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A statistical approach to represent small-scale variability of permafrost temperatures due to snow cover

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In permafrost environments exposed to strong winds, drifting snow can create a small-scale pattern of strongly variable snow heights which has profound implications for the thermal regime of the ground. Arrays of 26 to more than 100 temperature loggers were installed to record the distribution of ground surface temperatures within three study areas across a climatic gradient from continuous to sporadic permafrost in Norway. A variability of the mean annual ground surface temperature of up to 6°C was documented within areas of 0.5 km². The observed variation can to a large degree be explained by variation in snow height. Permafrost models employing averages of snow height for grid cells of e.g. 1 km² are not capable of representing such sub-grid variability. We propose a statistical representation of the sub-grid variability of ground surface temperatures and demonstrate that a simple equilibrium permafrost model can reproduce the temperature distribution within a grid-cell based on the distribution of snow heights.

1 Introduction

In permafrost areas, the seasonal snow cover is a crucial factor for the ground thermal regime (e.g. Luetschg et al., 2004, 2008; Ishikawa, 2003; Goodrich, 1982; Zhang et al., 2001). The main influences of the snow cover are the low thermal conductivity, the high surface albedo and the delay of active layer thawing during snowmelt in spring or summer (Harris et al., 2009). Snow heights above 0.6 to 0.8 m have been found to effectively insulate the ground from the atmosphere (e.g. Haeberli, 1973), and the importance of the timing of the early-winter snow has been emphasized in a number of studies (Zhang, 2005; Zhang et al., 2001; Luetschg et al., 2008; Gubler et al., 2011; Vonder Muhll et al., 1998). Snow redistribution caused by wind drift creates a pattern of different snow heights resulting in spatially variable ground temperatures (e.g. Isaksen et al., 2011; Farbrot et al., 2011). This effect is of great importance in mountain

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permafrost areas due to higher wind speeds, rough topography and open terrain. The strong redistribution of snow caused by wind drift accumulates most of the snow in deep hollows, while large areas are bare-blown.

The heterogeneity of the ground surface temperatures (GSTs) must be systematically quantified as an important background for understanding meso-scale variability of near-surface ground temperatures (Etzelmüller, 2013). In most applications in permafrost areas, grid-based numerical models calculating ground temperature or active layer thickness operate on spatial scales too coarse to resolve the variability of snow heights. For the Norwegian mainland, the ground thermal models CryoGRID 1 and CryoGRID 2 have been implemented at 1×1 km spatial resolution, at which they are capable of capturing the regional distribution of permafrost (Gisnås et al., 2013; Westermann et al., 2013). One of the principal uncertainties is the effect of sub-grid variation in snow cover (Langer et al., 2013) so that procedures for downscaling of snow heights are required to get a satisfactory representation of the ground thermal regime.

In this study, we present a proof-of-concept approach for a statistical representation of ground temperatures based on the distribution of snow heights within a grid cell, using the simple equilibrium permafrost model CryoGRID 1. The concept is validated for three study areas along a transect from continuous permafrost on Svalbard to sporadic permafrost in Southern Norway, where arrays of temperature loggers facilitate to estimate the true distribution of GSTs.

2 Study areas

In three areas in Norway, arrays of temperature loggers measuring the ground surface temperature have been implemented (Fig. 1): Ny-Ålesund, Svalbard ($78^{\circ}55' \text{ N}$, $11^{\circ}50' \text{ E}$), a high-Arctic site in the continuous permafrost zone; Juvvasshøe, Jotunheimen ($61^{\circ}41' \text{ N}$, $8^{\circ}23' \text{ E}$), a high mountain site in central southern Norway located in the continuous to discontinuous permafrost zone; and Finse, Hardangervidda mountain plateau ($60^{\circ}34' \text{ N}$, $7^{\circ}32' \text{ E}$), a low alpine site in southern Norway located in the

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sporadic permafrost zone. A summary of field locations and climatic settings is given in Table 1.

Ny-Ålesund (10–50 m.a.s.l.) is located on the Brøgger peninsula at the west coast of Spitsbergen, and has a maritime climate with cool summers and relatively mild winters.

