountain environments.

534 In a first-order approach, thermal anomalies can be translated into elevation differences by assuming a temperature
535 lapse rate, so that the impacts on the lower altitudinal limit of permafrost can be estimated. For a lapse rate of 0.5 °C per 100
536 m (e.g. Farbrot et al., 2011), the lower limit of permafrost in drained, blocky deposits in Norway would be 300 to 400 m lower
537 compared to “normal” permafrost represented by the other scenarios. For the Ivarsfjorden site, these numbers compare
538 favorably to the Scandinavian permafrost map (Gisnås et al., 2017) which shows a lower discontinuous permafrost limit in
539 Finnmark at around 400 m a.s.l., approximately 300 m above the rock glacier.

540 5.3 Implications for future work

541 In this study, we show that modelling the full subsurface water and ice balance in well-drained blocky deposits with air-filled
542 voids leads to significantly lower ground temperatures in permafrost environments. In modelling studies on the distribution of
543 permafrost, blocky ground usually is not accounted for, or the water balance is not simulated at this level of detail. For the
544 Northern hemisphere permafrost map (Obu et al., 2019), a coarse landcover classification was used in mountain areas which
545 did not represent blocky terrain. To produce the map, a semi-empirical equilibrium TTOP model was used in which the thermal
546 anomaly of blocky deposits likely could account for by adjusting the r_k parameter which accounts for the thermal offset of the
547 ground. A more sophisticated modelling approach as presented in this work could be used to train the more simple TTOP
548 model across a range of climate conditions. Permafrost mapping with transient models (e.g. Jafarov et al., 2012) often uses
549 fixed ground stratigraphies, in which the sum of water and ice contents does not change for a given layer. An example is the
550 transient permafrost map for southern Norway (Westermann et al., 2013) which featured a dedicated stratigraphic class for
551 blocky deposits, with a dry upper layer followed by an ice-saturated layer below, very similar to the stratigraphy used for the
552 *blocks only, drained* scenario in this study. However, both layers have a fixed thickness and the sum of water and ice contents
553 is constant, so that the temporal evolution of ground ice dynamics cannot be captured. If the seasonal thaw extends in the ice-
554 rich layer, a pool of meltwater forms which cannot drain and hence strongly delays refreezing in fall, potentially resulting in
555 the degradation of permafrost. In our simulations with full water/ice dynamics, the ice table instead varies over time, both
556 seasonally and over longer periods in response to the climatic forcing. Such changes in the ground ice table have for example
557 been observed at the Schilthorn site in the European Alps where the ground ice table was significantly lowered during a hot
558 summer and did not regrow in the following years although permafrost conditions persisted (Hilbich et al., 2008). As this
559 observation site is located on a slope, it is clear that such observed ground ice dynamics can only be reproduced if lateral
560 drainage of meltwater is taken into account. To improve transient modelling of mountain permafrost distribution, CryoGrid in
561 the configuration used in this study could be adapted for individual grid cells, especially by adjusting the strength of the lateral
562 drainage (i.e. the distance to seepage face) depending on the local slope. In flat areas and depressions, water would then pool
563 up as for the *undrained* cases, while both melt- and rainwater would drain in sloping terrain as in the *undrained* cases, with
564 corresponding changes to the ground thermal regime and permafrost distribution. Furthermore, our study suggests that the
565 presence of fine sediments in the voids between blocks can strongly alter the ground temperature compared to blocky terrain
566 with air-filled voids. For spatially distributed mapping, these two cases would have to be distinguished as separate stratigraphic
567 classes and maps of their spatial extent must be available. Especially the latter is expected to be significant challenge, as the
568 surfaces likely appear similar for remote sensors, so that detailed field mapping may be required.

