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► **To cite this version:**

Léo C P Martin, Jan Nitzbon, Johanna Scheer, Kjetil S Aas, Trond Eiken, et al.. Thermal erosion patterns of permafrost peat plateaus in northern Norway. 2023. hal-03967417

**HAL Id: hal-03967417**

**<https://hal.science/hal-03967417>**

Preprint submitted on 1 Feb 2023

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## Thermal erosion patterns of permafrost peat plateaus in northern Norway

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## 1 Abstract

2 Subarctic peatlands underlain by permafrost contain significant amounts of organic carbon and our  
3 ability to quantify the evolution of such permafrost landscapes in numerical models is critical to provide  
4 robust predictions of the environmental and climatic changes to come. Yet, the accuracy of large-scale  
5 predictions is so far hampered by small-scale physical processes that create a high spatial variability of  
6 surface ground thermal regime and thus of permafrost degradation patterns. In this regard, a better  
7 understanding of the small-scale interplay between microtopography and lateral fluxes of heat, water and  
8 snow can be achieved by field monitoring and process-based numerical modeling.

9 Here, we quantify the topographic changes of the Šuoššjávri peat plateau (Northern Norway) over a  
10 three-years period using repeated drone-based high-resolution photogrammetry. Our results show that edge  
11 degradation is the main process through which thermal erosion occurs and represents about 80 % of measured  
12 subsidence, while most of the inner plateau surface exhibits no detectable subsidence. Based on detailed  
13 investigation of eight zones of the plateau edge, we show that this edge degradation corresponds to a  
14 volumetric loss of  $0.13 \pm 0.07 \text{ m}^3 \text{ yr}^{-1} \text{ m}^{-1}$  (cubic meter per year and per meter of plateau circumference).

15 Using the CryoGrid land surface model, we show that these degradation patterns can be reproduced  
16 in a modeling framework that implements lateral redistribution of snow, subsurface water and heat, as well  
17 as ground subsidence due to melting of excess ice. We reproduce prolonged climate-driven edge degradation  
18 that is consistent with field observations and present a sensitivity test of the plateau degradation on snow  
19 depth over the plateau. Small snow depth variations (from 0 to 30 cm) result in highly different degradation  
20 behavior, from stability to fast degradation.

21 These results represent a new step in the modeling of climate-driven landscape development and  
22 permafrost degradation in highly heterogeneous landscapes such as peat plateaus. Our approach provides a  
23 physically based quantification of permafrost thaw with a new level of realism, notably regarding feedback  
24 mechanisms between the dynamical topography and the lateral fluxes through which a small modification of  
25 the snow depth results in a major modification of the permafrost degradation intensity. In this regard, these  
26 results also highlight the major control of snow pack characteristics on the ground thermal regime and the  
27 benefit of improving snow representation in numerical models for permafrost degradation projections.



28

## 29 **1. Introduction**

30 Observations show that permafrost is warming at a global scale (Biskaborn et al., 2019). Its  
31 thawing has major consequences on Arctic and Boreal ecosystems and landscapes (Beck et al., 2015;  
32 Farquharson et al., 2019; Liljedahl et al., 2016) and potentially represents an important climate feedback  
33 through the degradation of thawed organic matter (Koven et al., 2015; Schuur et al., 2009, 2015). Carbon  
34 emissions from permafrost regions towards the atmosphere are already observed (Natali et al., 2019);  
35 field measurements show that these emissions are influenced by the timing of the active layer deepening  
36 (Morgalev et al., 2017) and by the state of degradation of the permafrost terrains (Langer et al., 2015;  
37 Nwaishi et al., 2020; Serikova et al., 2018). Particularly, abrupt thaw of ice-rich permafrost is expected  
38 to become a dominant process regarding carbon emissions, offsetting the potential ecosystemic negative  
39 feedback that is expected for gradual thaw (McGuire et al., 2018; Turetsky et al., 2020).

40 As such, our ability to quantify and represent the physical evolution of permafrost landscapes is  
41 critical to provide robust predictions of the environmental and climatic changes to come (Aas et al.,  
42 2019; Andresen et al., 2020; Teufel and Sushama, 2019). While permafrost affects about 14 millions  
43 square kilometers throughout the Northern Hemisphere (Obu et al., 2019), the ground thermal response  
44 to climatic signal and morphological changes of permafrost are governed by processes occurring within  
45 a spatial scale of few meters (Gisnås et al., 2014; Jones et al., 2016; Martin et al., 2019; Way et al.,  
46 2018). Indeed, at the small-scale, Arctic and Boreal permafrost landscapes in low lands (such as peat  
47 plateaus and polygonal tundra) are characterized by low amplitude (0-3 m vertically) and high frequency  
48 (10-100 m horizontally) spatial variations of their topography called microtopography (French, 2018).  
49 On the one hand, microtopography drives the redistribution of snow, liquid water and heat and these  
50 lateral fluxes can dramatically modify the ground thermal regime and water content (Martin et al., 2019).  
51 On the other hand, permafrost microtopography result from the presence of excess ice in the ground (ice  
52 content superior to natural porosity) and permafrost thawing drives surface subsidence (“thermokarst”,  
53 Göckede et al., 2017, 2019; Nitzbon et al., 2019, 2020). However, the representation of this feedback  
54 between small-scale fluxes and dynamical topography in numerical models is still in its infancy. Robust



55 predictions of the physical state of permafrost landscapes thus require further field observations and  
56 numerical developments to refine our understanding of these phenomena.

57 Peat plateaus are permafrost landforms covering extensive regions throughout the Boreal and  
58 Arctic realms which store nearly 200 Pg of carbon (Lindgren et al., 2018). They are mainly located in  
59 the sporadic permafrost zone (Seppälä, 1972; Sollid & Sørbel, 1998) and are typically associated with a  
60 climatic envelope characterized by a mean annual ground temperature around 0°C and limited  
61 precipitation (< 800 mm yr<sup>-1</sup>) (Aalto et al., 2017; Parviainen and Luoto, 2007). Permafrost underneath  
62 peat plateaus is thus relatively warm and its distribution is highly sensitive to climate changes (Aalto et  
63 al., 2014, 2017; Fronzek et al., 2010; Luoto et al., 2004). The limit of continuous and discontinuous  
64 permafrost in the North Hemisphere is already moving northward (Thibault and Payette, 2009) and peat  
65 plateaus ‘degradation is observed in the North American Arctic (Jones et al., 2016; Mamet et al., 2017;  
66 Payette et al., 2004), Fennoscandia (Borge et al., 2017) and Western Siberia (Jones et al., 2016;  
67 Sherstyukov and Sherstyukov, 2015). In Northern Norway, the analysis of aerial imagery showed a  
68 decrease between 33 and 71 % in the lateral extent of the peat plateaus since the 1950’s (Borge et al.,  
69 2017), with the largest lateral changes since the 2000’s. These results suggest that “lateral erosion” (as  
70 coined by Borge et al., 2017) plays a crucial role in permafrost degradation.

71 The ongoing degradation of Fennoscandian peat plateaus is a potential analogue for the future  
72 of much larger peat plateau areas found in Russia, Canada and Alaska. It provides the opportunity for  
73 field measurements and process-based approaches to further understand the local drivers of permafrost  
74 peatland thermal dynamics. Several studies contributed to understand how microtopography drives the  
75 lateral fluxes of heat, water and snow and to quantify their influence on the ground thermal regime with  
76 both field measurements and numerical modeling experiments (Martin et al., 2019; Sannel, 2020; Sannel  
77 et al., 2016; Sjöberg et al., 2016). The snow and water transport from the plateau towards the surrounding  
78 mire is critical to maintain important surface temperature difference between them (1.5 to 2°C of mean  
79 annual temperature at 1m depth, Martin et al. 2019). It affects the subsurface thermal regime and enables  
80 the presence of permafrost in regions where the mean annual air temperature is above 0°C (Jones et al.,  
81 2016; Martin et al., 2019; Sannel and Kuhry, 2011). However, a comprehensive model that could

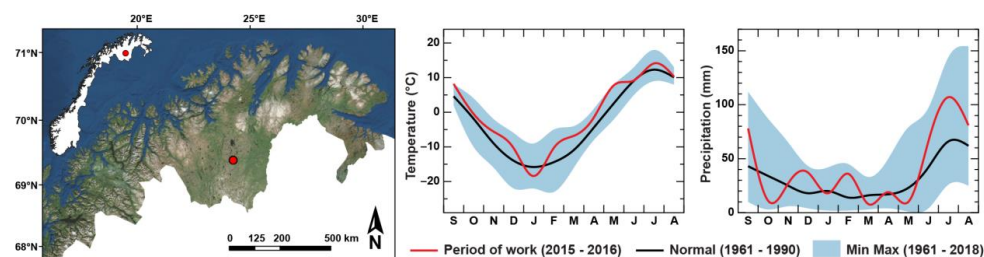


82 simulate the landscape evolution (including observed thermal erosion patterns) in a quantitative process-  
83 based fashion is lacking.