It is located within the continuous permafrost zone, with a mean annual air temperature (MAAT) during the hydrological year 2012/2013 of -3.8°C , 2.5°C higher than the long-term average in the period 1961–1990 (hereafter denoted the normal). The annual precipitation was 464 mm according to the climate database of the Norwegian Meteorological Institute (eKlima, 2013), 16% higher than the normal. However, a substantial undercatch of measured solid precipitation must be assumed in these climatic conditions, so that the true precipitation is probably 50% higher than the measured (Førland and Hanssen-Bauer, 2000). The seasonal snow cover normally lasts from mid September to well into June, with inter-annual variations in the length of the snow season from 215 to 315 days (Winther et al., 2002). The average snow height in the terrain is around 60 cm, with variations from 0 cm to 300 cm due to wind drift (Bruland et al., 2001). The prevailing wind direction is southeasterly along the Kongsfjord, with generally gentle wind speeds. The average wind speed was 3.5 m s^{-1} during winter 2012/2013, with a maximum wind speed of 20.8 m s^{-1} (eKlima, 2013). The Ny-Ålesund field site is located two kilometers southwest of Ny-Ålesund, half way between the fjord and the terminus of the nearest glacier Brøggerbreen, in gentle topography ranging from 10 to 50 m.a.s.l. The study area is located around the Bayelva climate and soil monitoring station (red circle in Fig. 1a) that has provided a record of climate and soil variables since 1998 (Westermann et al., 2009, 2011; Boike et al., 2003). Active layer thicknesses ranging from 1.6 to 2.0 m (Westermann et al., 2010) and mean annual ground temperatures at the bottom of the active layer between -2 and -3°C are reported from the field area (Roth and Boike, 2001; Boike et al., 2003).

The Juvvasshøe field site (1374–1497 m.a.s.l.) is located in the northeastern slope down from Juvvasshøe (1894 m.a.s.l.) (Fig. 1b), a small hill northwest of Galdhøpiggen (highest peak in Norway, 2469 m). The Galdhøpiggen massive is located at the divide

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between the more marine climate on the west coast and the continental climate in eastern parts of Norway. The terrain is open, with very sparse vegetation and a ground cover dominated by moraine deposits. The field area is well investigated in terms of ground thermal regime and permafrost distribution. A few hundred meters southwest of Juvvasshøe, mean annual ground temperatures measured in a 10 m deep borehole at 1851 m a.s.l. for the two periods 1982–1986 and 2008–2010 were -2.2°C (Ødegård et al., 1992) and -1.6°C (Farbrot et al., 2011), respectively. At the top of Juvvasshøe, data from a 129 m deep borehole (Sollid et al., 2000) show a very low geothermal gradient and indicates more than 300 m deep permafrost (Isaksen et al., 2001, 2007) and close relation of permafrost, ground surface and air temperatures (Isaksen et al., 2003). An altitudinal transect of shallow boreholes (10 m deep) was established in 2008 (Farbrot et al., 2011; Hipp et al., 2012). The lower limit of permafrost in this area is located at about 1450–1500 m (Farbrot et al., 2011; Hauck et al., 2004; Isaksen et al., 2002) and appears to have moved upslope the last decades (Isaksen et al., 2011).

The field site is located 400 m down the northeastern slope from Juvvasshøe, at a 15° steep slope facing northeast. The location is chosen to intersect the lower limit of permafrost (Hauck et al., 2004) and coincides with an established transect of 150 points with 14 yr of bottom temperature of snow (BTS) and continuous ground surface temperature measurements, maximum snow height measurements and DC resistivity profiles (Isaksen et al., 2011). For the measurement period, MAAT at the field site was -2.8°C , while annual precipitation estimated by seNorge.no (2013) was around 1200 mm yr^{-1} . According to this precipitation record, maximum snow heights of 1.5 to 2.0 m would be expected in the field area for the winter 2012/2013. However, the observed average snow height was less than 1.0 m, which is most likely caused by substantial wind drift of snow. During the winter of 2012/2013 the average wind speed at the meteorological station at Juvvasshøe was 6.5 ms^{-1} , with a maximum wind speed of 31.6 ms^{-1} . As a result, large parts of the area are bare-blown during the entire snow season, while some areas experience snow heights up to 6 m. The deep snow areas are covered by snow for approximately three quarters of the year.