569 The model approach in this study also offers significant potential to study ground ice derived runoff from blocky
570 deposits and rock glaciers. While the Norwegian study sites are both located in wet climate settings with ample water supply,
571 rock glaciers in more arid regions can be important sources of water (e.g. Croce and Milana, 2002). The global ratio of rock
572 glacier to glacier water volume equivalent is currently increasing as both systems react differently to a changing climate (Jones

573 et al., 2019). Therefore, simulations of ground ice volumes and seasonal runoff characteristics in both the present and future
574 climates can be a valuable tool for the assessment of water resources. Furthermore, rock glaciers are sensitive to climate change
575 (Haeberli et al., 2010) and recent studies have linked rock glacier acceleration to increasing air and ground temperatures (e.g.
576 Käab et al., 2007; Hartl et al., 2016; Eriksen et al., 2018; Thibert and Bodin, 2022). Our model approach is likely able to
577 simulate the seasonal ground ice mass balance at different points and elevations of a rock glacier which could be ingested in a
578 flow model for rock glaciers (e.g. Monnier and Kinnard, 2016). Finally, permafrost degradation and ground ice loss can also
579 play an important role for slope stability in mountain permafrost environments (e.g. Gruber and Haeberli, 2007; Saemundsson
580 et al., 2018; Nelson et al., 2001). Simulations of ground ice table changes, as well as the occurrence of strong melt events with
581 corresponding production of meltwater, could eventually improve assessments of the stability and hazard potential of
582 permafrost-underlain slopes (e.g. Mamot et al., 2021).

583 **6. Conclusions**

584 In this study, we used the CryoGrid permafrost model to simulate the effect of blocky terrain on the ground thermal regime
585 and ground ice dynamics at two Norwegian mountain permafrost sites (Juvvasshøe and Ivarsfjorden rock glacier). In particular,
586 we investigated the effect of subsurface drainage, as typical on slopes, for three idealized stratigraphies, named *blocks only*,
587 *blocks with sediment* and *sediment only*. From this study, the following conclusions can be drawn:

588

- 589 • Markedly lower ground temperatures are found in well drained, coarse blocky deposits with air-filled voids (*blocks*
590 *only, drained* scenario) compared to other scenarios which are either undrained or feature fine sediments. This
591 negative thermal anomaly can exceed 2 °C and is mainly linked to differences in the freeze-thaw dynamics caused
592 by the removal of meltwater and the build-up of new ground ice in spring. The largest anomalies occur in simulations
593 with a thick winter snow cover as ground temperatures in well drained blocky deposits are less sensitive to insulation
594 by snow than other soils. We emphasize that the model does not account for well-known factors, such as air
595 convection and the effect of blocks protruding through the winter snow cover.
- 596 • For the *blocks only, drained* scenario, thermally stable permafrost can exist at the Ivarsfjorden rock glacier site
597 (located near sea level), even for a mean annual ground surface temperature of 2.0–2.5 °C. At Juvvasshøe in the
598 southern Norwegian mountains, permafrost is simulated even for a very thick winter snow cover in the *blocks only*,
599 *drained* scenario, while all other scenarios in this case feature permafrost-free conditions.
- 600 • Transient simulations since 1951 at the Ivarsfjorden rock glacier show a completely or partially degraded ground ice
601 table for all scenarios, except the *blocks only, drained* scenario. This result is explained by the overall lower ground
602 temperatures in this scenario, while the simulated warming rates are generally similar for all scenarios, except for
603 periods when strong ground ice melt occurs.

604

605 This study suggests that including subsurface water and ice dynamics can drive simulations of mountain permafrost dynamics
606 towards reality, which can for example improve estimates of the lower altitudinal limit of permafrost in blocky terrain. In
607 addition to permafrost distribution mapping, the presented model approach could be used to simulate the seasonal and multi-
608 annual evolution of the ground ice table, in addition to ground-ice derived runoff. It therefore represents a further step to a
609 better understanding and model representation of the permafrost processes in mountain environments.

610

611 *Code and data availability.* The CryoGrid source code and model setup files are available
612 <https://doi.org/10.5281/zenodo.6563651> (Renette, 2022). Field measurements at Juvvasshøe are from Etzelmüller et al. (2020).
613 Field measurements at Ivarsfjorden are from Lilleøren et al. (2022).

