Permafrost map for Norway, Sweden and Finland

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Abstract

The thermal regime of permafrost is sensitive to changes in the climate system. A common, research-based understanding of the permafrost distribution at a sufficient spatial resolution is important to meet scientific, educational and societal demands. We present a new permafrost map for Norway, Sweden and Finland providing a more detailed and updated description of the permafrost distribution in this area. The CryoGRID1 model is implemented at 1 km^2 resolution, forced by a new operationally gridded dataset of daily air temperature and snow cover for Finland, Norway and Sweden. 100 model realizations are run for each grid cell, based on statistical snow distributions, allowing for the representation of sub-grid variability of ground temperatures. The new map indicates a total permafrost area (palsas excluded) of 23 400 km^2 in equilibrium with the average 1981 – 2010 climate, corresponding to 2.2 % of the total land area. About 56 % of the area is within Norway, 35 % in Sweden, and 9 % in Finland. The model results are thoroughly evaluated, both quantitatively and qualitatively, as a collaboration project including permafrost experts in the three countries. Observed ground temperatures from 25 boreholes are within ±2 °C from the average modelled grid cell ground

temperature, and all are within the range of the modelled ground temperature for the corresponding grid cell. The model results are also evaluated qualitatively by the representatives conducting field investigations on permafrost at several sites within the three countries, and are shown to reproduce observed lower altitudinal limits of permafrost and permafrost distributions mapped by the corresponding detailed field investigations.

1. Introduction

Permafrost has gained wide attention because of its sensitivity to climate change. With increasing surface temperatures, permafrost may degrade, with associated consequences like increased release of carbon to the atmosphere (Schuur *et al.*, 2008; Elberling *et al.*, 2010) or increased slope instability in mountain areas (Krautblatter *et al.*, 2013). Knowledge about the spatial distribution and temperatures of permafrost is crucial for several reasons, in particular as a baseline for validation of global and regional climate models and associated land surface models, and as a tool for planning of human activities.

Permafrost, defined as ground remaining at or below 0 °C for two or more consecutive years (French 2007), is not necessarily visible on the ground surface as it is solely a thermally defined phenomenon, corresponding to the subsurface regime below the active layer. It is therefore important to establish a common, research-based understanding of the permafrost distribution at a sufficiently high spatial resolution to meet scientific and societal demands. The first Northern Hemisphere permafrost map was the "*International Permafrost Association Circum-Arctic Map of Permafrost and Ground Ice Conditions*" (hereafter the IPA map) by Brown *et al.* (1997), with a scale of 1:10,000,000. This map was produced by gathering information from researchers working with permafrost in all different permafrost areas. Today the map has been widely used as a baseline for validating modelled permafrost in the northern hemisphere based on both global and regional Circulation models. It provides information of the spatial distribution of permafrost along with potential ground ice content, and is mainly based on experience and field observations. Because of the coarse map scale it has only limited application in regional areas, such as the Scandinavian Peninsula. In this region, numerous field-based studies have been conducted since the 1980s, including the establishment of around 30 new permafrost boreholes during the International Polar Year 2007-2009 (IPY) (Isaksen *et al.*, 2001; Christiansen *et al.*, 2010; Farbrot *et al.*, 2011; Sannel *et al.*, 2015). The new knowledge draws a more differentiated image of the permafrost distribution, justifying the production of a separate and more detailed regional permafrost map of this region.

A main objective for a new baseline map is to provide an improved regional-scale description of the current state of permafrost with greater accuracy and detail. The map should serve as a useful and reliable tool for both the cryosphere research community and practitioners interested in the state of permafrost in the area. A new map should therefore also to the largest extent possible be based on observations. For permafrost, this includes primarily the forcing climate parameters such as air temperature and snow conditions, along with modulating parameters such as the type of surface cover and soil thermal properties. To achieve the desired spatial resolution of ground temperatures, climate information must be transferred to soil temperatures and thus permafrost distribution. A model used for this purpose should be robust enough to give trustworthy results, and simple enough to have relevant input parameters over a larger area, preferably through field observations. In a second step, the resulting permafrost distribution must be validated with field information in several different areas, to reduce the spatial bias.

Here, we present a new permafrost map for the Norway, Sweden and Finland providing a detailed description of the permafrost distribution in this area, following the objectives mentioned above. The map is obtained by applying a simple numerical equilibrium model, the

CryoGRID1 (Farbrot *et al.*, 2011; Gisnås *et al.*, 2013), forced with 1 km^2 gridded data on daily air temperature and snow cover for the period 1981 – 2010 based on the recently developed Nordic Gridded Climate Data set (NGCD, http://blog.fmi.fi/nordmet/node/100). The CryoGRID1 model uses a limited set of parameters, mainly to describe the temperature offsets between air, ground surface and ground temperatures. These parameters were refined for snow and mires (including palsas) based on extensive field measurements. The model results are qualitatively evaluated by national permafrost experts with active field research sites, and compared to the available ground and ground surface temperature observations, "Bottom temperature of snow" (BTS, Haeberli (1973))-surveys and geomorphological maps showing permafrost landforms such as palsas, intact rock glaciers and stable ice-cored moraines. Finally, changes in the modelled permafrost distribution over the period 1981-2010 are evaluated.

2. Setting

Norway, Sweden and the north-western part of Finland form together the Scandinavian Peninsula. In the following we also include the remaining parts of Finland under the Scandinavian Peninsula, giving in a total land area of the peninsula of 1.111.890 km^2. The geology of the peninsula consists of a stable large crust of very old metamorphic rock (c. 2500-3100 Ma old). The bedrock in most of northern and central Sweden, together with the south-western parts of Finland, was formed during the Svecofennian orogeny (1750-1900 Ma ago). The Scandinavian Caledonides or the Scandes, stretching through most of Norway and adjacent parts of Sweden, form the highest mountains of Scandinavia, with peaks up to 2469 m a.s.l. (Galdhøpiggen, Norway). The mountain chain consists of metasedimentary and metavolcanic rocks, deposited in the predecessor of the present-day Atlantic Ocean c. 700 to 400 Ma ago, together with slices of older basement. These rocks were thrust several 100 km eastwards over the edge of the Fennoscandian Shield when it collided with North America and Greenland during the Caledonian orogeny c. 400 Ma ago. During the opening of the Atlantic Ocean in the Tertiary, the margin of Scandinavia was tilted, with the highest land heave in the west (see e.g. summary by Gabrielsen *et al.* (2010)).

The present topography of Scandinavia is a result of subsequent modulation by multiple glaciations the last 3 Ma years, while the sediment cover over the bedrock is mostly related to the last one or two major glaciations (Gabrielsen *et al.*, 2010). This has resulted in a large variety of landscapes on the peninsula. Pre-existing mountain river systems in the west were linearly carved by the glaciers, producing the present fjord landscape. Remains of the paleic surfaces were preserved both between the fjord systems and towards the east, indicating cold-based and non-erosive conditions during at least the latter glaciations. In some areas, local glaciations have dominated over longer time periods, leaving alpine relief forms.