The Finse field site (1293–1332 m.a.s.l.) is located in the upper part of a valley at the northern margin of the mountain plateau Hardangervidda. The 1 km × 1 km field area is located to the west of Vesle Hansbunut, 3.5 km southeast of Finse railway station (Fig. 1c). The area rises around 100 m above the valley bottom and features a rough and undulating topography. Finse is climatically situated in the transition zone between the maritime western coast and the more continental eastern parts of southern Norway. The general lower limit of permafrost in this area is estimated to 1550 m.a.s.l. (Etzelmüller et al., 2003). However, DC-resistivity soundings at 1450 m.a.s.l. (Etzelmüller et al., 1998) and observations of cold ice ($< 0^{\circ}\text{C}$) at the glacier front of Midtdalsbreen, an outlet glacier of Hardangerjøkulen (1400 m.a.s.l.) (Liestøl and Sollid, 1980; Andersen, 1996; Hagen, 1978; Etzelmüller et al., 1998), indicate permafrost at lower elevations at least at snow-free sites (Lilleøren et al., 2013). The MAAT at the field site was -2.8°C for 2012/2013. The measured precipitation at the official meteorological station (large black circle in Fig. 1c, left) located 1.5 km from the field area for the same period was 708 mm (eKlima, 2013), which was 24 % less than the normal. More than half of the measured precipitation came as snow, but a severe undercatch of up to 60 % must be expected also at Finse. The snow pack usually starts to build up in October, reaches a maximum of 3 m in the terrain (seNorge.no, 2013), and lasts until June or early July. The prevailing wind direction is from the west, and strong winds during winter season results in a pronounced spatial variability of the snow heights in the field area. The mean wind speed during the winter season 2012/2013 was 4.9 ms^{-1} , with maximum wind speeds of 28.7 ms^{-1} and mean wind speed during snow precipitation events of 5.5 ms^{-1} .

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GPR data were collected using an impulse radar system (RAMAC, Malå GeoScience) with a shielded antenna at the frequency of 800 MHz. A GPS receiver simultaneously recorded the positions together with the GPR. The GPR control unit, the antenna and the GPS were mounted on a sledge and pulled by snow mobile with a driving speed between 5 and 10 ms⁻¹. Radar traces were acquired at intervals of 0.25 s, resulting in intervals of 0.13–0.25 m. The surveys extend for 10 to 30 km at each site, and consist of in total more than 45 000 processed data points (Table 1). The tracks cover an area of 1 km × 1 km around the logger sites, and are sampled as a grid in Ny-Ålesund and Finse. At Juvvasshøe, the tracks were made along the terrain gradient instead of a grid due to the steep topography. Stake measurements taken along the radar tracks served as validation points for the processed radar data, and temperature and density profiles of the snow pack were measured at three sites in the field area at the day of the snow survey.

The processing of the GPR-data followed Dunse et al. (2009) and included static correction and frequency filtering, as well as removal of constant time-delay clutter. Since the snow pack was at freezing temperatures during all surveys, the wave speed of the radar signal in dry snow was derived from the permittivity and the speed of light in vacuum, with the permittivity obtained from snow density by an empirical relation (Kovacs et al., 1995). The snow height was determined from the two-way travel time of the reflection from the ground surface and the wave-speed.

Errors in the determination of snow heights mainly arise from the lateral variability of the snow density and from the picking accuracy of the radar reflection from the ground surface. The largest variability of snow density was found at Finse, with measured densities of 334 ± 40 kg m⁻³. This translates to wave speeds of 234 ± 6 m μs⁻¹ and an uncertainty of ±2 cm for a snow height of 1 m. On the other hand, inaccurate picking of the ground surface reflection may introduce errors in the range of 10–15 cm, so that it constitutes the largest source of uncertainty. For shallow snow packs, the strong vibrations of the GPR antenna can lead to a noisy signal and thus a lower accuracy.

4 Results

4.1 Observed distribution of ground surface temperatures

The observed average MAGST of all loggers at Ny-Ålesund, Juvvasshøe and Finse is -1.6°C , -0.5°C and 0.8°C , respectively. However, there is a large spread in MAGST of 5.0°C , 2.8°C and 4.5°C at the three sites (Fig. 2), with 98 %, 77 % and 30 % of the GST loggers featuring a MAGST below 0°C .