614

615 *Author contribution.* CR performed the model simulations, retrieved forcing data, wrote the draft manuscript and created all
616 figures. SW helped design the study, developed the model and provided ideas throughout the entire study. KA developed code
617 for retrieving and downscaling forcing data, assisted with this process and wrote text regarding the forcing data. KL and KI
618 provided field measurements, site descriptions and photos. RBZ and JA developed parts of the model. All authors contributed
619 with text and suggestions.

620

621 *Competing interests.* The authors declare that they have no conflict of interest.

622

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627 **References**

- 628 Aalstad, K., Westermann, S., Schuler, T. V., Boike, J., and Bertino, L.: Ensemble-based assimilation of fractional snow-
629 covered area satellite retrievals to estimate the snow distribution at Arctic sites, *The Cryosphere*, 12, 247–270,
630 <https://doi.org/10.5194/tc-12-247-2018>, 2018.
- 631 Arenson, L. U., Phillips, M., and Springman, S. M.: Geotechnical considerations and technical solutions for infrastructure in
632 mountain permafrost, in: *New permafrost and glacier research*, pp. 3–50, Nova Science Publishers, 2009.
- 633 Azócar, G. and Brenning, A.: Hydrological and geomorphological significance of rock glaciers in the dry Andes, Chile (27–
634 33 S), *Permafrost and Periglacial Processes*, 21, 42–53, <https://doi.org/10.1002/ppp.669>, 2010.
- 635 Croce, F. A. and Milana, J. P.: Internal structure and behaviour of a rock glacier in the arid Andes of Argentina, *Permafrost
636 and Periglacial Processes*, 13, 289–299, <https://doi.org/10.1002/ppp.431>, 2002.
- 637 Dahl, R.: Block fields, weathering pits and tor-like forms in the Narvik Mountains, Nordland, Norway, *Geografiska Annaler:
638 Series A, Physical Geography*, 48, 55–85, <https://doi.org/10.1080/04353676.1966.11879730>, 1966.
- 639 Eriksen, H., Rouyet, L., Lauknes, T., Berthling, I., Isaksen, K., Hindberg, H., Larsen, Y., and Corner, G.: Recent acceleration
640 of a rock glacier complex, Adjet, Norway, documented by 62 years of remote sensing observations, *Geophysical Research
641 Letters*, 45, 8314–8323, <https://doi.org/10.1029/2018GL077605>, 2018.
- 642 Etzelmüller, B., Berthling, I., and Sollid, J. L.: Aspects and concepts on the geomorphological significance of Holocene
643 permafrost in southern Norway, *Geomorphology*, 52, 87–104, [https://doi.org/10.1016/S0169-555X\(02\)00250-7](https://doi.org/10.1016/S0169-555X(02)00250-7), 2003.
- 644 Etzelmüller, B., Schuler, T. V., Isaksen, K., Christiansen, H. H., Farbrot, H., and Benestad, R.: Modeling the temperature
645 evolution of Svalbard permafrost during the 20th and 21st century, *The Cryosphere*, 5, 67–79, [https://doi.org/10.5194/tc-5-67-
2011](https://doi.org/10.5194/tc-5-67-
646 2011), 2011.
- 647 Etzelmüller, B., Guglielmin, M., Hauck, C., Hilbich, C., Hoelzle, M., Isaksen, K., Noetzli, J., Oliva, M., and Ramos, M.:
648 Twenty years of European mountain permafrost dynamics—the PACE legacy, *Environmental Research Letters*, 15, 104070,
649 <https://doi.org/10.1088/1748-9326/abae9d>, 2020.
- 650 Farbrot, H., Hipp, T. F., Etzelmüller, B., Isaksen, K., Ødegård, R. S., Schuler, T. V., and Humlum, O.: Air and ground
651 temperature variations observed along elevation and continentality gradients in Southern Norway, *Permafrost and Periglacial
652 Processes*, 22, 343–360, <https://doi.org/10.1002/ppp.733>, 2011.
- 653 Fiddes, J. and Gruber, S.: TopoSCALE v. 1.0: downscaling gridded climate data in complex terrain, *Geoscientific Model
654 Development*, 7, 387–405, <https://doi.org/10.5194/gmd-7-387-2014>, 2014.
- 655 Fiddes, J., Endrizzi, S., and Gruber, S.: Large-area land surface simulations in heterogeneous terrain driven by global data sets:
656 application to mountain permafrost, *The Cryosphere*, 9, 411–426, <https://doi.org/10.5194/tc-9-411-2015>, 2015.
- 657 Fiddes, J., Aalstad, K., and Westermann, S.: Hyper-resolution ensemble-based snow reanalysis in mountain regions using
658 clustering, *Hydrology and Earth System Sciences*, 23, 4717–4736, <https://doi.org/10.5194/hess-23-4717-2019>, 2019.