The superficial deposits in Scandinavia are governed by the architecture and deglaciation pattern of the Pleistocene ice sheets. Only c. 43 % of the land area is today covered by till(Ramberg *et al.*, 2006). In high mountain areas, exposed bedrock and only thin covers of till or regolith dominate, while valleys are often filled with glacio-lacustrine, glacio-fluvial and fluvial sediments. Over certain elevation limits, depending on latitude and distance from the coast, mountain slopes and plateaus are covered by coarse block fields. These block fields can be several meters thick, with coarse boulders overlying finer sediments. The origin of these block fields is disputed, but most likely they are remnants from a combination of pre-glacial weathering along with Pleistocene frost processes occurring above glacial trimlines (Ballantyne, 2010). During the Holocene, a climatic shift towards warmer and wetter conditions favoured the accumulation of organic material in wetlands. This material covers wide areas in central and especially northern Scandinavia. Both the block fields and the organic material play crucial roles for the thermal regime and distribution of permafrost.

The present (1981 – 2010) climate in Scandinavia is highly variable, with mean annual air temperatures (*MAAT*) of up to 9 °C in the maritime areas along the coast and in the southernmost parts (Figure 1). The highest parts of the Scandes (> 2000 m) feature *MAAT* below -5 °C. In the northernmost parts of Finnmark *MAATs* are often below 0 °C all the way down to sea level. The Scandes represents a significant orographic barrier for the prevailing westerly winds from the Atlantic Ocean, creating a strong east-west gradient in the annual precipitation pattern. This results in average maximum snow depths of 2 to 6 meters in the western parts of the mountains (average over 1981 – 2010), while on the eastern side maximum snow depths are normally 2 meters or less. Further east of the Scandes the maximum snow depths averaged over 1 km^2 are generally below 1 meter, according to the snow depth data based on the NGCD dataset (Figure 1).

3. Permafrost in Scandinavia

The majority of the permafrost in both Norway and Sweden is found in mountainous settings (e.g. Etzelmüller *et al.*, 2003; Gisnås *et al.*, 2013). However, in the northern parts of Scandinavia, much of the permafrost is located in mires, often producing palsa landforms and peat plateaus, signifying sporadic permafrost distribution (e.g. King and Seppälä, 1987; Sollid and Sorbel,1998; Johansson *et al*., 2006; Borge *et al.*, 2016). While permafrost was suggested present in the Scandinavian mountains already in the beginning of the 19th century (Reusch, 1902), its wide distribution in the mountains was not recognized until extensive studies were carried out by King (King, 1982; 1986; King and Seppälä, 1987; King and Åkerman, 1993) in both southern and northern Scandinavia. Before these fundamental studies, permafrost was mainly thought to be related to palsa mires (Reusch, 1902), which are wide-spread from sea level and up to 1000 meters north of approximately 68 °N in both Norway, Finland and Sweden (Figure 2).

Since the late 1980s, permafrost studies have followed different directions in Norway, Sweden and Finland. In Sweden and Finland much focus was directed towards sporadic permafrost, mostly related to palsas (e.g. Seppälä, 1997). In Abisko (northern Sweden), Åkerman and Johansson (2008) established a monitoring station for active layer thickness, which has been in operation since 1978. In Norway, a first 10 m borehole in permafrost was drilled at Juvvasshøe (southern Norway) in 1982 (Ødegård *et al.*, 1992). This was the start of a more systematic mapping of mountain permafrost in southern Norway (Ødegård *et al.*, 1996; Isaksen *et al.*, 2002), to a large degree based on BTS-surveys (e.g. Ridefelt *et al.*, 2008). In 1999 and 2000 the EU-funded PACE project (Harris *et al.*, 2001) funded two deep boreholes in Scandinavia (at Juvvasshøe and at Tarfalaryggen), described in Isaksen *et al.* (2001) and Sollid *et al.* (2000). These boreholes boosted the mountain permafrost research in Scandinavia, and a borehole monitoring network was established in southern Norway (Sollid *et al.*, 2003; Farbrot *et al.*, 2011). During the IPY monitoring networks were also built up in northern Norway, northern Sweden and Finland, along with Svalbard (Christiansen *et al.*, 2010).

From these studies, an improved understanding of the permafrost distribution in Scandinavia is obtained, as outlined in Figure 2. The southernmost observation of permafrost in Scandinavia is reported from Gaustadtoppen in southern Norway (59.9 °N), with frozen bedrock down to 1500 m a.s.l. (Etzelmüller *et al.*, 2003). The lower altitudinal limit of permafrost (LALP) is clearly lower in the eastern parts of southern Norway (transition zone at 900 – 1100 m a.s.l. (Heggem *et al.*, 2005; Juliussen and Humlum, 2007), than in the central and western parts (transition zone at 1300 -1550 m a.s.l.; Ødegård *et al.*, 1992; Isaksen *et al.*, 2002; Etzelmüller *et al.*, 2003; Sollid *et al.*, 2003). Both snow conditions and surface material favour a lower permafrost limit in the eastern parts of southern Norway (Farbrot *et al.*, 2011).

In northern Norway, there is a similar gradient of decreasing LALP with the increasingly continental climate away from the coast. In coastal areas permafrost exists above c. 1250 m a.s.l., while in Kilpisjärvi and Abisko, located on the eastern side of the Scandes, permafrost exists down to c. 800 – 850 m a.s.l. (Ridefelt *et al.*, 2008). In inner parts of Finnmark (Farbrot *et al.*, 2013) and northern Finland (Christiansen *et al.*, 2010) permafrost is widespread above treeline (about 400 m a.s.l.) and common in extensive mires with relatively thick (> 70 cm) peat deposits (Seppälä, 1988). Permafrost is also present at local sites with coarse surface material, favouring a cold anomaly, resulting in active rock glaciers down to sea level, e.g. at Nordkinnhalvøya (Lilleøren and Etzelmüller, 2011; Lilleøren, 2016).