The largest variability in monthly mean temperatures is found during the winter months with a developed snow cover (January to April). At Finse, the variation is up to 7°C in March (Fig. 3). During the snow-free period (July to October), the observed spatial variability is very low, with most of the loggers being within 1°C variation at each site. Finse features a significantly larger variability than the two other sites during winter, coinciding with a larger variation in snow cover. The median at Finse is close to 0°C during the entire winter, showing that GST is decoupled from the air temperature in large parts of the field area.

The duration of the snow cover at each logger site was calculated from the difference in diurnal temperature amplitudes at the ground surface and in the air. The duration of snow varies between 100 to 280 days at all three field sites, with an average of 220, 180 and 210 days of snow cover at Ny-Ålesund, Juvvasshøe and Finse. A large variability of GST is observed during the melt-out period in June (Fig. 3) due to the pattern of snow-covered and snow-free areas at this time. The low variability during May is related to the ripening phase of the snow pack, with many of the loggers featuring temperatures close to zero.

4.2 Distribution of snow heights

The distributions of annual maximum snow heights are displayed in Fig. 4. The average snow heights at the GST-logger sites were 0.67 m, 0.50 m and 0.98 m at the

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Ny-Ålesund, Juvvasshøe and Finse, respectively (Fig. 4a–c). The total spread in snow heights at the corresponding sites were 1.9 m, 2.8 m and 4.2 m.

The average snow heights from the GPR snow surveys in the three study areas were 0.53 m, 0.93 m and 1.39 m, with a corresponding total spread of 3.5 m, 5.8 m and 6.0 m (Fig. 4d–f). The areal fractions with less than 20 cm of snow were 9 %, 28 % and 10 % at the three sites, while the fraction with snow heights of less than 1 m were 92 %, 66 % and 49 %.

4.3 Modeling small-scale variability of MAGST

The thawing/freezing degree days in the air calculated at each field site were 533/1904 for Ny-Ålesund, 838/1943 for Juvvasshøe and 874/1684 for Finse. The observed values of nF varied between 0.02 and 1.25 at the three sites, with 43 % (Ny-Ålesund), 67 % (Finse) and 75 % (Juvvasshøe) of the variation in nF explained by maximum snow height. The observed values of nT varied less at the three sites, with values in the range 0.4–1.5 in Ny-Ålesund, 0.6–1.3 at Juvvasshøe and 0.4–1.2 at Finse. Snow height can explain 36 % (Ny-Ålesund), 80 % (Juvvasshøe) and 39 % (Finse) of the variation in nT , since the variation of nT is strongly related to the length of snow free days. The parameters for the empirical relations between snow height and n -factors for each field site are given in Table 2.

CryoGRID 1 was run for the distribution of snow heights measured with GPR at the field sites (Fig. 4, lower). The distributions of modelled MAGST are displayed in Fig. 5, with total variations of 4.8 °C (Ny-Ålesund), 2.8 °C (Juvvasshøe) and 4.1 °C (Finse). The corresponding average values of modelled MAGST are –2.1 °C, 0.2 °C and 1.1 °C.

The model results (Fig. 5) are in good agreement with the observed distributions (Fig. 2). The model result for Ny-Ålesund agree best with observation, with an average MAGST 0.5 °C lower than measured. With 4.8 °C compared to the observed 5 °C, CryoGRID 1 can well reproduce the variation of MAGST, while the fraction of loggers with sub-zero MAGST is 99 % compared to the observed 98 %. The obtained distribution is slightly narrower, which can be explained by differences between the snow

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Ålesund, most of the observed snow heights are in the range 40 to 70 cm, with 90 % of the measurements being below 1 m. The insulating effect of the snow cover varies most in this range (Haeberli, 1973; Keller and Gubler, 1993; Luetschg et al., 2008), which is reflected in the large variability of GST despite of the smaller variability in snow heights.

At lower latitudes at Juvvasshøe and Finse, the summer season is relatively more important for the ground thermal regime than the winter. Thus, even with almost double variation of snow heights at Finse compared to Ny-Ålesund, there is less variation in MAGST. Furthermore, smaller fractions of the observed snow heights at these two sites are below 1 m (65 % and 50 % at Juvvasshøe and Finse). Since GST becomes disconnected from the air temperatures at snow heights above 1 m, the large variation from 1 to 6 m found at the two southern sites are of minor importance for the GST variation.