659 Fiddes, J., Aalstad, K., and Lehning, M.: TopoCLIM: rapid topography-based downscaling of regional climate model output
660 in complex terrain v1. 1, *Geoscientific Model Development*, 15, 1753–1768, <https://doi.org/10.5194/gmd-15-1753-2022>,
661 2022.

662 Gislén, K., Eitzinger, B., Lussana, C., Hjørt, J., Sannel, A. B. K., Isaksen, K., Westermann, S., Kuhry, P., Christiansen, H.
663 H., Frampton, A., et al.: Permafrost map for Norway, Sweden and Finland, *Permafrost and periglacial processes*, 28, 359–378,
664 <https://doi.org/10.1002/ppp.1922>, 2017.

665 Gislén, K., S. Westermann, T. V. Schuler, T. Litherland, K. Isaksen, J. Boike, and B. Eitzinger.: A statistical approach to
666 represent small-scale variability of permafrost temperatures due to snow cover, *The Cryosphere*, 8.6, pp. 2063–2074.
667 <https://doi.org/10.5194/tc-8-2063-2014>, 2014.

668 Gislén, K., Westermann, S., Schuler, T. V., Melvold, K., and Eitzinger, B.: Small-scale variation of snow in a regional
669 permafrost model, *The Cryosphere*, 10, 1201–1215, <https://doi.org/10.5194/tc-10-1201-2016>, 2016.

670 Göckede, M., Kittler, F., Kwon, M. J., Burjack, I., Heimann, M., Kolle, O., et al.: Shifted energy fluxes, increased Bowen
671 ratios, and reduced thaw depths linked with drainage-induced changes in permafrost ecosystem structure, *The Cryosphere*,
672 11(6), 2975–2996. <https://doi.org/10.5194/tc-11-2975-2017>, 2017.

673 Goodrich, L. E.: The influence of snow cover on the ground thermal regime, *Can. Geotech. J.*, 19, 421–432,
674 <https://doi.org/10.1139/t82-047>, 1982.

675 Gruber, S. and Haeberli, W.: Permafrost in steep bedrock slopes and its temperature-related destabilization following climate
676 change, *Journal of Geophysical Research: Earth Surface*, 112, <https://doi.org/10.1029/2006JF000547>, 2007.

677 Gruber, S. and Hoelze, M.: The cooling effect of coarse blocks revisited: a modeling study of a purely conductive mechanism,
678 *Zurich Open Repository and Archive*, 2008.

679 Haeberli, W., Hallet, B., Arenson, L., Elconin, R., Humlum, O., Käab, A., Kaufmann, V., Ladanyi, B., Matsuoka, N.,
680 Springman, S., et al.: Permafrost creep and rock glacier dynamics, *Permafrost and periglacial processes*, 17, 189–214,
681 <https://doi.org/10.1002/ppp.561>, 2006.

682 Hanson, S. and Hoelzle, M.: The thermal regime of the active layer at the Murtèl rock glacier based on data from 2002,
683 *Permafrost and Periglacial Processes*, 15, 273–282, <https://doi.org/10.1002/ppp.499>, 2004.