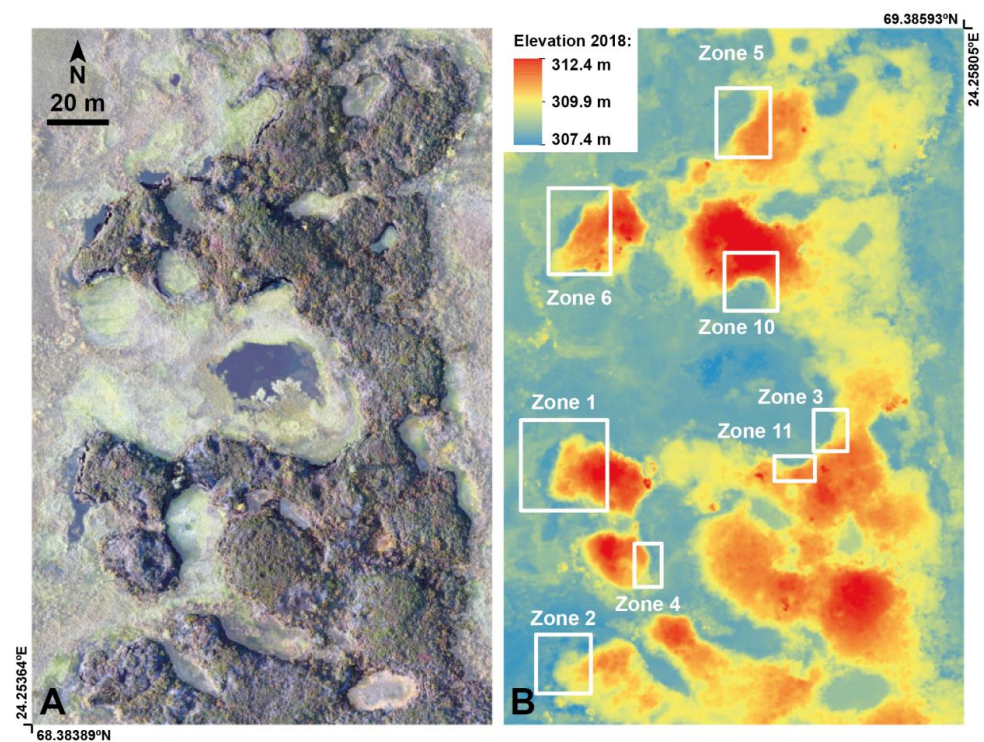
84 Here, we quantify volume changes of a peat plateau in Northern Norway using repeated digital  
85 elevation models compiled from drone aerial imagery. The surface topography of the plateau was  
86 reconstructed and compared for the years 2015 and 2018 using drone-based photogrammetry.  
87 Subsequently, we adapt the laterally coupled tiled version of the CryoGrid model (Langer et al., 2016;  
88 Nitzbon et al., 2019; Westermann et al., 2016) to reproduce observed patterns of microtopography  
89 change, including an analysis of model sensitivity towards snow depth. The work presented here  
90 prolongates the field observations and numerical simulations of Martin et al. (2019). Besides, it also  
91 builds on recent studies dedicated to numerical modelling of the ground thermal regime, lateral heat,  
92 snow and water fluxes and dynamical topography in arctic landscapes (Aas et al., 2019; Nitzbon et al.,  
93 2019, 2020).

## 94 **2. Study area: the Šuoššjávri peat plateaus**

95 The Šuoššjávri site (69.38° N, 24.25° E, around 310 m asl, Fig. 2) is situated in Finnmarksvidda,  
96 Northern Norway, and extends over approximately 23 ha. A detailed presentation of the Šuoššjávri peat  
97 plateaus can be found in Martin et al. (2019), and a map detailing the geomorphological context around  
98 the study site is presented in the Supplementary Material (Fig. A2). The climate of Finnmarksvidda (Fig.  
99 1) is continental with mean annual air temperature ranging from -4°C to -2°C (Aune, 1993) and mean  
100 annual precipitation of less than 400 mm. The Šuoššjávri site consists of a laterally carved peat plateaus  
101 and smaller peat mounds surrounded by wet mires and ponds. These peat bodies extend over meters to  
102 several tens of meters with tortuous horizontal geometries and rise 0.1 m to 3 m above the surrounding  
103 wet mire (Fig. 2). Several of the peat plateau edges show signs of advanced degradation and lateral  
104 erosion. The stratigraphy of the plateau and its spatial variability are presented in Sect. 3.3.3. Literature  
105 results for this site are discussed in Sect. 4.



106  
107 *Figure 1. Location and climate of the study site (Šuoššjávri, Finnmark). The red curve on the temperature and*  
108 *precipitation graphs correspond to the study period (2015-2016). The black curves indicate normal values and*  
109 *the blue shading the minimum and maximum values over the 1961–2018 period. Data from the Cuovdatmohkki*  
110 *station (no. 97350) located at 286 m asl, 7 km east from Šuoššjávri (310 m asl). Data from the Norwegian*  
111 *Meteorological Institute (sharki.oslo.dnmi.no). Landsat satellite image (ESRI). All three panels of the figure are*  
112 *from Martin et al. (2019).*



113  
114 *Figure 2. Area of interest of the Šuoššjávri peat plateau. A. Orthophoto of the peat plateau. We applied a*  
115 *transparent white shading on the mire to better distinguish it from the plateau. B. Digital Elevation Model of the*  
116 *plateau in 2018. The edge transect areas on which this study focuses are indicated with white boxes.*





### 117 3. Material and methods

#### 118 3.1. Field measurements

119 We used drone-based structure from motion photogrammetry to compute a high-resolution  
120 digital elevation model (DEM) of the Šuoššjávri peat plateau. Aerial imagery was acquired thanks to a  
121 digital camera mounted on a drone during two flights in September 2015 and 2018 (02/09/2015 and  
122 05/09/2018). Flights were done 120 m above the ground and the photo acquisition presented a side  
123 overlap of 40 % and a forward overlap of 80 %. The horizontal and vertical coordinates of natural and  
124 artificial ground control points were acquired with a differential GPS (dGPS) to support the topographic  
125 reconstruction. Images were processed, including the ground control points, using the Agisoft Photoscan  
126 software (version 1.2.6). The final digital elevation models have a grid resolution of 0.1 m. To guarantee  
127 a meaningful subsidence signal, we only considered subsidence values exceeding 5 cm in this study.

128 To frame our modeling approach with field constraints, we measured snow depth on the  
129 Šuoššjávri peat plateau. Snow measurements for winter 2016 were presented in Martin et al. (2019).  
130 Here we extend these measurements to winters 2017 and 2018 for the plateau top part only. Snow depth  
131 was measured in the end of March with an avalanche probe at precise locations covering the top of the  
132 plateau, as defined in Martin et al. (2019). The exact location of each measurement point under the snow  
133 was localized using a dGPS system to ensure a 5–10 cm horizontal accuracy.

#### 134 3.2. Quantification of thermal erosion patterns

135 Our drone-based photogrammetric approach (Sect. 3.1) enabled us to derive repeat DEMs of the  
136 peat plateau for September 2015 and 2018. We computed the elevation difference (elevation from 2018  
137 minus elevation from 2015) to quantify the spatial pattern of elevation changes of the plateau. From this  
138 elevation difference map, we selected eight degrading zones (later referred as *edge transect areas*)  
139 presenting a 10-30 m long and roughly straight lateral extent for comparison with modeling results (Sect.  
140 3.3). Based on the changes of elevation and lateral extent of the plateau between 2015 and 2018, we  
141 used the eight edge transect areas (Fig. 2) to calculate the subsided volume per meters of circumference  
142 and per year ( $\text{m}^3 \text{m}^{-1} \text{yr}^{-1}$ ). Because elevation changes occurred in the mire due to water level variations





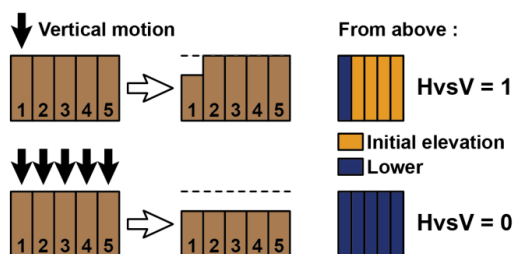
143 between the two dates, we relied on an estimation of the elevation of the plateau edge inflection point  
 144 (around 309.7-309.8 m asl, yellow color on Fig. 2B) to delimitate the plateau from the mire and thus  
 145 identify elevation changes associated to the plateau only.

146 Permafrost degradation creates different subsidence patterns depending on the size of the  
 147 permafrost landform. Small structures like palsas tend to sink entirely from the edge to the top, while  
 148 peat plateaus show stability of their top part and pronounced lateral retreat. To distinguish between these  
 149 two types of thermal erosion patterns, we introduce a so-called *Horizontal vs Vertical* (HvsV) shape  
 150 index that we can apply to both field observations and model results. The basic concept of the HvsV  
 151 index is illustrated in Fig. 3. To compute the HvsV index, the plateau edge elevation is averaged over  
 152 five points, from its very base ( $z_1$ ) to the first meters where its flat top is reached ( $z_5$ ). The elevation  
 153 difference between 2015 and 2018 ( $\Delta z$ ) for  $z_1, z_2, z_4$  and  $z_5$  is then used as follow:

$$154 \quad HvsV \text{ shape index} = \max \left( 0, 1 - \frac{\Delta z_4 + \Delta z_5}{\Delta z_1 + \Delta z_2} \right) \quad (1)$$

155 For the field observations, the HvsV index was obtained by first laterally averaging the slope of  
 156 each edge transect area using five to fourteen parallel elevation profiles across the zone, for the 2015  
 157 and 2018 DEMs. These two synthetic elevation profiles were then averaged into the five required points  
 158 (from the edge to the top of the plateau). The HvsV index was then calculated based on these synthetic  
 159 elevation differences (between 2018 and 2015).