4. Model description

4.1 Implementation of CryoGRID1

The equilibrium model CryoGRID1, earlier implemented for Norway (Farbrot *et al.*, 2013; Gisnås *et al.*, 2013), is a TTOP-approach (Smith and Riseborough, 1996) providing an estimate for the temperature at the top of permafrost or at the bottom of the seasonal frost layer (*TTOP*) based on annual freezing (FDD_a) and thawing (TDD_a) degree days in the air:

$$TTOP = \begin{cases} \frac{(TDD_a * nT * r_k + FDD_a * nF)}{P} & \text{for } K_t TDD_s + K_f FDD_s \leq \\ \frac{\left(TDD_a * nT + \frac{1}{r_k} * FDD_a * nF\right)}{P} & \text{for } K_t TDD_s + K_f FDD_s \geq \end{cases} \quad (1)$$

Here, P is the total time period for which FDD_a and TDD_a are integrated over (i.e. the number of days in one year), while r_k is the ratio of ground thermal conductivities in thawed and frozen states. The factors nT and nF are empirical transfer-functions for the surface offset, defined as the offset between air and ground surface temperatures. The transfer-functions include a variety of heat flow attenuation processes in one single variable, such as vegetation, snow cover, soil moisture and topography. The winter *nF*-factor relates the freezing degree days at the ground surface to the air and thus accounts for the effect of the winter snow cover. Likewise the *nT*-factor relates the thawing degree days at the surface (subscript s) to the air (subscript a) and accounts for the surface cover type:

$$FDD_s = nF * FDD_a \quad \text{and} \quad TDD_s = nT * TDD_a \qquad (2)$$

In non-vegetated areas the snow cover and the ground surface temperatures can be highly variable over short distances due to wind drifting of snow (Gisnås et al. 2014). To account for this variation we assume that the distribution of maximum snow depths within a grid cell follows a gamma distribution, following Gisnås *et al.* (2015). The distribution is defined by the average maximum snow depth (μ) of the grid cell and here a constant coefficient of variation (*CV*) of 0.6 is adopted. The probability density function over a range of snow depths (*SD)* is then given as (Skaugen *et al.*, 2004):

$$f(SD;\ \alpha, \beta) = \frac{1}{\beta^{\alpha}\Gamma(\alpha)} SD^{\alpha-1} e^{-\frac{SD}{\beta}} \qquad (3)$$

with shape parameter $\alpha = CV^{-2}$ and rate parameter $\beta = \mu * CV^{2}$. We run 100 model realizations based on snow depths selected from the probability distribution *f(SD)*. Corresponding *n*-factors are computed for each realization based on Eq. 4 and 5 below, giving a distribution of *TTOP* and mean annual ground surface temperatures *(MAGST)* for each grid cell. Based on the *TTOP* distribution, the percentage of permafrost, defined as sub-zero *TTOP,* is derived for each grid cell. This percentage is used to classify each grid cell into one of the four distinct permafrost zones reflecting the spatial coverage of permafrost (Brown et al. 1997); no and isolated permafrost (0 – 10 %), sporadic permafrost (10 – 50 %), discontinuous permafrost (50 – 90 %) and continuous permafrost (90 – 100 %). Model set up is described in the flow chart in Figure 3.

4.2 Parameterization of n-factors

The *n*-factors are classified for the following surface cover classes: (1) coniferous forest, (2) broad-leaved forest, transitional woodland and shrub, (3) mire and peat, (4) bedrock and sparsely vegetated area and (5) block field. Variation in observed *n*-factors for forest and shrub is relatively small, with *nT* typically in the range 0.85 – 1.1, and *nF* in the range 0.3 – 0.5 (Gisnås *et al.*, 2013). For mires and peat land the *n*-factors in the main model run are parameterized for general mire conditions with a uniform snow cover. One year of data from four temperature loggers over mires with a developed snow cover at Iskoras in Finnmark showed *nT* and *nF* values in the range 1.1 to 1.4 and 0.15 to 0.35, respectively. Based on these data *nT* and *nF* at mires without snow drift was given the average values of 1.25 and 0.26, respectively (Table 1). However, as a palsa grows, the surface offset changes significantly due to the reduced snow cover and also potentially because of changes in vegetation cover at the top of the palsa. In order to represent the situation at snow free sites in mires and peat land, such as on palsas, a second model run is made with *n*-factors adjusted for "snow free mire conditions". Several years of data from temperature loggers located at the very of top of palsas in Tavvavuoma (8 yrs), northern Sweden (Sannel *et al.*, 2015), Vaisjeaggi (4 yrs) and Kilpisjärvi (2 yrs) in northern Finland, and Iskoras (1 yr), northern Norway, show mean *nT* and *nF* values in the range 0.7 to 1.2, and 0.35 to 0.85, respectively. The average values of *nT* and *nF* are 1.2 and 0.6, respectively, which is used for this condition in model (Table 1) .

Observed variations in *nT* and *nF* within the open non-vegetated areas are comparably large, with values typically in the range 0.4 – 1.2 for *nT* and 0.0 – 1.0 for *nF*. The large range is related to the high impact and high spatial variability of snow depths (Gisnås *et al.*, 2014). While *nF* accounts for the insulation from snow due to low thermal conductivity, *nT* indirectly compensates for the shorter season of thawing degree days at the ground surface in areas with a thick snow cover. Relationships between *n*-factors for open areas and maximum snow depths (*SD*) are established based on air and ground temperature observations, together with snow depth observations at the end of accumulation season at nearly 100 sites in southern Norway, presented in Gisnås *et al.* (2015) as:

$$nF = -0.16 * \ln(SD) + 0.22 \tag{4}$$

$$nT = -0.14 * SD + 1.1 \tag{5}$$

The thermal conductivities of the different surficial materials, excluding mires, are derived from 28 000-point measurements of petrophysical data such as bedrock density and thermal conductivity provided by the Geological Survey of Norway (NGU) (Olesen et al., 2010), along with thermal properties used in other models (GIPL 1.0 – Geophysical Institute Permafrost Laboratory) (Table 2). A comprehensive overview is given in Gisnås *et al.* (2013). For mires, r_k is increased to 0.75, based on a model fit over borehole temperatures measured in peat plateaus in Tavvavuoma (Sannel *et al.*, 2015) and Vaisjeaggi.

5. Data and field observations

5.1 Climate forcing data

The Nordic Gridded Climate Data set (NGCD) of daily gridded air temperature and precipitation data at 1 km^2 resolution for the period 1981 – 2010 is provided by the Norwegian Meteorological Institute. NGCD is developed within the Nordic Framework for Climate Services (http://blog.fmi.fi/nordmet/node/100), based on the Norwegian Climate

Database in addition to the European Climate Assessment Dataset (ECA&D, Klein Tank *et al.*, 2002). Original non-homogenized time series were used. The number of stations used for the interpolation varies with time due to data availability. In more than 80 % of the time steps the interpolation is based on data from more than1100 precipitation stations and 371 temperature stations. These are distributed over the three countries with approximately 25% of the stations in each of Norway and Finland and 50% in Sweden. The station distribution is rather stable in the time period considered. In NGCD, the operational choices we've made were to use all the available observations for each timestep and to keep the statistical interpolation settings fixed in time such that its filtering and smoothing properties are also constant in time. As a result, any significant variation in time of the NGCD statistics should be attributed to an actual modification in the underlying climatological probability density function. However, as pointed out in Masson and Frei (2016), the variations in the station network have an impact in the estimation of long-term trends from observational gridded datasets and further investigations are needed to accurately assess this impact on the NGCD.