684 Harris, C., Haeberli, W., Vonder Mühl, D., and King, L.: Permafrost monitoring in the high mountains of Europe: the PACE
685 project in its global context, *Permafrost and periglacial processes*, 12, 3–11, <https://doi.org/10.1002/ppp.377>, 2001.

686 Harris, C., Arenson, L. U., Christiansen, H. H., Eitzinger, B., Frauenfelder, R., Gruber, S., Haeberli, W., Hauck, C., Hoelzle,
687 M., Humlum, O., et al.: Permafrost and climate in Europe: Monitoring and modelling thermal, geomorphological and
688 geotechnical responses, *EarthScience Reviews*, 92, 117–171, <https://doi.org/10.1016/j.earscirev.2008.12.002>, 2009.

689 Harris, S. A. and Pedersen, D. E.: Thermal regimes beneath coarse blocky materials, *Permafrost and periglacial processes*, 9,
690 107–120, [https://doi.org/10.1002/\(SICI\)1099-1530\(199804/06\)9:2<107::AID-PPP277>3.0.CO;2-G](https://doi.org/10.1002/(SICI)1099-1530(199804/06)9:2<107::AID-PPP277>3.0.CO;2-G), 1998.

691 Hartl, L., Fischer, A., Stocker-waldhuber, M., and Abermann, J.: Recent speed-up of an alpine rock glacier: an updated
692 chronology of the kinematics of outer hochebenkar rock glacier based on geodetic measurements, *Geografiska Annaler: Series*
693 *A, Physical Geography*, 98, 129–141, <https://doi.org/10.1111/geoa.12127>, 2016.

694 Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., Nicolas, J., Peubey, C., Radu, R., Schepers,
695 D., et al.: The ERA5 global reanalysis, *Quarterly Journal of the Royal Meteorological Society*, 146, 1999–2049,
696 <https://doi.org/10.1002/qj.3803>, 2020.

697 Hilbich, C., Hauck, C., Hoelzle, M., Scherler, M., Schudel, L., Völksch, I., Vonder Mühll, D., and Mäusbacher, R.: Monitoring
698 mountain permafrost evolution using electrical resistivity tomography: A 7-year study of seasonal, annual, and long-term
699 variations at Schilthorn, Swiss Alps, *Journal of Geophysical Research: Earth Surface*, 113,
700 <https://doi.org/10.1029/2007JF000799>, 2008.

701 Hinkel, K. M. and Outcalt, S. I.: Identification of heat transfer processes during soil cooling, freezing, and thaw in central
702 Alaska, *Permafrost Periglac.*, 5, 217–235, <https://doi.org/10.1002/ppp.3430050403>, 1994.

703 Hinkel, K. M., Paetzold, F., Nelson, F. E., and Bockheim, J. G.: Patterns of soil temperature and moisture in the active layer
704 and upper permafrost at Barrow, Alaska: 1993–1999, *Global Planet. Change*, 29, 293–309, 2001.

705 Hipp, T., Etzelmüller, B., Farbrot, H., Schuler, T., and Westermann, S.: Modelling borehole temperatures in Southern Norway–
706 insights into permafrost dynamics during the 20th and 21st century, *The Cryosphere*, 6, 553–571, [https://doi.org/10.5194/tc-](https://doi.org/10.5194/tc-6-553-2012)
707 [6-553-2012](https://doi.org/10.5194/tc-6-553-2012), 2012.

708 Humlum, O.: Active layer thermal regime at three rock glaciers in Greenland, *Permafrost and Periglacial Processes*, 8, 383–
709 408, [https://doi.org/10.1002/\(SICI\)1099-1530\(199710/12\)8:4<383::AID-PPP265>3.0.CO;2-V](https://doi.org/10.1002/(SICI)1099-1530(199710/12)8:4<383::AID-PPP265>3.0.CO;2-V), 1997.

710 Isaksen, K., Holmlund, P., Sollid, J. L., and Harris, C.: Three deep alpine-permafrost boreholes in Svalbard and Scandinavia,
711 *Permafrost and Periglacial Processes*, 12, 13–25, <https://doi.org/10.1002/ppp.380>, 2001.