160 For the simulation results (Sect. 4.2), a 10 m long window was used to capture the topography  
 161 from the base of the plateau to its flat top. These 10 m profiles were then averaged into the five required  
 162 points. Then, elevation differences were observed over three-year-long time periods to compute the  
 163 HvsV index.



164  
 165 *Figure 3. Presentation of the HvsV index used to quantify the observed and simulated degradation types. The index*  
 166 *evolves from 1 when the plateau undergo pure lateral degradation (subsidence restricted to the areas at contact*  
 167 *with the mire) and 0 when it experiences a global collapse (subsidence of all the areas).*



### 168 3.3. Modeling climate-induced changes of peat plateau topography

#### 169 3.3.1 The CryoGrid3 model

170 We simulate the ground thermal regime and related topographic evolution of the Šuoššjávri peat  
171 plateau using the CryoGrid model (Westermann et al., 2016). CryoGrid is a land surface model designed  
172 for permafrost modeling. It consists of a physically based description of 1D heat transfer in the soil  
173 column, including freeze–thaw process of soil water/ice. The model features a simple snowpack module,  
174 which includes heat conduction, dynamic buildup, melt, sublimation, water infiltration, and refreezing.  
175 CryoGrid uses a surface energy balance module to calculate the ground surface temperature. The  
176 turbulent fluxes of sensible and latent heat are calculated using a Monin – Obukhov approach (Monin  
177 and Obukhov, 1954). Evapotranspiration is adjusted to soil moisture and the water uptake is distributed  
178 vertically so that it decreases exponentially with depth. The soil moisture computation along the soil  
179 column relies on a bucket scheme (Martin et al., 2019, Nitzbon et al., 2019).

180 CryoGrid represents ground subsidence resulting from the melt of the excess ice in the ground  
181 (thermokarst process, see Westermann et al., 2016). Subsidence calculation is based on soil stratigraphy  
182 (ice content and natural porosity) and modifies the 1D vertical soil mesh when excess ice (soil grids for  
183 which the ice content exceeds the natural soil porosity) melts. This functionality was first implemented  
184 in Westermann et al. (2016) and later used in (Nitzbon et al., 2019, 2020) to represent the transient  
185 evolution of polygonal tundra landscapes caused by permafrost degradation for various hydrological  
186 scenarios. In the present study, this scheme is used to account for the thermal erosion of the peat plateaus  
187 and the microtopography changes induced by permafrost degradation within the plateau.

188 Following Nitzbon et al. (2019, 2020), CryoGrid includes a parallel framework to  
189 simultaneously compute several 1D tiles that can exchange numerical information at defined time steps  
190 (quantities of water, snow and heat). This approach, denoted as *laterally coupled tiling*, allows us to  
191 couple 1D tiles with different stratigraphies/topographies. With this method, the spatial variability tied  
192 to Arctic landscape such as the topographic gradient between the center and the rim of polygonal tundra  
193 (Nitzbon et al., 2019) or the stratigraphy differences between Yedoma and Holocene deposits (Nitzbon  
194 et al., 2020) has been simulated. It also permits to calculate the lateral fluxes of snow, subsurface water



195 and heat that are small-scale key drivers of the ground thermal regime (Nitzbon et al., 2019). In the  
196 present study, laterally coupled tiling is used to simulate the Šuoššjávri peat plateau and its interface  
197 with the surrounding wet mire and represent the subsidence and lateral fluxes along this transition (Sect.  
198 3.2.4).

### 199 3.3.2 Modelling framework and sensitivity analysis to plateau snow depth

200 Snow depth is a major parameter influencing the ground thermal regime (cf. Introduction) and  
201 assessing its influence on the degradation dynamics of peat plateau is one of the objectives of this study.  
202 Because of microtopography-dependent wind-driven snow redistribution, snow depths over the peat  
203 plateaus are usually shallow (10-40 cm, Martin et al. 2019). In the Cryogrid model, the efficiency of this  
204 redistribution is adjusted with a parameter that sets, for each tile, the maximum snow depth which cannot  
205 be transported towards the mire. Snow accumulated beyond this depth is systematically transported by  
206 the model towards the lower-lying tiles, here representing the mire.

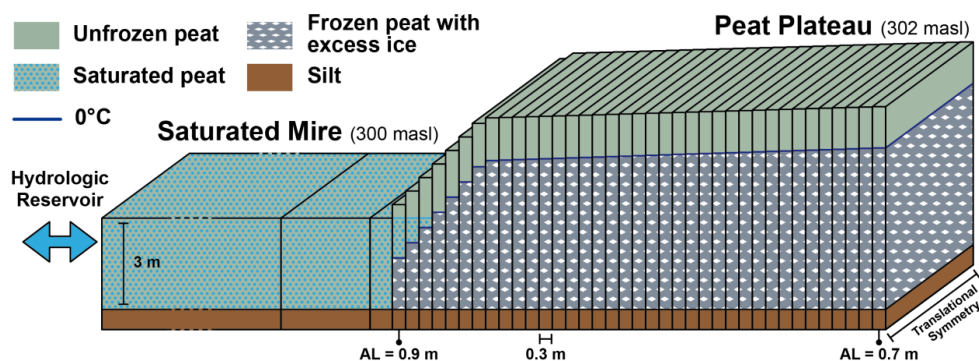
207 The present-day topography of the peat plateau is clear evidence that the long-term evolution is  
208 not only governed by thermal erosion of the plateau edges, but also by a complex interplay of  
209 thermokarst process in the interior of the plateau, with pond formation and drainage, as well as drainage  
210 gully development and deepening. As a full three-dimensional simulation of these phenomena is beyond  
211 the capability of present-day models, we focus modeling on the simplified situation of a laterally  
212 symmetric peat plateau edge. Furthermore, we conduct simulations for steady-state climate forcing from  
213 the period of the field observations (Sect. 3.3.3). This makes it possible to compare the magnitudes of  
214 modeled volumetric plateau degradation over three-year periods with field observations for sufficiently  
215 straight sections of the plateau edge (Sect. 3.2, Fig. 2). As field observations of snow depth reveal a  
216 considerable spread of snow depths values on the plateau (that cannot be reproduced by modeling), we  
217 investigate the sensitivity of modeled topography changes towards snow depths on the plateau (which  
218 can be controlled by the above mentioned model parameter), using a realistic range of values between  
219 which snow depths vary during the simulations (Sect. 3.1). We therefore designed four simulations  
220 called: *0 cm snow*, *5-10 cm snow*, *10-20 cm snow* and *20-30 cm snow*.



221 *3.3.3 Model setup*

222 To simulate the edge thermal erosion of the Šuoššjávri peat plateau, we used the laterally  
223 coupled tiling framework of CryoGrid (Nitzbon et al., 2019) with 40 coupled tiles in a linear  
224 configuration so that they can exchange snow, subsurface water and heat (Sect. 3.2.1). In the third  
225 dimension, translational symmetry is assumed in order to represent the evolution of the 10 to 30 meters  
226 long and roughly straight geometry of these specific zones of the plateau. These edge transect areas are  
227 detailed in Fig. 2.

228 Fig. 4 provides details on the model setup. The wet mire is divided into three tiles with surface  
229 elevation at 300 m asl. They are composed of a 3 m thick layer of unfrozen saturated peat above a 7 m  
230 thick silty mineral layer that also extends below the plateau. Their respective widths are 50, 2 and 0.5  
231 m. These values are significantly higher than those of the tiles initially hosting permafrost to ensure  
232 stable unfrozen lateral boundary conditions in the mire side, as is observed in the plateau mire complex.  
233 The largest and most external mire tile is linked to a hydrologic reservoir (Nitzbon et al., 2019) to ensure  
234 a steady water level in the mire at its surface (at 300 m asl), keeping it always saturated and preventing  
235 ponding of surface water above 300 m asl. The peat plateau tiles are 0.3 m wide, giving a total width of  
236 the plateau of 11.1 m. It stands at 302 m asl, making it 2 m high, in line with observations (Table 1). For  
237 the initial steady state simulation (Sect. 3.2.5), the topographic transition between the mire and the  
238 plateau was initially set as an arbitrary regular slope between 300 and 302 m asl as well as the active  
239 layer depths, which were linearly interpolated between 0.9 m (at the contact with the mire) and 0.7 m at  
240 the other hand of the plateau. The excess ice is distributed homogeneously along depth and the soil grid  
241 ice content is set so that the plateau surface reaches 300 m asl (mire surface elevation) when the excess  
242 ice is fully melted.