The interpolation of the NGCD gridded data set is based on two original implementations of the well-established statistical interpolation method called *Optimal Interpolation* (OI; Gandin (1965)). The goal of OI is to provide the best (i.e. minimum error variance for the final prediction), linear, unbiased estimate of the unknown meteorological field. Two distinct OI-based schemes are developed for temperature and precipitation because of the different statistical properties of the two meteorological variables. Nonetheless, for both variables the spatial interpolation relies on the scale-separation concept, which is based on the idea that the final predicted value is a combination of large scale and local scale effects. As a first step, we aim at estimating the contribution of atmospheric processes operating on scales greater than the local station density (i.e. *large* scale), thus influencing dozens of stations. Then we include the effects of atmospheric processes acting at the *local* scale, as simultaneously observed by a few stations over an area significantly smaller than the one used in the previous step. The small-scale component of an observed value that the spatial interpolation scheme is unable to resolve constitutes the so-called representativity error (Lussana *et al.*, 2010), which is part of the observation error in OI. For example, very local responses to mesoscale dynamics may affect just one or very few observations, thus determining large uncertainties in the final prediction.

An empirical Bayesian approach has been adopted for the implementation of the spatial interpolation scheme, where a priori information on the grid is combined with point observations. The a priori information is estimated from the observations and serves as a representation on the *large* scale. Given the actual station distribution, it is expected that the NGCD would properly resolve atmospheric processes from the synoptic down to the meso-beta scale (Orlanski, 1975; Thunis and Bornstein, 1996).

The first step for spatial interpolation for the daily mean temperature is the estimation of a large-scale temperature field, both on the grid and at station locations, by means of a non-linear de-trending procedure. Similar procedures are described in Uboldi *et al.* (2008) and Frei (2014). The large-scale field information can represent several regional vertical temperature profiles. This allows for both ground-based inversions as well as temporal changes of the vertical profiles at every time step, depending on the actual atmospheric conditions. In the second step, the large-scale background field is modified on a local basis by a few neighbouring stations employing an OI as described in Uboldi *et al.* (2008). In this way the best linear unbiased estimate of the unknown true temperature state is achieved. The procedure used to set the de-correlation scale values is based on a trade-off between the necessity to incorporate the added value of the local scale information into as many large-scale (background field) grid points as possible, and the need to keep a substantial distinction between large and local scale effects. Several tests were conducted with different de-correlation scale values in order to optimize the leave-one-out cross-validation score (Uboldi *et al*., 2008). Based on these test the values of 60 km and 600 m were found to be optimal for the horizontal and vertical directions, respectively. To deal with the presence of gross measurement errors in the temperature observations, a spatial consistency test is included in the statistical interpolation, as described in Lussana *et al.* (2010).

In the case of daily accumulated precipitation, the spatial interpolation begins with the identification on the grid of simultaneous observed areas of precipitation (i.e. precipitation events), followed by the statistical interpolation on each area considered independent from the others. The statistical properties of the field are allowed to change between different observed areas of precipitation. Eventually, the analysis field is a composition of several precipitation events, which are considered individually. In each individual area of observed precipitation, the statistical interpolation is based on a multi-scale separation concept through iteration from the large-scale (up to the synoptic scale, depending on the event extension) down to the finer scale, predefined as 10 km. In each iteration the statistical interpolation scheme is based on OI, and the parameters for the spatial covariance are optimized independently for each event by minimizing the deviations between leave-one-out cross-validated analysis and observed values.

The snow depth data was produced from the air temperature and precipitation data by employing the *SeNorge* snow model v.1.1.1., described in Saloranta (2012). Precipitation at temperatures below 0.5 °C is considered as snow, and snow water equivalents are calculated directly from precipitation and temperature data. Snow depths are derived from snow water equivalents using an algorithm that takes into account snow accumulation, temperature during snowfall, compaction and snow melt. Average mean annual air temperature and maximum snow depths over the 30-year period 1981-2010 are shown in Figure 1.

5.2 Surface cover and soil data

n-factors were assigned to groups of surface cover classes in the newest CORINE dataset, CLC2012 (Aune-Lundberg and Strand, 2010). Sediment and bedrock maps are provided by the Geological Surveys of Finland, Sweden and Norway. The maps are unified into one classification for r_k, following CryoGRID1 (Table 2).

5.3 Field observations used for evaluation

The modelled *TTOP* results are evaluated against 25 boreholes shown as blue and yellow dots in Figure 2 (Isaksen *et al.*, 2001; Christiansen *et al.*, 2010; Farbrot *et al.*, 2011), all incorporated in the GTN-P database (Biskaborn *et al.*, 2015). Modelled *MAGST* is evaluated with an extensive dataset of temperatures from more than 100 ground surface temperature (GST)-loggers distributed over Norway (Figure 2). For evaluation with temperature data the model is run for the hydrological year corresponding to the ground or ground surface temperature observations.

The overall modelled distribution of permafrost is evaluated towards maps of palsa distribution derived from aerial photographs from Norway (Borge *et al.*, 2016), Sweden (Backe, 2014) and Finland (Metsähallitus, 2002) (red dots in Figure 2). The Swedish maps are in grid squares of 100 m x 100 m, where the percentage of palsas and water related to palsas were specified. Furthermore, the modelled permafrost zonation is evaluated against permafrost probability maps derived from BTS-surveys at Kilpisjärvi (Majava, 2014), Abisko (Ridefelt *et al.*, 2008), Dovrefjell and Juvvasshøe (Isaksen *et al.*, 2002), and Elgåhogna and Sølen (Heggem *et al.*, 2005) (outlined in purple in Figure 2).

6. Results and discussion

6.1 Modelled permafrost distribution in the Scandinavian Peninsula

The model results indicate a permafrost area in Scandinavia of c. 23 400 km^2 in equilibrium with the 1981-2010 climate, when snow free mires are excluded (Figure 4 and Figure S1). The areas modelled to have permafrost are exclusively in non-forested areas, of which 21 % of the total distribution is found to be in areas classified as block fields. 56 % is located in Norway, 35 % in Sweden, and 9 % in Finland. In total, the area with a grid cell percentage of permafrost above 10 % is c. 62 600 km^2, whereof 2 % of the area is classified as continuous permafrost (> 90 %), 20 % discontinuous (50 – 90 %) and 78 % is sporadic (10 – 50 %).

In the “snow free mire” model run c. 5570 km^2 are modelled as potential permafrost. Norway, Finland and Sweden holds 51 %, 38 % and 11 % of these areas, respectively. In reality, permafrost can be found only in sites with favourable local conditions for permafrost development (e.g. thin snow cover, >70 cm peat deposits and no regular flooding) (Seppälä, 1988; Luoto and Seppälä, 2002; Seppälä, 2011).