712 Isaksen, K., Heggem, E., Bakkehøi, S., Ødegård, R., Eiken, T., Etzelmüller, B., and Sollid, J.: Mountain permafrost and energy
713 balance on Juvvasshøe, southern Norway, in: 8th International Conference on Permafrost, Zurich, Switzerland, ISI:
714 000185049300083, pp. 467–472, 2003.

715 Isaksen, K., Sollid, J. L., Holmlund, P., and Harris, C.: Recent warming of mountain permafrost in Svalbard and Scandinavia,
716 *Journal of Geophysical Research: Earth Surface*, 112, <https://doi.org/10.1029/2006JF000522>, 2007.

717 Jafarov, E. E., Marchenko, S. S., and Romanovsky, V.: Numerical modeling of permafrost dynamics in Alaska using a high
718 spatial resolution dataset, *The Cryosphere*, 6, 613–624, <https://doi.org/10.5194/tc-6-613-2012>, 2012.

719 Jones, D. B., Harrison, S., Anderson, K., and Whalley, W. B.: Rock glaciers and mountain hydrology: A review, *Earth-Science*
720 *Reviews*, 193, 66–90, <https://doi.org/10.1016/j.earscirev.2019.04.001>, 2019.

721 Juliussen, H. and Humlum, O.: Thermal regime of openwork block fields on the mountains Elgåhogna and Sjølen, central-
722 eastern Norway, *Permafrost and Periglacial Processes*, 19, 1–18, <https://doi.org/10.1002/ppp.607>, 2008.

723 Kääb, A., Frauenfelder, R., and Roer, I.: On the response of rockglacier creep to surface temperature increase, *Global and*
724 *Planetary Change*, 56, 172–187, <https://doi.org/10.1016/j.gloplacha.2006.07.005>, 2007.

725 Langer, M., Westermann, S., Boike, J., Kirillin, G., Grosse, G., Peng, S., and Krinner, G.: Rapid degradation of permafrost
726 underneath waterbodies in tundra landscapes—toward a representation of thermokarst in land surface models, *Journal of*
727 *Geophysical Research: Earth Surface*, 121, 2446–2470, <https://doi.org/10.1002/2016JF003956>, 2016.

728 Liestøl, O.: Lokalt omrøde med permafrost i Gudbrandsdalen, *Norsk Polarinstitutt Aarbok*, 1965, 129-133, 1966.

729 Liljedahl, A. K., Hinzman, L. D., Harazono, Y., Zona, D., Tweedie, C. E., Hollister, R. D., et al.: Nonlinear controls on
730 evapotranspiration in Arctic coastal wetlands. *Biogeosciences*, 8(11), 3375–3389. <https://doi.org/10.5194/bg-8-3375-2011>,
731 2011.

732 Lilleøren, K. S. and Eitzelmüller, B.: A regional inventory of rock glaciers and ice-cored moraines in Norway, *Geografiska*
733 *Annaler: Series A, Physical Geography*, 93, 175–191, <https://doi.org/10.1111/j.1468-0459.2011.00430.x>, 2011.

734 Lilleøren, K. S., Eitzelmüller, B., Rouyet, L., Eiken, T., and Hilbich, C.: Transitional rock glaciers at sea-level in Northern
735 Norway, *Earth Surface Dynamics Discussions*, pp. 1–29, <https://doi.org/10.5194/esurf-2022-6>, 2022.

736 Liston, G. E. and Sturm, M.: A snow-transport model for complex terrain, *Journal of Glaciology*, 44, 498–516,
737 <https://doi.org/10.3189/S0022143000002021>, 1998.

738 Luetsch, M., Stoeckli, V., Lehning, M., Haeblerli, W., and Ammann, W.: Temperatures in two boreholes at Flüela Pass,
739 Eastern Swiss Alps: the effect of snow redistribution on permafrost distribution patterns in high mountain areas, *Permafrost*
740 *Periglac.*, 15, 283–297, doi:10.1002/ppp.500, 2004.