243  
244 *Figure 4. Setup used to simulate peat plateau degradation. We coupled 40 CryoGrid tiles to reproduce the contact*  
245 *between the mire and the peat plateau. The AL occurrences indicate the active layer used for the CryoGrid tiles*  
246 *with permafrost. We linearly interpolated them between 0.9 and 0.7 m. The model implement lateral fluxes of snow,*  
247 *subsurface water and heat between the tiles as well as ground subsidence due to excess ice melt.*

### 3.3.4 Steady state climatic forcing and model spin-up

248  
249 As presented in Martin et al. (2019), we obtained the forcing data for the model with a dynamical  
250 downscaling of the ERA Interim reanalysis data (Dee et al., 2011). We used for this Weather Research  
251 and Forecasting model (WRF v.3.8.1; Skamarock and Klemp, 2008) with an option set as in Aas et al.  
252 (2016) with the exceptions mentioned in Martin et al. (2019). We used the nearest grid points in the 3-  
253 km domain to derive the forcing data for CryoGrid. In Finnmark, the hydrological year 2015–2016  
254 revealed itself to be particularly warm. Both 2015 and 2016 ranged 2 to 3°C above the 1961 - 1990  
255 normal period. Additionally, the period was wetter than average, with 2015 and 2016 being, respectively,  
256 33 and 50 % wetter than the normal period. As such, the simulated period gives a good opportunity to  
257 study the response of ground surface temperature to an anomalously warm and wet 12 month-long  
258 climatic conditions (Grinde et al., 2018; Heiberg et al., 2017). This is of particular interest for this study  
259 because both of these two factors enhance permafrost degradation (Sect. 1).

260 The forcing data are looped to generate 100-year time series with steady-state climate forcing.  
261 To achieve a realistic initial temperature profile also in deeper layers, a 100-year spin-up is performed  
262 for all simulations using the 0 cm snow scenario, for which the peat plateau is stable (Sect. 4.2). Note  
263 that the other snow scenarios cannot be used for model spin-up, as the plateau edge starts to retreat  
264 instantly, so that a true steady-state cannot be reached.



265 *3.3.5 Input parameters*

266 Similar to Martin et al. (2019), the soil stratigraphy used in the model is based on analyzed soil  
267 samples from the peat plateau site of Iškoras located 40 km east from Šuoššjávri, the site of the present  
268 study. Consistently with Kjellman et al. (2018), the results we obtained consist of a 3 meters thick peat  
269 soil layer with 5 % of mineral and 15 % of organic material in total volumetric content (and thus a 80 %  
270 porosity), underlain by a 7 m saturated mineral silt layer with 50 % porosity, above a mineral bedrock  
271 layer (3 % porosity, as in Westermann et al., 2013).

272 Over the Šuoššjávri plateau, the soil stratigraphy presents a significant spatial variability. While  
273 the above described stratigraphy matches the western parts of the plateau, on the eastern and southern  
274 parts, organic soil thickness is limited by the near surface underlying morainic deposits. Doing manual  
275 soil probing, we encountered mineral soil within the first meter in the close vicinity of the plateau. Such  
276 stratigraphy is not prone to subsidence and for this reason the modeling work presented here focuses on  
277 reproducing the subsidence patterns observed on the western part of the peat plateau where edge already  
278 started retreating (Borges et al., 2017).

279 Snow and soil parameters are based on the field measurements and the sensitivity tests from  
280 Martin et al. (2019). The simulations use a snow density of  $230 \text{ kg m}^{-3}$ , consistent with measurements  
281 on top of the peat plateaus. Similarly, the soil field capacity used for the simulations is set to 0.55 (in  
282 terms of volumetric water content). Peat soil field capacity can display a pronounced variability (20 to  
283 60 % of the volumetric content; Walczak and Rovdan, 2002) and our value is consistent with field  
284 observation from Southern Siberian peatlands (Motorin et al., 2017). All other parameters (e.g. the  
285 surface energy balance parameterization etc.) were chosen as in Martin et al. (2019).



## 286 4. Results

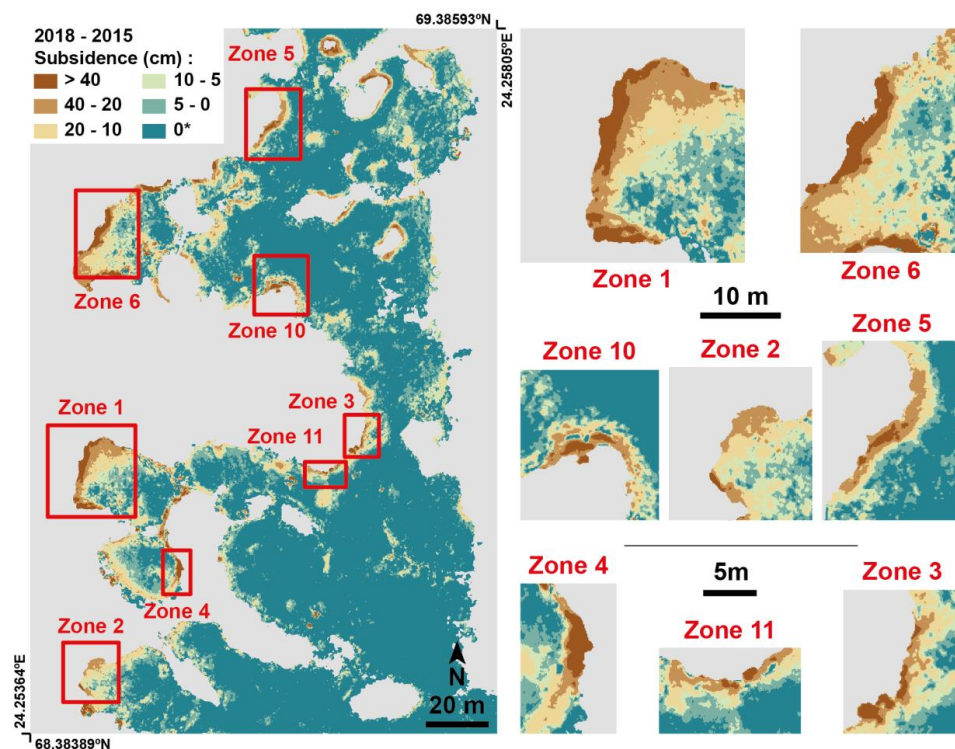
### 287 4.1 Measurements of microtopography evolution

288 The topographic changes of the Šuoššjávri peat plateau between September 2015 and September  
289 2018 are presented in Fig. 5. From the DEM difference, we found that 19 % of the plateau exhibits 5 cm  
290 or more subsidence (i.e. the sensitivity threshold of the measurements), which consequently implies that  
291 81 % of the plateau is stable during the observation period. The mean subsidence value (considering  
292 values larger than 5 cm) is  $17\pm 15$  cm ( $1\sigma$ ) and the median 12 cm, with 1.2 % the total plateau surface  
293 subsiding by more than 40 cm. The maximum observed subsidence is a one square meter patch in Zone  
294 6, exhibiting between 1.5 and 1.7 m of subsidence over the 3 years.

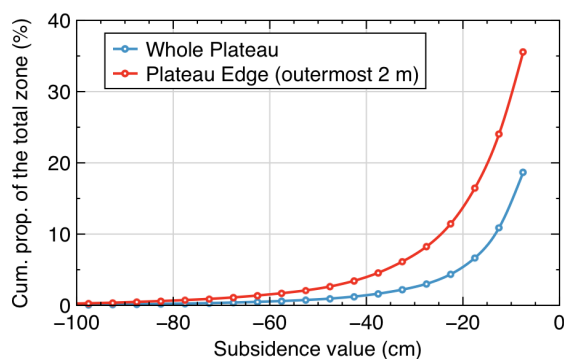
295 We extracted the plateau edge over its outermost 2 m (i.e. the band of the 20 outermost pixels  
296 delimiting the plateau). We find that this surface corresponds to one third of the total plateau surface,  
297 but represents 77 % of the total subsidence (including the rims of the depressions within the plateau).  
298 The distribution of subsidence values for the whole plateau and its edge only are presented in Fig. 6.

299 Due to the differences in peat plateau stratigraphies detailed in Sect. 3.3.3., the west side of the  
300 plateau features higher subsidence values than the east side. On the eastern edge, ground subsidence is  
301 lower due to the limited thickness of the peat layer, with mineral soil at a depth of less than one meter  
302 below the surface. A description of the eight edge transect areas and their subsidence between 2015 and  
303 2018 is presented in Table 1. For these eight zones, the average volumetric loss per year (normalized by  
304 the structure length, i.e. the length of the plateau edge of the different sections) is  
305  $0.13\pm 0.08$  m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup>. The mean HvsV shape index (Sect. 3.2.) is  $0.78\pm 0.08$ , which suggests a  
306 dominance of horizontally- over vertically-driven ground subsidence.