The size of the field areas in this study is chosen to exclude ground temperature variation related to factors varying on larger spatial scales, such as air temperature and precipitation patterns. Based on hundreds of BTS points taken in Jotunheimen, partly at Juvvasshøe, Isaksen et al. (2002) found that there is a high correlation to elevation at a larger scale (> 500–600 m), while the very small-scale variation below 20–30 m mainly is explained by the variation in the interaction of a set of climatic, surface and subsurface factors. Between 20–30 m and 500–600 m they found that the variation is explained by snow drift and variation in direct radiation. The results from the study presented in this paper confirm that variation in snow cover is the dominating factor of influencing spatial variability in GST at scales below 500–600 m.

A 10 yr data series of 8 ground surface temperature loggers distributed over the 500 m long transect at Juvvasshøe (partly used in this study) shows a small-scale variation in MAGST by ± 1.5 – 2.0 °C within a distance of 30–50 m and more than 3 °C within a distance of 100 m (Isaksen et al., 2011). While the total amount of snow varies from year to year, the relative variation in MAGST between the loggers is more or less the same for all years. Snow height measurements taken at the logger sites at maximum snow height show the same pattern of snow depths in every year. These findings

suggest that an area has a distinct snow distribution pattern, which may facilitate estimating the annual snow distributions using simple snow redistribution models (e.g. Winstral et al., 2002) and larger-scale precipitation fields.

5.2 Representation of sub-grid variability in permafrost models by including snow distribution functions

This study demonstrates that a statistical representation of sub-grid variability is feasible even in simple permafrost models such as CryoGRID 1 (Fig. 5 compared to Fig. 2). Regional permafrost models are forced with spatial fields of climate parameters, such as precipitation, air temperature and surface radiation. A main challenge in permafrost modeling is the pronounced small-scale variability of GSTs, which occur at spatial scales several orders of magnitude below the variation of climatic parameters. For the Scandinavian mountains, a spatial resolution of 1 km can sufficiently account for regional air temperature and precipitation patterns (Gisnås et al., 2013; Westermann et al., 2013; Eitzelmüller, 2013). At such resolution, the sub-grid variability in MAGST is mainly explained by snow cover variations induced by wind drift. This study suggests that the sub-grid distribution of ground temperatures can be statistically modeled if a distribution of snow depths is available. Such distributions can be inferred from snow redistribution models, e.g. Alpine-3D (Lehning et al., 2006), Snow-model (Liston and Sturm, 1998) or by applying the Winstral terrain parameterization (Winstral et al., 2002). A statistical representation of sub-grid variability is computationally much less demanding than a deterministic representation on a refined grid, so that application for large areas becomes feasible: for a 1 km grid cell, the model can be run for e.g. 10 different bins of snow depths instead of increasing the resolution to 10 m, with computation time increasing by a factor of ten instead of 10 000. More importantly, distributions of ground temperatures can be directly related to the traditional concept of permafrost zonation in continuous (> 90 % of an area), discontinuous (50–90 %), sporadic (10–50 %) and isolated (< 10 %). The transition from continuous to sporadic permafrost in a warming climate can thus be inherently described by such a sub-grid

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concept, while area-averaged formulations for large grid cells can only deliver a single ground temperature for each grid cell. Permafrost degradation can occur in a small part of a grid cell, although the permafrost model (or the closest borehole) suggests stable permafrost. Even if degradation occurs only over a small fraction of the area, it can trigger non-linear processes, such as thermokarst or thaw slumps, affecting the larger-scale surface fluxes of energy, water and carbon. The occurrence of positive MAGST in Ny-Ålesund for a small fraction of the GST loggers (Fig. 2) suggests the onset of degradation at localized sites, which would be obscured when only considering area-averages of MAGST.