The modelled gradient in LALP with distance from the coast is in accordance with earlier observations. In southern Norway the modelled distribution of discontinuous and sporadic LALP decreases from 1750 m a.s.l. and 1450 m a.s.l. in the west to 1350 m a.s.l. and 1050 m a.s.l. in the east, respectively (Figure 5, left). In northern Norway it decreases from 1400 m a.s.l. and 1200 m a.s.l. down to 350 m a.s.l. and 150 m a.s.l. in the eastern parts, respectively (Figure 5, left). The gradient is very low east of the Scandes in the north, where the LALP for discontinuous permafrost stabilizes at around 400 m a.s.l. through Finnmark and northern Finland. This elevation corresponds to the upper limit of birch forest in this area. The model results indicate occurrences of sporadic permafrost all the way out to the coast in northern

Norway, while in southern Norway there is no permafrost modelled closer than 150 km from the coast line (Figure 5, left).

There is also a strong latitudinal gradient in the LALP. Sporadic permafrost is present as far south as 60 °N, here above 1500 m a.s.l. The LALP for discontinuous permafrost is 1650 at 61°N. From 64 °N to 70 °N the LALP decreases gradually down to 150 m a.s.l. for sporadic and 350 m a.s.l. for discontinuous permafrost, respectively. The LALP for sporadic permafrost increases again in the northernmost maritime areas of northern Norway.

The extent of the modelled distribution of permafrost follows the general distribution of permafrost in the IPA map (Brown *et al.*, 1997). However, the correspondence is higher in the northern than the southern parts of the Scandinavian Peninsula (Figure 4). At the time that the IPA map was derived, the permafrost research in Scandinavia was focused in the northernmost areas, which might explain this pattern. The main difference between the maps is the level of detail. While the new permafrost map presented here has a raster resolution of 1 km^2, the IPA map is made in a mapping scale of 1:10,000,000. Assuming details down to 1 mm are visible in the IPA map, this corresponds to 10 km on the ground. The extent of the permafrost area outlined by the IPA map coincides well with the outer limits of permafrost in the new permafrost map. However, with the finer resolution we obtain a much higher level of detail of the permafrost distribution, in relation to the variation in topography and surface cover. This is reflected in the large area classified as lowland permafrost in Sweden, where only smaller occurrences of permafrost are present in the new map. Still, there is a very good accordance in the outer limits of permafrost occurrences between the maps.

In the IPA map 175 000 km^2 are classified as highland permafrost and 80 000 km^2 as lowland permafrost in the Scandinavian Peninsula. Because the new permafrost map does not separate between highland and lowland permafrost, but treat all permafrost areas except mires the same way, the areal numbers are not directly comparable. However, the total area of highland permafrost in the IPA map is almost 8 times as large as the total area of permafrost, mires excluded, in the new permafrost map. The total permafrost area including all types of permafrost (highland and lowland in the IPA map, and including mires in the new map) are almost 9 times larger in the IPA map than in the new permafrost map. The total permafrost area in the IPA map is distributed between the countries with 53 %, 34 % and 13 % in Norway, Sweden and Finland, respectively. This is in good accordance with the distribution in the new permafrost map.

6.2 Validation of the NGCD temperature data

The network of meteorological stations in Scandinavia follows the density of the population pattern and infrastructure. Consequently, the northern as well as the more mountainous areas feature a much more sporadic network of observations. This influences the quality of the interpolated meteorological data. Permafrost is restricted to areas with a very sparse station density. A validation with daily air temperatures measured at six meteorological stations located in permafrost areas are performed (Figure 6): Storkløftfjellet located at the top of the plateau at Varangerhalvøya (70.54°N, 29.34°E, 486 m a.s.l.), Iskoras (69.30°N, 25.35°E, 585 m a.s.l.), Vaisjeaggi (69.82°N, 27.17°E, 290 m a.s.l.), Abisko Research Station (67.92°N, 18.87°E, 355 m a.s.l.), Tarfalaryggen (67.92°N, 18.63°E, 1550 m a.s.l.) and Juvvasshøe (61.68°N, 8.38°E, 1894 m a.s.l.) (see all locations in Figure 2). Temperature observations from the Abisko and Juvvasshøe sites are included in the meteorological station network of the NGCD dataset, and the modelled temperatures at these two stations have therefore a significantly lower bias compared to the others, which are not included. In other words, given that Abisko and Juvvasshøe observations enter the spatial interpolation scheme for these sites we are evaluating the uncertainty due to the representativity error component of the observation (see Sect. 5.1). For the other sites we are evaluating the uncertainty is due to the prediction error.

The impact of the representativity error on the uncertainty in predictions is in general higher in winter than in summer, as can be seen in Figure 6 for Abisko and Juvvasshøe. However, it is quite evident that the agreement between modelled and observed values is better in Juvvasshøe than in Abisko, both in terms of bias and dispersion around the perfect prediction. The reason is in part due to the close relation between station density and representativity error: for areas characterized by a denser station network, as in southern Norway where Juvvasshøe is located, the uncertainties due to the representativity error are smaller than for areas with a sparser station network, as for the Abisko area in northern Sweden. In general, in OI, as in any Bayesian spatial interpolation scheme, the impact of representativity error on the prediction uncertainties can be reduced by improving the quality of the a priori information, for example by using output fields from high-resolution numerical models.

The distribution of summer (May – September) temperatures (red) are relatively well represented at all sites, with coefficients of determination (R^2) from 0.84 to 0.99 (Figure 6). The stations at Varangerhalvøya and Tarfalaryggen still feature warm biases of 1.6 °C and 1.1 °C, respectively. The representations of winter (October – April) temperatures (blue) are less accurate, with R^2 of 0.8 at Varangerhalvøya and Iskoras, and as low as 0.6 at Tarfalaryggen. All stations except for the one at Varangerhalvøya feature cold biases during the winter season, as large as -2 °C for Iskoras and Vaisjeaggi. These two stations are located at each side of the border in eastern Finnmark (Figure 2), where strong temperature inversions dominate during winter. The lack of stations at higher elevations prevents the inversions from being captured by the interpolation routine. The station Storkløftfjellet at Varangerhalvøya is 1.5 °C too warm in both summer and winter. The northernmost part of Finnmark has in general very few stations, all located along the coast in maritime settings. Because of the lack of stations at higher elevations, the maritime climate highly influences the interpolation at the top of Varangerhalvøya, but also higher elevated areas in large parts of northern Finnmark.