307  
 308 *Figure 5. Subsidence over the Šuoššjávri peat plateau. Left: Overview of the subsidence on the plateau. Note that*  
 309 *the color scale is truncated for subsidence values higher than 40 cm and lower than 0 cm, which, for the latter,*  
 310 *correspond to increase in surface elevation due to vegetation change. Right: Edge transect areas of the plateau*  
 311 *used to compare observed and simulated degradation patterns (Sect. 3.2.1).*



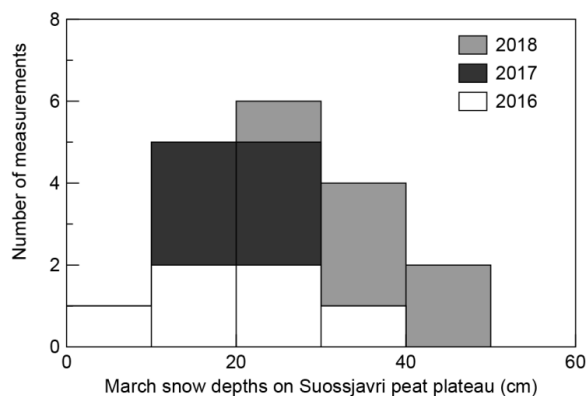
312  
 313 *Figure 6. Subsidence distribution for the Šuoššjávri peat plateau. The blue curve gives the subsidence distribution*  
 314 *for the whole plateau while the red one gives it for the plateau edge only. The edge is taken as the outermost 2 m*  
 315 *of the plateau. Only subsidence values greater than 0.05 m are considered in this graph to guarantee a meaningful*  
 316 *subsidence signal. Data are derived from the topography difference between 2015 and 2018. Percentages indicate*  
 317 *which proportion of the total area (whole plateau or plateau edge) is affected by a subsidence superior (in absolute*  
 318 *value) or equal to a given subsidence value (e.g. 36 % of the edge exhibits a subsidence higher than or equal to 5*  
 319 *cm).*



Zone	Palsa Height (cm)	Structure Length (m)	Subsided volume per structure length (m <sup>3</sup> /m/a)	HvsV Shape Index (-)
1	200	25	0.295	0.74
2	130	16	0.100	0.64
3	120	16	0.074	0.68
4	140	13	0.132	0.81
5	120	25	0.045	0.94
6	180	30	0.232	0.75
10	220	13	0.124	0.81
11	145	11	0.052	0.84
		<b>Mean</b>	<b>0.132</b>	<b>0.78</b>
		<b>Standard Dev.</b>	<b>0.078</b>	<b>0.08</b>

320 *Table 1. Field observations at the 8 edge transect areas presented in Fig. 2. Elevation changes and surface*  
 321 *measurements are derived from the digital elevation models. The HvsV shape index (Fig. 3) was calculated*  
 322 *according to equation (1). See Sect. 3 for more details.*

323 Snow measurement results are presented in Fig. 7. Consistently with Martin et al. (2019), peat  
 324 plateau tops are commonly covered with 10 to 40 cm of snow, with most of the values between 10 and  
 325 30 cm. We used these observations to design the numerical simulations presented in this study and their  
 326 four different snow depths: 0, 5-10, 10-20 and 20-30 cm (Sect. 3.3.2).



327 *Figure 7. March snow depth distribution on top of the Šuoššjávri peat plateau for 2016, 2017 and 2018. Data for*  
 328 *2016 are from Martin et al. (2019).*  
 329

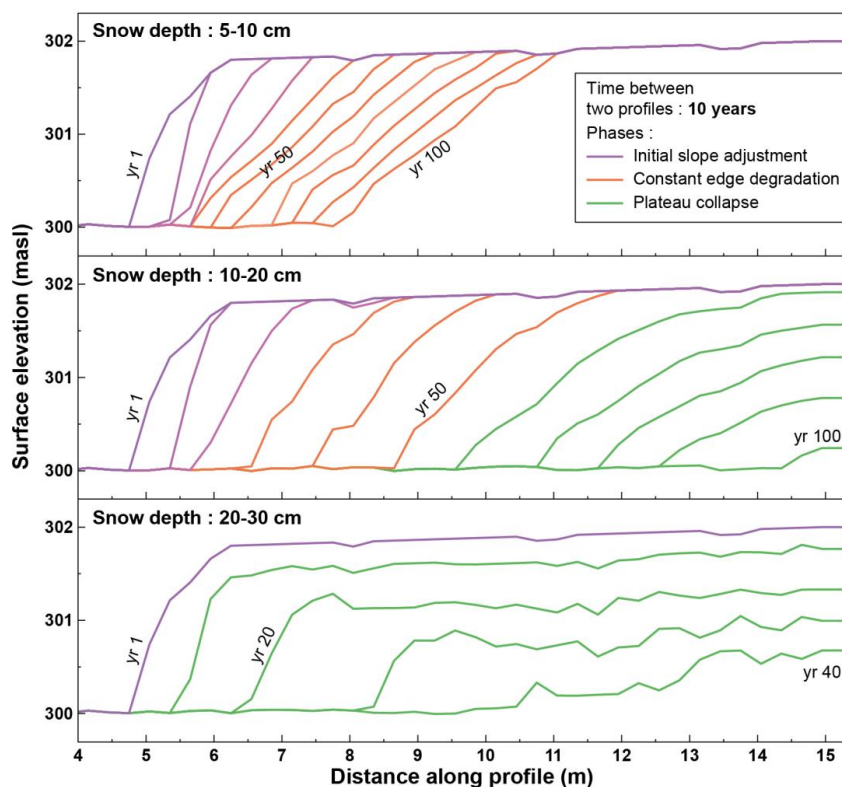
## 330 4.2 Simulations of microtopography evolution

331 Results from the model simulations are presented in Fig. 8. The temporal evolution of the peat  
 332 plateau microtopography shows an edge retreat, while a large part of the plateau is stable, as observed  
 333 on the field. The temporal evolution and patterns of edge retreat exhibit a pronounced dependence on



334 the snow depth on the peat plateau. The *0 cm snow* simulation (with complete transport of the snow  
335 from the plateau towards the mire) shows no thermal erosion of the plateau, whereas the simulation with  
336 the thinnest snow depth (*5-10 cm snow*) triggers an edge retreat of 4 to 5 meters over the 100 years  
337 duration of the simulation. For the *10-20 cm snow* and *20-30 cm snow* simulations, the plateau fully  
338 degrades within the simulation time with notable differences in profile evolution between simulations.  
339 While the plateau fully degrades at the end of the 100 years simulation for the *10-20 cm snow* simulation,  
340 it occurs within 40 years for the *20-30 cm snow* simulation.

341 The topographic evolution observed over the three simulations lead us to identify three different  
342 types of simulated thermal erosion. For both the *5-10 cm snow* and *10-20 cm snow* simulations, the  
343 plateau degradation first shows a phase of slope adjustment during which the slope profile smooths from  
344 year to year. We denote this phase as the “Initial Slope Adjustment” (ISA, Fig. 8 and 9). This phase lasts  
345 for 40 years in the *5-10 cm snow* simulation and 30 years for the *10-20 cm snow* simulation. For the  
346 same two simulations, the thermal erosion later affects the slope in a more uniform way and the plateau  
347 edge retreats with a constant slope at a constant rate; during 60 years for the *5-10 cm snow* simulation  
348 and 20 years for the *10-20 cm snow* one. We denote this phase the “Constant Edge Degradation” (CED,  
349 Fig. 8 and 9). During the second half of the *10-20 cm snow* simulation, both the edge and the top of the  
350 plateau subside. We denote this phase as the “Plateau Collapse” (PC, Fig. 8 and 9). Contrary to the *10-*  
351 *20 cm snow* simulation, the *20-30 cm snow* simulation does not exhibit the phases of initial slope  
352 adjustment and constant edge degradation but only the evolution corresponding to a plateau collapse.



353  
354 *Figure 8. Surface elevation profiles of the peat plateaus as simulated with CryoGrid for different snow depths on*  
355 *top of the plateau, in time increments of 10 years. The evolution between the lines lead us to identify different*  
356 *periods in the plateau degradation: Initial Slope Adjustment (slope modifications along time), Constant Edge*  
357 *Degradation (slope conserved) and Plateau Collapse (subsidence over the full plateau). Note that the “0 cm snow”*  
358 *simulation did not produce any subsidence and for this reason, does not appear on this figure.*

359 The top panel of Fig. 9 presents the volumetric rate loss of the plateau for the three simulations.  
360 For the 5-10 cm snow simulation, this rate is constant around 0.06 to 0.08 m<sup>3</sup> lost per meter of lateral  
361 extension of the plateau (i.e. per meter of plateau circumference) and per year (m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup>) for the whole  
362 simulation. For the 10-20 cm snow simulation, the volumetric loss during the initial slope adjustment  
363 and the constant edge degradation phases shows a steady increase from 0.08 to 0.28 m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup>. During  
364 the plateau collapse phase, this rate steadily decreases to 0.12 m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup> at the end of the simulation.  
365 For the 20-30 cm snow simulation, the rate reaches 0.28 m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup> over the first decades and stabilizes  
366 at this value for 10-20 years before undergoing another rapid increase to 0.35 m<sup>3</sup> m<sup>-1</sup> yr<sup>-1</sup>, at which it  
367 stabilizes until the end of the simulation. Further quantification of the thermal erosion is given in Sect.  
368 4.3. A comparison between simulated and measured ground surface temperatures (temperature logger