The statistical representation of sub-grid variability can be extended to include other factors, e.g. exposition, land cover or ground thermal properties. However, this would require multi-dimensional distribution functions, which can describe the potentially complex correlations between statistically not independent parameters, e.g. between snow depths and surface soil moisture. It is thus important that such modeling efforts are strongly guided by field studies on the governing parameters for the ground thermal regime in a particular region. In the study areas, the variability of ground temperatures can to a large degree be described by the variability of a single parameter, snow depth. However, this is not necessarily true for other permafrost areas, for instance for more southerly permafrost areas, where exposition can be a strong additional control due to larger values of incoming radiation (Gruber et al., 2004; Gubler et al., 2011).

6 Conclusions

At three sites along a transect from Svalbard to southern Norway the distributions of ground surface temperatures are recorded within areas of 0.5 km² by in total 171 temperature loggers. With the data set, a statistical approach to include the small-scale variability of temperatures in a simple permafrost model could be validated.

From this study the following conclusions can be drawn:

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1. In high mountain and arctic areas dominated by strong winds, wind drift of snow creates a pattern of snow depths, resulting in a pronounced small-scale variability of mean annual ground surface temperatures.

2. The small-scale variability of ground surface temperatures is most pronounced during the winter months with a developed snow cover and during snow melt, while the variability is comparatively small during summer and early winter.

3. If the spatial distribution of snow heights is known, a statistical representation of the spatial variability of ground surface temperatures is feasible even in a simple permafrost model.

The study exemplifies the necessity of representing the sub-grid variability of ground temperatures in larger-scale model approaches. A statistical representation of snow cover and ground temperatures within model grid cells can facilitate application over large areas.

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Table 1. Site and climate characteristics for the measurement period. MAAT: mean annual air temperature in the field area. MHS: average snow height in the terrain at snow maximum; MWWS: mean winter (December–March) wind speed; MaxWWS: maximum winter (December–March) wind speed. Wind speed at Juvvasshøe is measured at the meteorological station at the top of Juvvasshøe (1894 m a.s.l.).

	Ny-Ålesund	Juvvasshøe	Finse
Latitude	78.6° N	61.3° N	60.2° N
Altitude	10–50 m a.s.l.	1374–1497 m a.s.l.	1293–1332 m a.s.l.
MAAT °C	−3.8 °C	−2.8 °C	−2.2 °C
MHS	0.6 m	0.7 m	1.6 m
MWWS	3.9 ms ^{−1}	6.5 ms ^{−1}	4.7 ms ^{−1}
Max WWS	18.7 ms ^{−1}	31.6 ms ^{−1}	28.7 ms ^{−1}
No. GST loggers	104	26	41
Measurement period	9 Sep 2012– 9 Sep 2013	27 Jul 2012– 14 Jul 2013	25 Sep 2012– 25 Sep 2013
Temporal resolution	HOBO: 4 h iButtons: 4 h	UTL-3: 2 h iButtons: 4 h	Hobo: 2 h iButtons: 4 h
Snow survey	23 Apr 2013	14 Mar 2013	20 Mar 2013
Snow density	305 kg m ^{−3}	340 kg m ^{−3}	338 kg m ^{−3}
Avg velocity	238 m μs ^{−1}	232 m μs ^{−1}	233 m μs ^{−1}
Length of GPR survey	28 km	11 km	16 km
Sample points	23 330	7957	14 665

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Table 2. Parameters of transfer functions representing the surface offset in the model approach. MHS is the maximum height of snow.

Site	$nF(a \cdot \log(\text{MHS}) + b)$			$nT(a \cdot \text{MHS} + b)$		
	a	b	p value	a	b	p value
Ny-Ålesund	-0.3	0.39	3.44×10^{-13}	-0.28	1.3	1.55×10^{-4}
Juvvasshøe	-0.17	0.29	5.80×10^{-3}	-0.2	1.1	8.06×10^{-8}
Finse	-0.19	0.19	2.70×10^{-11}	-0.089	1.0	6.25×10^{-4}