6.3 Evaluation of the permafrost model

The model results reproduce the measured mean annual ground temperature (*MAGT)* at the top of permafrost in boreholes within the range of the grid cell for 94 % of the observations (Figure 7). All observations are within ±2 °C of the average temperature of the grid cell, with an overall *RMSE* of 0.75 °C. The ground temperatures for the coldest boreholes are generally about 1 °C lower than the average modelled temperature. This is because they are all located at very exposed and snow free sites, and are representative for the colder end of the modelled distribution, and not the average.

The distribution of mountain permafrost corresponds well to all available permafrost probabilities derived from BTS-surveys (Figure 8). The agreement is particularly high in the Juvvasshøe area in central Jotunheimen, but also at Abisko and Dovrefjell the limits for both 50% (black contour line and purple colour) and the 80% (red contour line and blue colour) probability of permafrost are well represented (Figure 8). At Dovrefjell sporadic permafrost are modelled down to about c. 1350 m a.s.l., which is in accordance with observations indicating permafrost at bare blown areas down to this elevation (Sollid *et al.*, 2003). The model shows too little permafrost compared to the BTS-maps in Kilpisjärvi and at the two sites in southeast Norway; Elgåhogna and Sølen. In southeast Norway the temperature forcing has a warm bias, resulting in underestimation of permafrost in this area. From visual inspection, the distribution of permafrost is in good accordance with observations for most of Scandinavia. However, the model results show no permafrost in the northernmost parts of Finnmark, including Varangerhalvøya and Norkinnhalvøya. Permafrost has been documented from the higher parts of the plateau at Varangerhalvøya (Isaksen *et al.*, 2008; Farbrot *et al.*, 2013), while the permafrost model show sporadic to no permafrost in this area. This is related to the warm bias in the temperature forcing data for this area, seen in the validation at Storkløftfjellet (Sect. 6.2 and Figure 6)

6.4 Palsa distribution in Scandinavia

For the evaluation of the modelled permafrost in “snow free mires”, the national maps of palsa observations were resampled to 250 m resolution with all grid cells containing an area mapped as palsa included. At this resolution the total areal distribution of palsas is c. 1510 km^2, with 55 % in Norway, 32 % in Sweden and 13 % in Finland. The palsas are mainly located at elevations below 1000 m a.s.l., with almost 60 % at elevations between 350 and 500 meters a.s.l. (Figure 9). The amount of annual FDDs are mainly lower than -1500 °C*days, with more than 60 % of the locations colder than -2000 °C*days. This is in accordance with the findings by Aalto and Luoto (2014). Annual TDDs for most palsa areas are around 1000 °C*days, while maximum snow depths are below 0.6 meters. This pattern is fairly well reproduced by the model (Figure 9).

However, the modelled potential permafrost area at snow free mires covers as much as c. 5570 km^2, with 51 % of the area in Norway, 11 % in Sweden and 38% in Finland. Only 15 % of this area corresponds to grid cells with observed palsas, and the total modelled area significantly exceeds the mapped palsa distribution. The main reason for this is that only 60 % of the observed palsas are located in grid cells classified as “mire or peatland” in the CLC2012 land use map with 250 m resolution. In most cases there are grid cells classified as mires in the very near vicinity of palsas. However, in a few areas such as in the southernmost palsa area in western Sweden, the palsas are often classified as *sparsely vegetated area*. Another main reason for this overestimation is that the formation of a palsa depends on a range of local conditions, and even though the climate is cold enough to allow for a sustainable palsa, there may not be any palsa formation because of the local wind exposure and related snow drifting, peat layer, soil moisture or drainage patterns (Seppälä, 1988). A few years of favourable conditions can initiate the formation of a palsa that will remain in much less favourable conditions than when it was first established (Seppälä, 2011). The equilibrium model used in the new permafrost map can only provide an estimate of areas where already formed palsas would be likely to exist in equilibrium with the average climatic condition of the period 1981-2010. Palsas are therefore not present on all mires that are modelled to potentially have a palsa. In addition, permafrost in mires and peatland may be present even though they are not included in the palsa map. These maps, which are made from aerial photographs, do only include palsas that are visible in the aerial photographs. It should be noted that permafrost can also be present in peaty hummocks smaller than palsas (e.g. Luoto and Seppälä, 2002).

In some areas there are also biases introduced by inaccuracies in the map of mires and the climate forcing data. Out of the 60 % of the grid cells with observed palsas correctly classified as mire, 90 % are modelled as permafrost in the model run. The remaining 10 % are located in the northern parts of Finnmark, and in valleys close to Abisko. The underestimation in the northern parts of Finnmark is explained by the observed warm bias in the gridded temperature data at Varangerhalvøya (discussed in Section 6.2). The temperature validation to the meteorological station at Abisko shows very good correlation (Figure 6). However, in Katterjokk (~25 km west of Abisko) a palsa mire has degraded, and since the year 2000 the permafrost has largely disappeared (Åkerman and Johansson, 2008). In Stordalen (~10 km east of Abisko) palsas have degraded between 1970 and 2000 as a result of permafrost thaw (Johansson *et al.*, 2006). Because the model shows the distribution of potential palsa areas in an equilibrium situation, the palsas might correctly be represented as not having permafrost in equilibrium with the average climate over 1981 – 2010. In addition, local inversion phenomena in the surrounding valleys to Abisko could cause very cold winter conditions, which might not be well represented in the temperature data due to few meteorological stations located in this area.

The distribution map of potential palsa areas provides a good indication of areas where we can expect to find permafrost in mire environments (i.e. palsas and other permanently frozen peat hummocks). This map is also useful to detect areas where palsas are likely to thaw in the near future. The exact distribution is beyond the limitations of what an equilibrium model can represent, and would require a better representation of site-specific input parameters and the distribution of mires.

6.5 Response of permafrost to the climate change over the period 1981-2010

Significant increases in degree days are seen in the NGCD dataset over the period from 1980 to 2010. Time series are extracted from the dataset for four permafrost field sites Vaisjeaggi, Abisko, Tarfalaryggen and Juvvasshøe (Figure 2). All stations are validated with observations from the last few years of the period in Figure 6. The time series are shown as running means of annual thawing (TDD) and freezing (FDD) degree days in the air (Figure 10, a and b), and as annual maximum snow depths (Figure 10, c). The average value over each decade is drawn as a line.

At the two palsa mire sites, Vaisjeaggi and Abisko, the NGCD dataset shows an increase in annual TDD of 200 °C*days from the second to the third decade. The increase in TDD at the two high-mountain sites, Tarfalaryggen and Juvvasshøe, is less significant, while FDD increase with almost 1000 °C*days at Tarfalaryggen and 500 °C*days at Juvvasshøe. The inter-annual variation in maximum snow depth is high at all sites, but a general decrease in snow depth can be seen towards the end of the period. The first two decades are similar, and there is even a slight increase in snow depth from the first to the second decade at Tarfalaryggen and Vaisjeaggi (Figure 10, c).