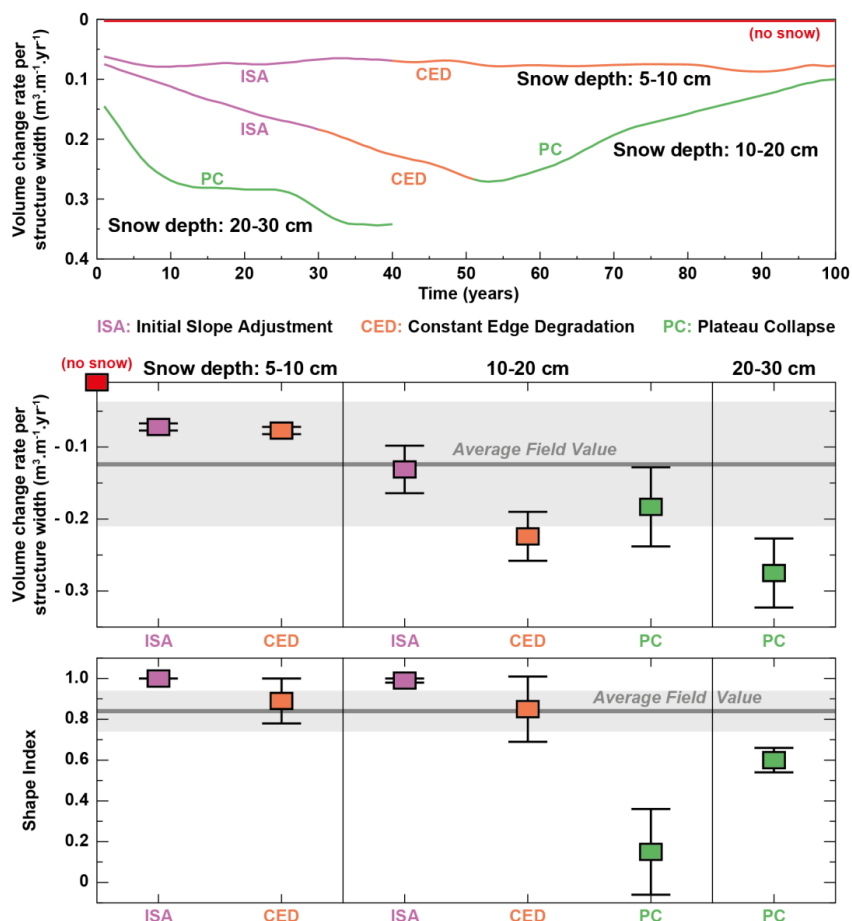
369 time series from Martin et al., 2019) is presented in the supplementary material (Fig. A1), showing an  
370 overall good agreement.

### 371 4.3 Comparison of model results and topographic measurements

372 A comparison of the Šuoššjávri peat plateau thermal erosion patterns between the field data and  
373 the simulations is presented in the two bottom panels of Fig. 9. Field values are displayed in grey and  
374 represent average and standard deviation of the field measurements of the respective measured variables  
375 (Sect. 3.3 and Table 1). For each simulation, we average the volume loss and shape index over the  
376 degradation periods presented in Sect. 4.2 (ISA,CED,PC).

377 Field-based and simulated values are in good agreement regarding volume changes. The mean field  
378 value ( $0.13 \pm 0.08 \text{ m}^3 \text{ m}^{-1} \text{ yr}^{-1}$ ) is compatible with the different degradation phases observed for the *5-10*  
379 *cm snow* and *10-20 cm snow* simulations. The *5-10 cm snow* simulation shows little spread and smaller  
380 values than the average field value ( $< 0.1 \text{ m}^3 \text{ m}^{-1} \text{ yr}^{-1}$  in absolute value) whereas the *10-20 cm snow*  
381 simulation shows greater spread and greater volume changes than the average field value (between 0.1  
382 and  $0.25 \text{ m}^3 \text{ m}^{-1} \text{ yr}^{-1}$ ). The *20-30 cm snow* simulation stands out from this trend and exhibits volume  
383 losses substantially higher than the field values ( $> 0.25 \text{ m}^3 \text{ m}^{-1} \text{ yr}^{-1}$ ).

384 Regarding the HvsV shape index, the Initial Slope Adjustment phases for both the *5-10 cm snow*  
385 and the *10-20 cm snow* simulations exhibit values of 1, slightly larger than the field value ( $0.84 \pm 0.09$ ).  
386 Both Constant Edge Degradation phases are in line with field observations, exhibiting a greater spread  
387 among the simulations than among the field values. Because both are characterized by simultaneous  
388 edge degradation and global subsidence, the two plateau collapse phases (for the *10-20 cm snow* and  
389 *20-30 cm snow* simulations) have HvsV values significantly smaller than the field values ( $< 0.6$ ).



390  
 391 *Figure 9. Top: Volume changes normalized to structure width for the 3 simulations. Bottom: Comparison of the*  
 392 *degradation patterns between observations and simulations. Observations are means and standard deviations of*  
 393 *the variables measured in the eight edge transect areas presented in Table 1. They appear as a grey line and*  
 394 *shading. The values derived from simulations are mean and standard deviations taken over the different periods*  
 395 *of the simulations. As detailed in Fig. 8, the profile evolution allows identifying different periods in the plateau*  
 396 *evolution, namely Initial Slope Adjustment (ISA), Constant Edge Degradation (CED) and Plateau Collapse (PC).*  
 397 *The red square reminds that no subsidence was observed for the “0 cm snow” simulation.*



## 398 5. Discussion

### 399 5.1 Field measurements

400 At the western edge of the Šuoššjávri plateau, subsidence is highly variable, ranging from 0 to more  
401 than 1 meter within 3 years. This pattern highlights the chaotic behavior of permafrost landscapes facing  
402 degradation due to a positive feedback between subsidence, snow accumulation and water drainage.  
403 When an initial perturbation, such as intense rainfall or extraordinary snow accumulation triggers  
404 subsidence (Seppälä, 1988, 2011), both the snow redistribution and the subsurface drainage towards the  
405 mire are affected, which creates warmer surface conditions and, in return, triggers more subsidence.  
406 Considering the complex geometry of the Šuoššjávri plateau edges, meter-scale variability of the snow  
407 and hydrological conditions likely contribute to observed variability of ground subsidence. Besides, due  
408 to the dependence of heat transfer to the surface area of the interface between the mire and the plateau,  
409 the geometry of edges affects degradation speed. As such, zones 1, 2, 4 and 6 belong to convex features  
410 of the plateau edges and show particularly high subsidence rates.

411 Our results confirm that edge degradation is a major degradation pathway of peat plateaus with 77 %  
412 of the total subsidence occurring within the outermost 2 meters of the Šuoššjávri plateau. This result  
413 shows consistency with Jones et al. (2016) who reported that 85 % of the degradation of forested  
414 permafrost plateaus was due to lateral degradation along the margins. Estimating the elevation of the  
415 inflection point of the plateau edge from the DEMs allows to quantify the plateau surface area at a given  
416 time, even though in practice the precise positioning of such limit can be discussed. Between 2015 and  
417 2018, we find that the Šuoššjávri plateau lost 3.2 % of its surface area. If we take the percentage of  
418 annual loss rate of plateau area as  $100 * (1 - S_{i+1}/S_i)$  (where the fraction corresponds to the ratio between  
419 the area at year  $i$  and the area at year  $i+1$ ), then over an observation period of  $n$  years, the average ratio  
420 can be expressed as:

$$421 \quad r = 100 \times \left( 1 - \sqrt[n]{\frac{S_n}{S_0}} \right) \quad (2)$$

422 Where  $r$  is the annual loss rate in %  $\text{yr}^{-1}$ ,  $n$  is the number of year between the 2 observations,  $S_n$  the  
423 plateau surface at the end of the observation period and  $S_0$  the plateau surface at the beginning of the





424 period. Applying Eq. 2, the aerial change in this study corresponds to a surface loss rate of 1.1 % yr<sup>-1</sup>.  
425 Reconstructing the Šuoššjávri peat plateau extent from 1956 to 2011 with aerial imagery, Borge et al.  
426 (2017) observed annual loss rates that they compared to the 1956 extension. Using Eq. 2, we compute  
427 the annual loss rate from their data to be 0.5 % yr<sup>-1</sup> from 1956 to 1982, 0.8 % yr<sup>-1</sup> from 1982 to 2003  
428 and of 1.4 % yr<sup>-1</sup> from 2003 to 2011. Hence, our retreat rate is in good agreement with the long-term  
429 edge retreat rates, although the two values cannot be compared in a strict sense, since Borge et al. (2017)  
430 also included small palsas in the surrounding area (which show faster degradation rates than the peat  
431 plateau) in their assessments.

## 432 5.2 Model results

### 433 5.2.1 The plateau degradation

434 Our modeling framework relies on an idealized geometry and steady-state climate forcing, so the  
435 full variety of the observed degradation patterns cannot be reproduced. However, the comparison  
436 between model results and observations clearly shows that the numerical model framework can capture  
437 the correct order of magnitude of the degradation processes, while also reproducing key patterns in the  
438 observed ground temperature regime (supplementary material, Fig. A1).