741 Mamot, P., Weber, S., Eppinger, S., and Krautblatter, M.: A temperature-dependent mechanical model to assess the stability
742 of degrading permafrost rock slopes, *Earth Surface Dynamics*, 9, 1125–1151, <https://doi.org/10.5194/esurf-9-1125-2021>,
743 2021.

744 Martin, L. C. P., Nitzbon, J., Aas, K. S., Eitzelmüller, B., Kristiansen, H., and Westermann, S.: Stability conditions of peat
745 plateaus and palsas in northern Norway, *Journal of Geophysical Research: Earth Surface*, 124, 705–719,
746 <https://doi.org/10.1029/2018JF004945>, 2019.

747 Monnier, S. and Kinnard, C.: Interrogating the time and processes of development of the Las Liebres rock glacier, central
748 Chilean Andes, using a numerical flow model, *Earth Surface Processes and Landforms*, 41, 1884–1893,
749 <https://doi.org/10.1002/esp.3956>, 2016.

750 Nelson, F. E., Anisimov, O. A., and Shiklomanov, N. I.: Subsidence risk from thawing permafrost, *Nature*, 410, 889–890,
751 <https://doi.org/10.1038/35073746>, 2001.

752 Nesje, A., Matthews, J. A., Linge, H., Bredal, M., Wilson, P., and Winkler, S.: New evidence for active talus-foot rock glaciers
753 at Øyberget, southern Norway, and their development during the Holocene, *The Holocene*, 31, 1786–1796,
754 <https://doi.org/10.1177/09596836211033226>, 2021.

755 Obu, J., Westermann, S., Bartsch, A., Berdnikov, N., Christiansen, H. H., Dashtseren, A., Delaloye, R., Elberling, B.,
756 Eitzelmüller, B., Kholodov, A., et al.: Northern Hemisphere permafrost map based on TTOP modelling for 2000–2016 at 1
757 km² scale, *Earth-Science Reviews*, 193, 299–316, <https://doi.org/10.1016/j.earscirev.2019.04.023>, 2019.

758 Painter, S. L. and Karra, S.: Constitutive model for unfrozen water content in subfreezing unsaturated soils, *Vadose Zone*
759 *Journal*, 13, <https://doi.org/10.2136/vzj2013.04.0071>, 2014.

760 Porter, C., Morin, P., Howat, I., Noh, M., Bates, B., Peterman, K., Keeseey, S., Schlenk, M., Gardiner, J., Tomko, K., et al.:
761 ArcticDEM, Harvard Dataverse [data set], V1, <https://doi.org/10.7910/DVN/OHHUKH>, 2018.

762 Renette, C.: Parameter files and code for simulations in "Simulating the effect of subsurface drainage on the thermal regime
763 and ground ice in blocky terrain, Norway" [Data set], Zenodo, <https://doi.org/10.5281/zenodo.6563651>, 2022.

764 Romundset, A., Bondevik, S., and Bennike, O.: Postglacial uplift and relative sea level changes in Finnmark, northern Norway,
765 *Quat. Sci. Rev.*, 30, 2398-2421, <https://doi.org/10.1016/j.quascirev.2011.06.007>, 2011.

766 Sæmundsson, Þ., Morino, C., Helgason, J. K., Conway, S. J., and Pétursson, H. G.: The triggering factors of the Móafellshyrna
767 debris slide in northern Iceland: Intense precipitation, earthquake activity and thawing of mountain permafrost, *Science of the*
768 *total environment*, 621, 1163–1175, <https://doi.org/10.1016/j.scitotenv.2017.10.111>, 2018.

769 Saloranta, T.: Simulating snow maps for Norway: description and statistical evaluation of the seNorge snow model, *The*
770 *Cryosphere*, 6, 1323–1337, <https://doi.org/10.5194/tc-6-1323-2012>, 2012.