The response in ground temperatures to this climatic change is demonstrated by comparing the potential permafrost distribution for three time slices (Figure 11). For this analysis, permafrost in snow free mires is not included. The total permafrost area in equilibrium with the climate of the first period is 36 800 km^2. The modelled decrease to the second decade given a new equilibrium is about 36 %, while to the third the total permafrost area is reduced by 64 %. The relative reduction for each period is similar for all three countries. A particularly large decrease is seen in the areas of discontinuous lowland permafrost in Finland, northern Sweden and the eastern parts of Finnmark (Figure 11, b). This clearly demonstrates that large areas with modelled occurrences of permafrost during the period 1980-1990, would not have permafrost in equilibrium with climate conditions similar to the period 2000-2010 (cf. Kukkonen and Šafanda, 2001).

The decrease in potential palsa areas is also significant, with a distribution restricted to Finnmark, Tavvavuoma and the area around Kilpisjärvi for the equilibrium situation with the average climate over 2001 – 2010 (Figure 11**Error! Reference source not found.**, d). In southern Norway, where the distribution of palsas is more marginal, an increase in elevation is seen from the first to the second period, while there are hardly any modelled potential palsa areas in the third time slice. However, palsas can still be found within the areas, but modelled as degraded, because palsas are slow systems with a significant time lag, and can remain in warmer climates (Seppälä, 2011). However, this indicates that palsas within these areas are vulnerable to degradation if the climate of the 2001 – 2010 period would continue, which is in accordance with the projected degradation of palsa mires by Fronzek *et al.* (2006) and Aalto *et al.* (2014).

6.6 A baseline map for permafrost

The main goal of this paper is to provide a first detailed baseline for permafrost distribution in Scandinavia. Compared to the more generalized IPA map this map provides a similar, but much more detailed picture of the distribution of mountain and lowland permafrost. The model is forced with climate data produced from operational meteorological stations, parameterized and evaluated with the use of all available field observations, and thoroughly validated by a large collaborating permafrost research community in Scandinavia. By combining simple modelling, a large ensemble of observations (for parameter calibration) and qualitative validation, the first permafrost map for the Scandinavian Peninsula provides a higher level of detail and stronger confidence in the model results than the larger scale IPA permafrost map (Brown *et al.*, 1997) for this region. For Scandinavia this map will certainly serve as a baseline model for validation of GCM or RCM-based permafrost evaluation in the future, along with already published large-scale attempts as published by Gruber (2012) and Westermann *et al.* (2015).

The implementation of a sub-grid variability of snow depths enables computation of the percentage of $MAGT$ < 0 °C in each grid cell. This allows for a representation of the permafrost extent with a gradual zonation. It enables sensitivity studies with increased snow depths or air temperatures, and because the full range of ground temperatures is represented, it will give a more realistic response to changes in the climate forcing. The map therefore provides a more dynamic aspect than the static IPA map, even one must be cautious to use such type of models for permafrost prediction under a changing climate..

Furthermore, the combination of climate, snow and soil parameters opens for local estimations of e.g. freezing depth and active layer thickness within a grid cell by employing semi-empirical relations like the Stefan solution (e.g. Nelson and Outcalt, 1987). The resolution is too coarse for detailed local assessments such as permafrost temperature distributions in steep slopes. However, the resolution and zonation is sufficient to assess in which areas problems related to frozen ground can be expected to be found. Thereby it may also serve to provide useful estimates of boundary conditions necessary for physically-based modelling accounting for active layer dynamics (e.g. Frampton and Destouni, 2015).

In the IPA map permafrost, palsas were not represented separately in Finland, Norway and Sweden, but instead incorporated in the permafrost class "*Sporadic to isolated patches of lowland permafrost*". Some perennial frost mounds were identified, however, in other permafrost regions. Still, compared to the observed palsa distribution, there is a large discrepancy in the level of detail between the IPA map and the permafrost map for the Scandinavian Peninsula. This new permafrost map shows clear improvements in the representation of the areas of potential palsas and other permafrost features by using a fine representation of the distribution of peat and mire areas, combined with air temperatures and snow at 1 km^2 resolution. In this way we are able to model the distribution of mires with potential for permafrost under favourable local conditions.

The spatial extent of the mire areas with potential permafrost fit relatively well with the extent of palsas. However, while the model projects permafrost in all mires of an area with a climate cold enough for palsa formation, palsas are only present at sporadic locations in the mires. Because of the highly sporadic nature and slow climatic response of palsas, it is difficult to obtain a precise representation of palsa features in regional models. The spatial variation is very high for a range of variables, including soil moisture and drainage patterns, snow distribution, and thickness and thermal properties of the peat layer (Seppälä, 1988; 2011). To sufficiently model the transient evolution of a palsa, a detailed description of all these parameters and related processes must be included (e.g. An and Allard, 1995; Kujala *et al.*, 2008). This is a challenging task on a local scale, and not possible in such regional models.

7. Conclusions

In this paper a baseline permafrost map for Norway, Sweden and Finland is presented. The map is thoroughly validated and consistent with field thermal and landform observations. Borehole observations are well within ±2 °C of the average modelled top permafrost temperature in the grid cell with an overall *RMSE* of 0.75 °C. Qualitative evaluation of the permafrost zonation indicates that the general accuracy in the lower altitudinal limit of permafrost is within 100 meters, and that the distribution fits well against BTS-mappings.

Based on these more detailed descriptions of the permafrost distribution of the Scandinavian Peninsula, the following main conclusions are obtained:

- C. 23 400 km^2 of the land area is underlain by permafrost (palsas excluded) in equilibrium with the 1981 – 2010 climate. About 56 % of the permafrost area is within Norway, 35 % in Sweden, and 9 % in Finland.
- The distribution of permafrost (mires excluded) is 60 % less in equilibrium with the climate over the period 2000 – 2010 than with the climate of 1981 – 1990. This indicates that large areas are thawing with possible degradation in the lowland permafrost in northeast Scandinavia, as well as degradation of sporadic permafrost in the coastal mountain areas in Troms and Finnmark.
- The mapped palsas cover c. 1510 km^2; 55 % in Norway, 32 % in Sweden, and 13 % in Finland. According to the model results, large parts of the mapped palsas are located in areas with significant warming of the permafrost during the 1981 – 2010 period. This indicates that the palsas are not in equilibrium with the climate of the last decade, due to their slow climatic response.
- The modelled permafrost extent coincides well with the outer boundaries of the permafrost in the IPA map. However, because of the much higher level of detail in the

new permafrost map, the total permafrost area of the IPA map is 9 times that in the new map.

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9. Supporting Information

Figure S1: The new permafrost map for Norway, Sweden and Finland classified in continuous (blue), discontinuous (purple) and sporadic (pink) permafrost zones. Areas with isolated patches of permafrost in mires are shown in brown colour.