439 Among the different degradation phases (Initial Slope Adjustment, Constant Edge Degradation and  
440 Plateau Collapse), the ISA and PC phases are less relevant than the CED phase for model to field  
441 comparisons: the ISA phase is strongly affected by the initial response of the model to the applied  
442 climate forcing and snow depths on the plateau, whereas the CED phase corresponds to the prolonged  
443 edge retreat observed in the field, while the bulk of the peat plateau remains stable. The PC phase  
444 corresponds to the sustained collapse of the plateau with ground subsidence in all parts, which is not  
445 observed for the Šuoššjávri peat plateau, but regularly occurs for smaller circular palsas in the vicinity.  
446 For such palsas, the assumption of translational symmetry inherent in the model setup is not valid and  
447 simulations should be performed assuming rotational rather than translational symmetry. This suggests  
448 that our simulations are indeed most realistic during the CED phase, whereas changes of the peat plateau  
449 shape need to be taken into account to model the final stages of degradation.



450                                   5.2.2 *The role of snow*

451           Our simulations confirm the crucial role of snow on the ground thermal regime and peat plateau  
452 degradation. The sensitivity of modeled ground temperatures towards snow depth is in good agreement  
453 with the field measurements for the Šuoššjávri and the Iškoras peat plateau (40 km east of Šuoššjávri)  
454 presented by Martin et al. (2019). For example, they showed that the measurement points on the plateau  
455 exhibiting a mid-march snow depth smaller than 10 cm were associated with coldest mean annual  
456 ground temperatures, and that 70 % of the plateau locations with stable permafrost had a March snow  
457 depth smaller than 30 cm.

458           However, our idealized model approach assumes snow depths to be constant on the peat plateau,  
459 which does not capture the significant spatial and interannual variability of snow depths on the plateau  
460 observed in measurements (Fig. 7). In particular, the complex geometry of snow drift patterns (snow  
461 accumulation) along the plateau edges, with snow drifts forming on lee sides, is not captured by the  
462 simple snow redistribution model implemented in CryoGrid. Field observations show that snow drifts  
463 along the plateau edges feature considerably higher snow depths than the surrounding wet mire, thus  
464 introducing additional winter warming in the zone of maximum change. Additionally, persistent wind  
465 patterns can strongly influence the distribution of snowdrifts. Unfortunately we do not possess field  
466 measurements to discuss this parameter at the Šuoššjávri site. In CryoGrid, on the other hand, snow  
467 removed from the plateau is evenly distributed over the entire mire, not taking edge effects into account.

468           Furthermore, our model assumes a fixed value for the snow density and thus snow thermal  
469 properties, while this parameter in reality varies with e.g. snow depth and time, responding strongly to  
470 synoptic conditions and imposing metamorphosis of the snowpack. Measurements of snow density in  
471 Šuoššjávri showed that the snow on top of palsas is slightly less dense than in the mire. This could be  
472 due to a thinner snowpack leading both to greater kinematic metamorphism and the formation of depth  
473 hoar, notorious for high porosity and a real difficulty to measure density and high effective thermal  
474 conductivity (Domine et al., 2016). A thinner snowpack also implies a lower overburden load and  
475 therefore less compaction. Such limitations could be moderated by using more sophisticated snow  
476 models taking snow microphysics and the transient evolution of snow density into account, such as  
477 CROCUS (Vionnet et al., 2012) or SNOWPACK (Bartelt and Lehning, 2002). Yet, even these models



478 show limitations to reproduce the thermal characteristics of snow deposited in Arctic regions because  
479 they do not account for the vapor fluxes in the snow pack, which significantly affect the snow thermal  
480 conductivity profile (Domine et al., 2016).

## 481 5.3 Implications for climate driven permafrost landscapes changes

### 482 5.3.1 Sensitivity to climate forcing and perturbations

483 In this study, we present idealized numerical simulations of peat plateau thermal erosion that  
484 reproduce the general patterns of edge retreat as observed in-situ through repeat digital elevation models.  
485 We demonstrate that the snow depth on the plateau is a strong control for subsidence patterns and  
486 dynamics. This result indicates that peat plateau systems will react sensitively to changes in the applied  
487 climate forcing, not only regarding temperature but also regarding (snow) precipitation and windspeed  
488 variations which all affects in turn snow pack building. In this regard, predictions regarding snow fall  
489 for the coming decades are complex. An overall decrease of mean snowfall is expected at the global  
490 scale (O’Gorman, 2014), consistently with observed trends over the past decades (Liston and Hiemstra,  
491 2011) but this decrease will be accompanied by strong regional trend (Brown and Mote, 2009; Lader et  
492 al., 2020) and the frequency and distribution of extreme snowfall remain unclear (O’Gorman, 2014). In  
493 any case, it appears that for future modelling works, the accuracy of both the snowpack buildup and its  
494 thermal properties at the small-scale (10-100 m) should be considered of major importance to robustly  
495 simulate permafrost degradation.

496 Additionally, our implementation illustrates the idea of small-scale feedback mechanism on  
497 permafrost degradation. The feedback between the dynamical microtopography and the lateral fluxes  
498 shows how a limited increase in snow cover (when comparing the 10-20 cm snow and the 20-30 cm  
499 snow simulations) results in a dramatically faster degradation rate. Such a sensitivity to minor  
500 perturbation resulting into major modifications of permafrost degradation finds consistency with the  
501 observed permafrost destabilization when punctually augmenting the snow depth with a fence (Hinkel  
502 and Hurd, 2006), when implanting linear road infrastructures (Deimling et al., 2020) or due to the traffic  
503 of heavy vehicle in Alaskan lowlands (Raynolds et al., 2020).



504 *5.3.2 Spatiotemporal stability and degradation conditions*

505 Our simple approach is clearly not able to capture the complex patterns of different subsidence rates  
506 that are observed around the edges of the plateau (Fig. 5). On top of small-scale variations of ground  
507 stratigraphy, excess ice contents and plateau heights, we suggest that the irregular plateau outline with  
508 both concave and convex shapes also affects the lateral fluxes of heat, water and snow, which in turn  
509 exert a control on the edge dynamics. While computationally demanding, our multi-tile approach could  
510 be embedded in an ensemble framework to represent a range of edge geometries and other critical  
511 parameters, yielding a range of different degradation scenarios and therefore capture the high spatial  
512 variability of subsidence at the plateau scale.

513 Over longer time scales, future studies should clarify if the simple multi-tile setup can capture  
514 climate-induced changes of peat plateau stability, for example between the cold Little Ice Age (when  
515 most of the present-day peat plateaus were formed, Kjellman et al., 2018) and the warmer conditions of  
516 the 20<sup>th</sup> century during which peat plateaus in Finnmark likely entered their current state of accelerating  
517 degradation (e.g. Borge et al., 2017).

518 *5.3.3 Permafrost modeling with Land Surface Models*

519 Most Land Surface Models (LSMs) used to simulate the future response of permafrost to climate  
520 changes still rely on simplified one-dimensional implementations of permafrost thaw dynamics which  
521 ignores subsidence and only reflects gradual top-down thawing of the frozen ground (Andresen et al.,  
522 2020). Excess ice melt and the resulting microtopography changes exert a major control on the evolution  
523 of hydrologic conditions, which in turn greatly influence the timing of permafrost degradation, as  
524 demonstrated for polygonal tundra (Nitzbon et al., 2019, 2020). Aas et al., (2019) presented a similar  
525 approach for peat plateaus in Northern Norway. It is based on two tiles (one for the wet mire, one for  
526 the plateau) and reproduces both climate-induced stability and degradation. However, in this approach,  
527 the plateau subsides as a whole when a climate-related threshold is exceeded and excess ice begins to  
528 melt. This contrasts with our field observations, which show ongoing edge retreat on decadal timescales,  
529 while the plateau interior is largely stable. Our approach uses a larger number of tiles to explicitly  
530 represent the temperature and soil moisture gradients across the plateau edges, which causes excess ice



531 melt to only occur in a narrow zone at the plateau edge, in agreement with our observations. Over longer  
532 timescales, on the other hand, this process leads to the reshaping and finally the complete collapse of  
533 the entire peat plateau. In ESMs, implementing a multi-tile approach is challenging due to its complexity  
534 and computational demands. Yet, parameterized approaches could likely be developed, based on  
535 sensitivity tests using our framework. In particular, future studies should investigate to what extent the  
536 two-tile approach demonstrated by Aas et al. (2019) can emulate the results of the multi-tile model,  
537 especially when averaging the results over the range of climatic conditions under which peat plateaus  
538 occur in the sub-Arctic domain. However, a multi-tile setup would probably reveal different  
539 hydrological regimes (zones of well-drained plateaus would remain until the end), which would in turn  
540 affect carbon decomposition.

## 541 6. Conclusion

542 We present and compare field measurements and numerical modeling of thermal erosion patterns  
543 of the Šuoššjávri peat plateau in Northern Norway. We use high resolution digital elevation models  
544 derived from drone-based photogrammetry to quantify changes of surface elevations of the plateau  
545 between September 2015 and 2018. Thermal erosion of the plateau edges is the main process through  
546 which thermal erosion occurs and accounts for 80 % of the total measured subsidence (in terms of  
547 subsided soil volume), while most of the total plateau surface exhibits no detectable subsidence. We  
548 show that this retreat corresponds to a volumetric loss of  $0.13 \pm 0.07 \text{ m}^3 \text{ m}^{-1} \text{ yr}^{-1}$  for the edge transect  
549 zones we studied.