771 Scudeler, C., Paniconi, C., Pasetto, D., and Putti, M.: Examination of the seepage face boundary condition in subsurface and
772 coupled surface/subsurface hydrological models, *Water Resources Research*, 53(3), 1799-1819, 2017.

773 Schmidt, J. U., Etzelmüller, B., Schuler, T. V., Magnin, F., Boike, J., Langer, M., and Westermann, S.: Surface temperatures
774 and their influence on the permafrost thermal regime in high-Arctic rock walls on Svalbard, *The Cryosphere*, 15, 2491–2509,
775 <https://doi.org/10.5194/tc-15-2491-2021>, 2021.

776 Smith, M. and Riseborough, D.: Climate and the limits of permafrost: a zonal analysis, *Permafrost and Periglacial Processes*,
777 13, 1–15, <https://doi.org/10.1002/ppp.410>, 2002.

778 Thibert, E, Bodin, X.: Changes in surface velocities over four decades on the Laurichard rock glacier (French Alps), *Permafrost*
779 *and Periglac Process*, 33(3): 323- 335, <https://doi.org/10.1002/ppp.2159>, 2022.

780 Van Everdingen, R. O.: Multi-language glossary of permafrost and related ground-ice terms, *International Permafrost*
781 *Association*, 1998.

782 Vionnet, V., Brun, E., Morin, S., Boone, A., Faroux, S., Le Moigne, P., Martin, E., and Willemet, J.-M.: The detailed snowpack
783 scheme Crocus and its implementation in SURFEX v7. 2, *Geoscientific Model Development*, 5, 773–791,
784 <https://doi.org/10.5194/gmd-5-773-2012>, 2012.

785 Westermann, S., Schuler, T., Gislås, K., and Etzelmüller, B.: Transient thermal modeling of permafrost conditions in Southern
786 Norway, *The Cryosphere*, 7, 719–739, <https://doi.org/10.5194/tc-7-719-2013>, 2013.

787 Westermann, S., Langer, M., Boike, J., Heikenfeld, M., Peter, M., Etzelmüller, B., and Krinner, G.: Simulating the thermal
788 regime and thaw processes of ice-rich permafrost ground with the land-surface model CryoGrid 3, *Geoscientific Model*
789 *Development*, 9, 523–546, <https://doi.org/10.5194/gmd-9-523-2016>, 2016.

790 Westermann, S., Ingeman-Nielsen, T., Scheer, J., Aalstad, K., Aga, J., Chaudhary, N., Etzelmüller, B., Filhol, S., Kääb, A.,
791 Renette, C., Schmidt, L. S., Schuler, T. V., Zweigel, R. B., Martin, L., Morard, S., Ben-Asher, M., Angelopoulos, M., Boike,

792 J., Groenke, B., Miesner, F., Nitzbon, J., Overduin, P., Stuenzi, S. M., and Langer, M.: The CryoGrid community model
793 (version 1.0) – a multi-physics toolbox for climate-driven simulations in the terrestrial cryosphere, *Geosci. Model Dev.*
794 *Discuss.* [preprint], <https://doi.org/10.5194/gmd-2022-127>, in review, 2022.

795 Wicky, J. and Hauck, C.: Numerical modelling of convective heat transport by air flow in permafrost talus slopes, *The*
796 *Cryosphere*, 11, 1311–1325, <https://doi.org/10.5194/tc-11-1311-2017>, 2017.

797 Zhang, T.: Influence of the seasonal snow cover on the ground thermal regime: An overview, *Reviews of Geophysics* 43.4,
798 <https://doi.org/10.1029/2004RG000157>, 2005.

799 Zweigel, R., Westermann, S., Nitzbon, J., Langer, M., Boike, J., Eitzelmüller, B., and Vikhamar Schuler, T.: Simulating snow
800 redistribution and its effect on ground surface temperature at a high-Arctic site on Svalbard, *Journal of Geophysical Research:*
801 *Earth Surface*, 126, e2020JF005673, <https://doi.org/10.1029/2020JF005673>, 2021.

802