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Figure captions:

Figure 1: Spatial distribution of mean annual air temperatures (left) and seasonal maximum snow depths (right) averaged over the period 1981 – 2010, based on the NGCD dataset.

Figure 2: Boreholes with permafrost and seasonal frost are indicated with blue and yellow dots, respectively, and basal temperature of snow (BTS)-surveys are outlined in purple. The main sites of permafrost field investiagations are marked in letters: A = Varangerhalvøya, B = Vaisjeaggi, C = Iskoras, D = Kilpisjärvi, E = Tavvavuoma, F = Abisko, G = Tarfalaryggen, H = Dovrefjell, I = Elgåhogna, J = Sølen, K = Juvvasshøe and central Jotunheimen, L = Finse, and M = Gaustatoppen. The lower altitudinal limits of permafrost for the respective sites [m a.s.l.] are given in the parenthesis behind the letters.

Figure 3: Schematic of the model setup.

Figure 4: Comparison of the permafrost map for the Scandinavian Peninsula, to the International Permafrost Association Circum-Arctic permafrost map (Brown *et al.*, 1997).The inset panel (lower right) is the continuation of the main map to the south.

Figure 5: Lower altitudinal limits of permafrost (LALP) related to distance from the coast (left) in southern (S) and northern (N) Norway, and to latitude (right). Red, blue and black lines indicate LALP for areas with 10 %, 50 % and 90 % permafrost, respectively.

Figure 6: Temperature validation at six permafrost sites in Scandinavia. Temperature data from Abisko and Juvvasshøe weather stations are included in the interpolation for the gridded weather data.

Figure 7: Measured vs. modelled *MAGT* at top of permafrost atall boreholes in mountain areas in Scandinavia being part of the GTN-P. The error bars represent the 2.5th and the 97.5th percentiles of all model realization; the dashed lines are the ±2 °C intervals around the 1:1 line (in solid). Figure 8: Comparison of modelled lower altitudinal permafrost limits, here represented as percentage of modelled permafrost per grid cell, and permafrost probability derived from BTS-surveys at Kilpisjärvi (Majeva, 2014), Abisko (Ridefelt et al., 2008), Dovrefjell and Juvvasshøe (Isaksen *et al.*, 2002), and Elgåhogna and Sølen (Heggem *et al.*, 2005). The 50 % probability of permafrost (thin, grey contour line) should correspond to the purple area in the modelled permafrost zonation, while the 80 % probability of permafrost (thick, black contour line) should outline the dark blue area.

BTS-surveys at Kilpisjärvi (Majava, 2014), Abisko (Ridefelt *et al.*, 2008), Dovrefjell and Juvvasshøe (Isaksen *et al.*, 2002), and Elgåhogna and Sølen (Heggem *et al.*, 2005)

Figure 9: The distribution of elevation, maximum snow depths (Snow depth), freezing and thawing degree days in the air (FDD/TDD) in locations mapped as palsas (*Palsas*, green), modelled as permafrost in mires with snow free conditions (*PF mires*, blue) and areas represented as mires in the CLC2012 raster map with 250 m resolution (*Mires*).

Figure 10: The graphs show running mean of annual thawing and freezing degree days in the air (*TDD* and *FDD*), as well annual maximum snow depth (max SD), over the 30-year period 1981 – 2010. Data are extracted from the NGCD dataset for the four locations Vaisjeaggi, Abisko, Tarfalaryggen and Juvvasshøe (see Figure 2 for locations). The stippled lines indicate the average values over each of the three decades.

Figure 11: The upper row is a comparison of modelled distribution of a) sporadic and b) discontinuous permafrost in equilibrium with the climate of the three decades in the period 1981 – 2010. In the lower row the distribution of potential permafrost in mires under “palsa conditions” is shown for northern Scandinavia (c) and Dovrefjell (d).

Tables:

Table 1: n-factor classification assigned to surface cover classes in the 4th CORINE Land Cover Map finalized in 2012 (CLC12). The classification is based on Gisnås *et al.* (2013) and additional data from Tavvavuoma (Sweden), Vaisjeaggi (Finland), Kilpisjärvi (Finland) and Iskoras (Norway). In open areas with sparse vegetation cover, n-factors vary with maximum snow depth (SD) following the given functions. For mire and peat, values for snow free conditions are given in the parenthesis.

	Surface cover	CLC12	*nT*	*nF*
1	Coniferous forest	312, 313	0.95	0.35
2	Broad-leaved forest, transitional woodland and shrub	311, 324	1.05	0.25
3	Mire and peat	412, 413	1.25 (1.20)	0.26 (0.60)
4	Bedrock and sparsely vegetated area, glaciers	321, 322, 331, 332, 333, 334, 335	-0.14*SD + 1.1	-0.16*log(SD) + 0.22
5	Blockfield	From sediment maps	-0.14*SD + 0.1	-0.16*log(SD) + 0.22
	Agricultural and artificial surfaces, water bodies	100 to 299, 500 to 599	No Value	No Value

Table 2: Thermal conductivities ($Wm^{-1}K^{-1}$) are given for the ground in thawed (κ_t) and frozen (κ_f) states. r_k is the ratio of κ_t and κ_f.

ID	Surficial Deposits	κ_t	κ_f	r_k
2	Beach and coastal deltas	1,46	1,73	0,84
3	Coarse and fine rubble / mountain alluvium and colluvium	1,83	2,15	0,85
4	Valley loess and alluvium / eolian	1,57	1,80	0,87
5	Uppland loess / eolian	1,59	1,90	0,83
6	Lightly modified moraine	1,47	2,13	0,69
7	Current moraine / glacial moraines & drift	1,52	1,87	0,81
11	Glaciofluvial deposits	1,97	2,05	0,96
12	Galciolacustrine deposits	1,65	2,23	0,74
13	Glacio - fluvial deposits	1,45	1,88	0,77
16	Fluvial deposits	1,65	1,99	0,83
17	Undifferentiated alluvium & colluvium	1,43	1,82	0,79
18	Coastal delta / coastal	1,86	2,11	0,88
19	Fine rubble, mountain alluvium & colluvium	1,87	2,13	0,88
20	Weathering material	1,96	2,16	0,91
21	Eolian deposits	1,98	2,01	0,99
22	Old marine & alluvium / coastal	1,41	1,97	0,72
26	Highly modified moraine / glacial moraines & drift	1,91	1,92	1,00

27	Mires and peat, thick	0,37	0,50	0,74
28	Mires and peat, thin	1,65	2,23	0,74
29	Bedrock	3,50	3,50	1,00
30	Blockfield	2,95	3,00	0,98
31	Thin sediment class / dry moraine	3,00	3,00	1,00