550 Using the CryoGrid land surface model we show that these degradation patterns can be reproduced  
551 numerically in a framework that implements lateral redistribution of snow, subsurface water and heat,  
552 as well as excess-ice-melt-triggered subsidence. Overall, the modeled volumetric change rates are in the  
553 same order of magnitude as the measurements. Based on a steady-state climate forcing, our simulations  
554 demonstrate the importance of lateral snow transport and resulting snow depths on the plateau. The  
555 modelled peat plateau is fully stable when all snow on its top is transported towards the mire (0 cm snow  
556 depth on the plateau), whereas its edges degrade at increasing rates with increasing snow depths. While



557 a maximum of 5-10 cm on the plateau only triggers a 4-5 meters edge retreat within 100 years, a 10-20  
558 cm cover fully degrades the plateau in 100 years and a 20-30 cm cover degrades it in 40 years. Our  
559 simulations reproduce the observed lateral edge degradation with a stable plateau top, but the final phase  
560 of plateau degradation corresponds to complete plateau collapse with subsidence occurring throughout  
561 the entire plateau.

562 These results highlight the fast and high spatial variability of permafrost landscape evolution in  
563 response to climate change. They also show that the related microtopographic, thermal and hydrological  
564 modifications can be represented in numerical models, opening the way for substantial improvements  
565 in simulating permafrost landscape evolution and its impact on greenhouse gas emissions from thawing  
566 permafrost.

567 **Author contribution.** L.M. and S.W. designed the study and conducted the numerical simulations. L.M.  
568 led the manuscript preparations. S.W., M.L., J.N. and L.M. contributed to the model development. T.E.,  
569 S.W., L.M., J.S. acquired field data. K.A. provided forcing data. L.M. and S.F. analyzed field data. All  
570 authors contributed to result interpretation and to manuscript preparation.

571 **Code availability.** The model code and settings used for the simulations is available from  
572 [github.com/CryoGrid/CryoGrid3/tree/xice\\_mpi\\_palsa\\_newsnow](https://github.com/CryoGrid/CryoGrid3/tree/xice_mpi_palsa_newsnow). It will be permanently deposited upon  
573 acceptance of the manuscript. The code is published under the GNU General Public License v3.0.

574 **Data availability.** Field data will be permanently deposited on [archive.sigma2.no](https://archive.sigma2.no) upon acceptance of  
575 the manuscript.

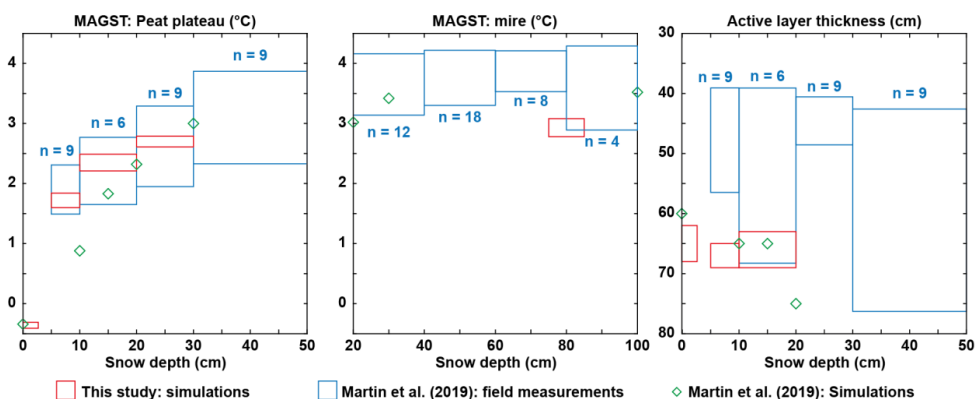
576 **Competing interests.** The authors declare that they have no conflict of interest.

577 **Acknowledgements.** This work was funded by PERMANOR (Norwegian Research Council,  
578 KLIMAFORSK program, NFR project 255331), Nunataryuk (EU grant agreement no. 773421) and the  
579 Department of Geosciences of the University of Oslo, Norway. WRF simulations were performed on  
580 the Abel high- performance computing facility with resources provided by the Department of  
581 Geosciences of Oslo University. The land surface simulations were performed on resources provided by  
582 UNINETT Sigma2 - the National Infrastructure for High Performance Computing and Data Storage in  
583 Norway The authors declare that they have no conflicts of interests. All data are available in the  
584 manuscript or the Supplementary Material.



## 585 Appendix A: Model output comparison with field measurements

586 Fig. A1 compares the Mean Annual Ground Surface Temperatures (MAGST) and the Active  
587 Layer Thickness (ALT) as they are simulated in this study and in Martin et al. (2019) with field  
588 measurements from the same study. Overall, both simulation works yield similar results and show good  
589 agreement with field measurements.



590 **□ This study: simulations**    **□ Martin et al. (2019): field measurements**    **◇ Martin et al. (2019): Simulations**  
591 *Figure A1. Mean Annual Ground Temperature (MAGST) and Active layer thickness as they are simulated in this*  
592 *study and in Martin et al. (2019) compared to the field measurements from Martin et al. (2019) for the same region.*  
593 *Values indicated with the letter n correspond to the number of field observations in Martin et al. (2019).*

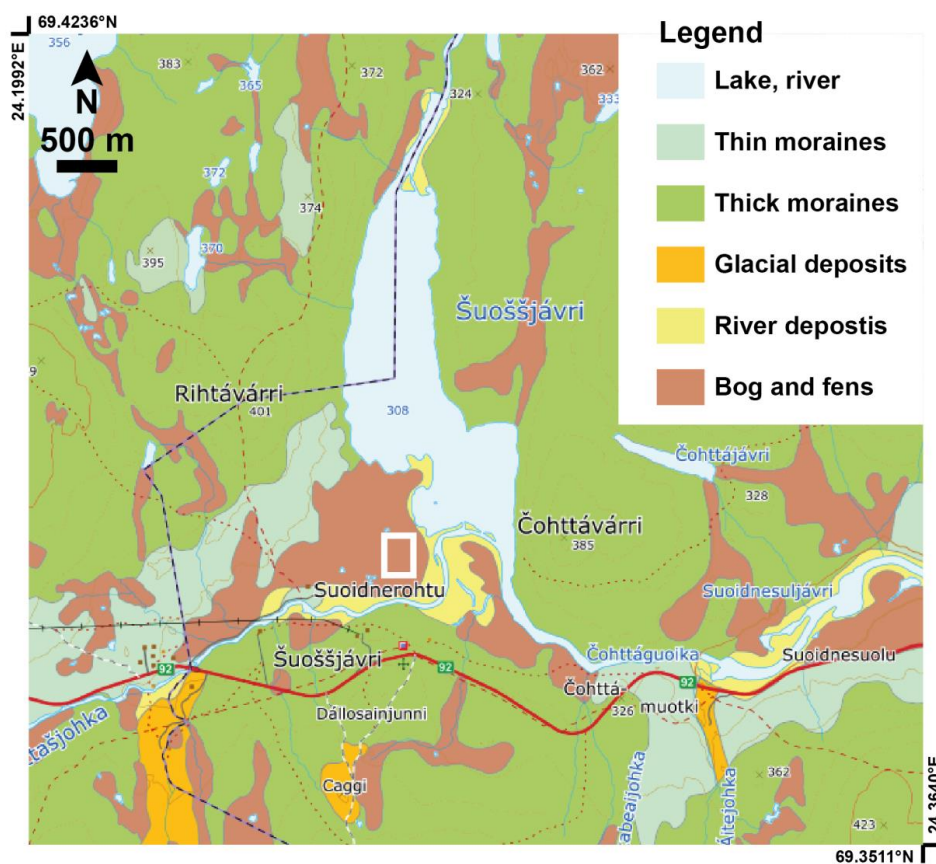
594 While MAGSTs from the present study are in perfect agreement with the measured ones on the  
595 peat plateau, the model seems to underestimate them slightly (0.5°C too cold) in the mire. This is  
596 mainly due to a too low water inflow from the reservoir which does not manage to always fully saturate  
597 the wet mire. As a consequence, the uppermost layer of the mire partially dries during summer, which  
598 shifts the heat flux during autumn from latent to sensible and imposes colder temperatures in the mire  
599 during winter.

600 Active layers are overestimated by 20 to 30 cm for snow depths smaller than 10 cm. This  
601 probably arises from the difference between the real and simulated snow conditions. While real  
602 conditions can be erratic and show important variations during winter (from snow free to snow covered,  
603 back and forth), snow scenarios in the simulations are smoother and show a prolonged covered  
604 conditions, leading to deeper ALT.





## 605 Appendix B: Geomorphological settings



606  
607 *Figure A2. Geomorphological map of the surroundings of the study site. The white rectangle indicates the study*  
608 *site. Source: Geological Survey of Norway (NGU).*

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