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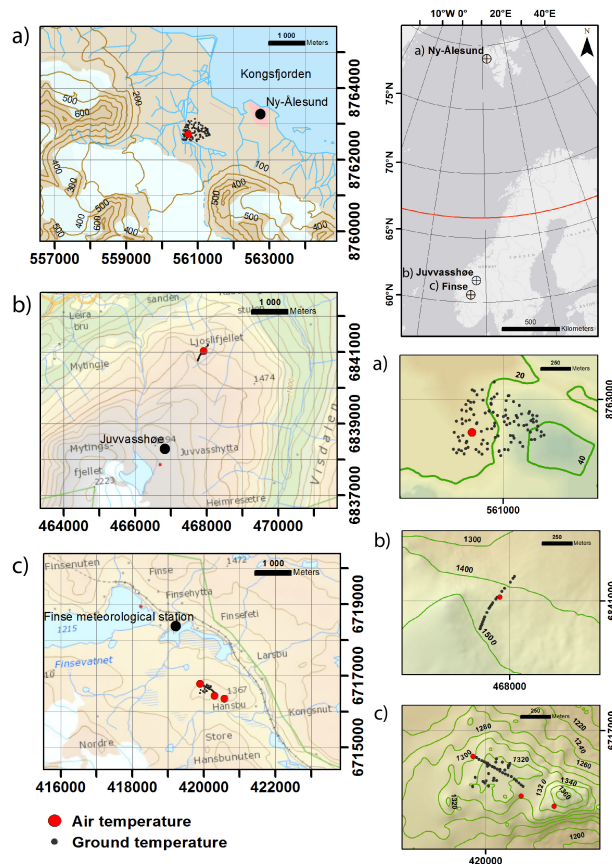


Fig. 1. Study areas located along the gradient from continuous to sporadic permafrost in northern Europe: **(a)** Ny-Ålesund, Svalbard; **(b)** Juvvasshøe, Jotunheimen, southern Norway; **(c)** Finse, Hardangervidda, southern Norway. Right column: distribution pattern of ground surface temperature (GST). Black dots: location of GST loggers, red dots: air temperature at 2 m height.

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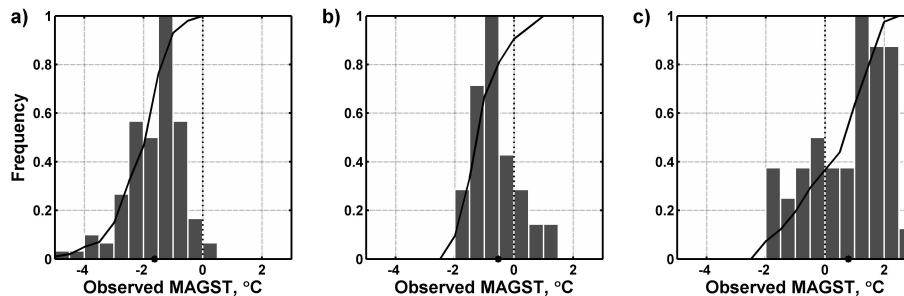


Fig. 2. Histograms of measured mean annual ground surface temperatures (MAGST) at the logger sites in **(a)** Ny-Ålesund, **(b)** Juvvasshøe and **(c)** Finse. The cumulative curve is given as a solid, black line, and average MAGST is marked with a black dot at the x axis. The fraction of samples below 0°C is given by the intersection of the cumulative curve and the 0°C line (stippled line).

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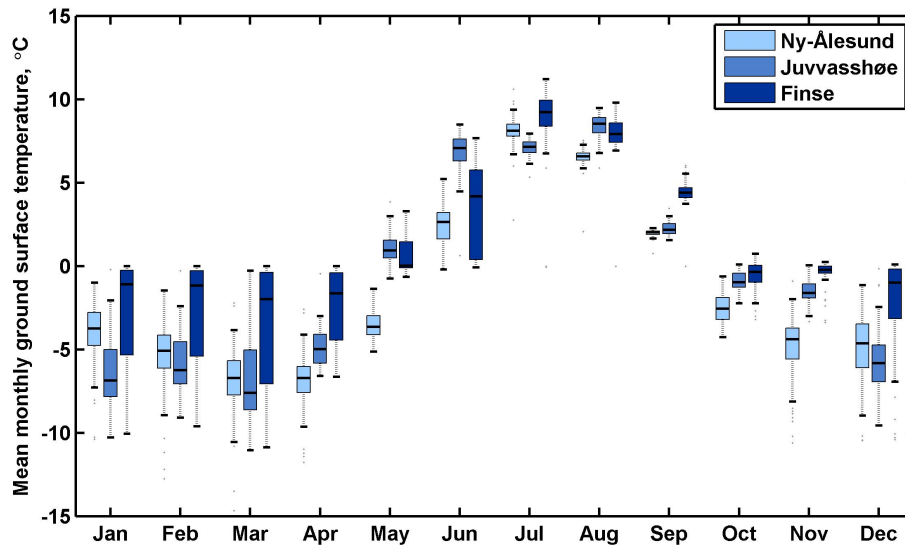


Fig. 3. Variation of ground surface temperatures between the loggers at the three field sites. The bars indicate the 25th and 75th percentiles, while the whiskers indicate minimum and maximum values. The median is given as a horizontal line in each bar.

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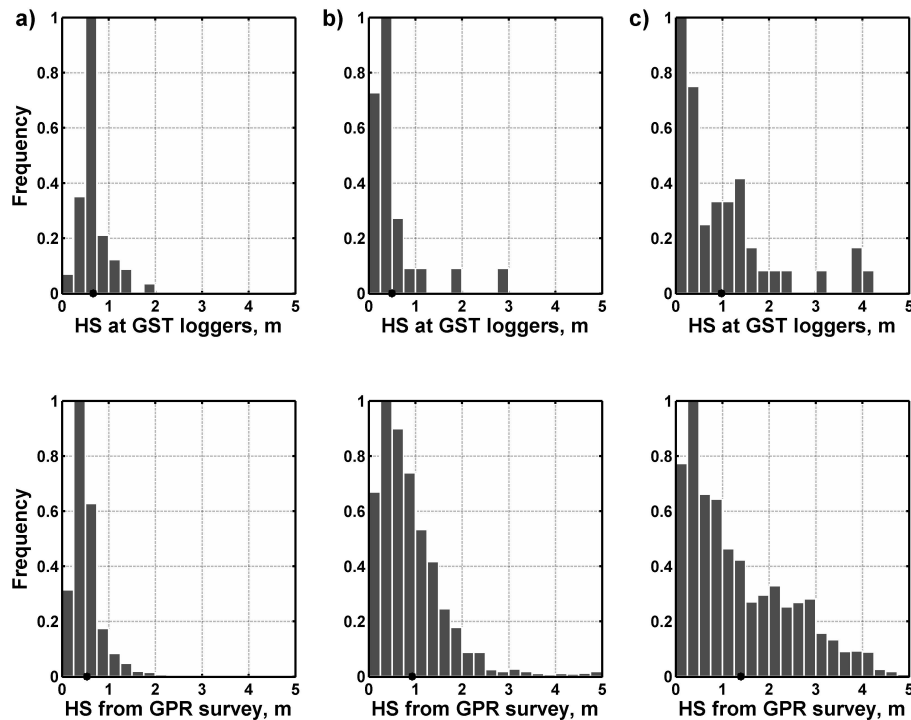


Fig. 4. Distributions of snow heights (HS) at the respective field sites **(a)** Ny-Ålesund, **(b)** Juvvasshøe and **(c)** Finse. Upper row: snow heights measured manually at each logger site; Lower row: snow heights from GPR snow surveys over a 1 km × 1 km area around the loggers. Black dot: average snow height.

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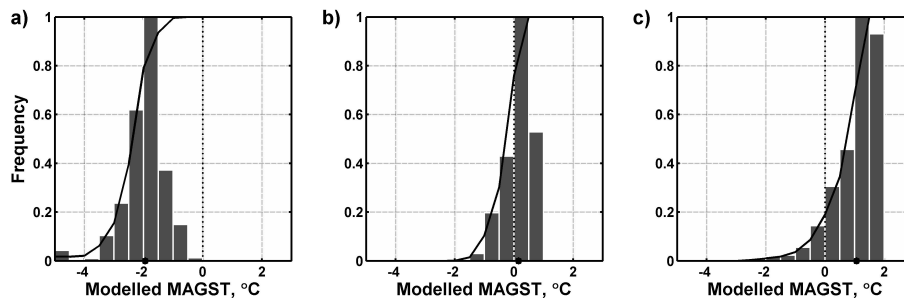


Fig. 5. Distribution of modelled MAGST at **(a)** Ny-Ålesund, **(b)** Juvvasshøe and **(c)** Finse. The cumulative curve is given as a solid, black line, and average MAGST is marked with a black dot at the x axis.