



Modelling of the thermal regime of permafrost during 1990–2014 in Hornsund, Svalbard

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Abstract: The thermal state of permafrost is a crucial indicator of environmental changes occurring in the Arctic. The monitoring of ground temperatures in Svalbard has been carried out in instrumented boreholes, although only few are deeper than 10 m and none are located in southern part of Spitsbergen. Only one of them, Janssonhaugen, located in central part of the island, provides the ground temperature data down to 100 m. Recent studies have proved that significant warming of the ground surface temperatures, observed especially in the last three decades, can be detected not only just few meters below the surface, but reaches much deeper layers. The aim of this paper is evaluation of the permafrost state in the vicinity of the Polish Polar Station in Hornsund using the numerical heat transfer model CryoGrid 2. The model is calibrated with ground temperature data collected from a 2 m deep borehole established in 2013 and then validated with data from the period 1990–2014 from five depths up to 1 m, measured routinely at the Hornsund meteorological station. The study estimates modelled ground thermal profile down to 100 m in depth and presents the evolution of the ground thermal regime in the last 25 years. The simulated subsurface temperature trumpet shows that multiannual variability in that period can reach 25 m in depth. The changes of the ground thermal regime correspond to an increasing trend of air temperatures observed in Hornsund and general warming across Svalbard.

Key words: Arctic, Spitsbergen, Hornsund, permafrost modelling, permafrost evolution.

Introduction

Permafrost and the active layer are crucial factors for all land ecosystems on Svalbard. Changes of ground temperatures have an impact on groundwater flow, hydrological cycles (Kurylyk *et al.* 2014), periglacial processes and geomorphological landforms (Humlum *et al.* 2003; Harris *et al.* 2009), vegetation (Myers-Smith *et al.* 2015), as well as fauna (Moe *et al.* 2009) and linkages between them.

Ground temperatures and thermal properties, as well as active layer thickness are influenced by many factors. Of major importance are the geographical, climatological and meteorological variables, such as location, slope exposition, elevation, distance to open waters and seas, sediment type, water content, snow cover, albedo of the ground surface, vegetation and others (*e.g.* Westermann *et al.* 2009). Due to that, the active layer thickness varies in non-glaciated, coastal areas in Svalbard from less than 1 m in higher latitudes, for example in Kinnvika (80°N), to around 2 m in the Hornsund area (77°N) (Dolnicki *et al.* 2013).

Being one of the most important indicators of climate change, permafrost phenomena have been topics of many scientific studies. Measurements of ground temperature have been part of the environmental monitoring for many years in just few places in Svalbard. There are few boreholes deeper than 10 m in Spitsbergen (Permafrost Observatory Project), but none of them is located south of Van Mijenfjorden, so that modelling is so far the only solution to describe the thermal conditions of the ground in the Hornsund area.

The aim of this study is the recognition of the permafrost state in the vicinity of the Polish Polar Station in Hornsund using a heat conduction model. We focus on the estimation of initial conditions at depths between 2 and 100 m and a sensitivity analysis of the model parameters on model output. The model is calibrated with the ground temperature data collected in the period from September 2013 to June 2014 from eight depths: 5, 10, 20, 50, 100, 150, 175 and 200 cm. The model is validated using the data from period 1990–2014 from five depths: 5, 10, 20, 50, 100 cm measured routinely at Hornsund meteorological station. As a result we estimate ground thermal regime up to 100 m below the surface.

Material and methods

Study area. — The study site is located on the northern shore of Hornsundfjord in Wedel Jarlsberg Land in SW Spitsbergen, in the close vicinity of the Polish Polar Station (77°00' N, 15°33' E; Fig. 1). Due to the strong influence of a warm West Spitsbergen current, the climate is maritime and mild compared to other Arctic areas on the same latitude. The mean annual air temperature in the years 1979–2014 has been measured to -4.0°C. The coldest month is March with the average temperature -10.4°C. The mean annual sum of precipitation amounts 453 mm. The annual average of days with the snow cover in the period 1983–2014 is 238 and the mean snow cover is 17 cm. The state of the climate in Hornsund, with 31 years long data record, is described in a monograph by Marsz *et al.* (2013a). Meteorological data are available online in the series of bulletins at <http://www.glacio-topoclim.org/index.php/reports> (Hornsund Glacio-Topoclim Database).

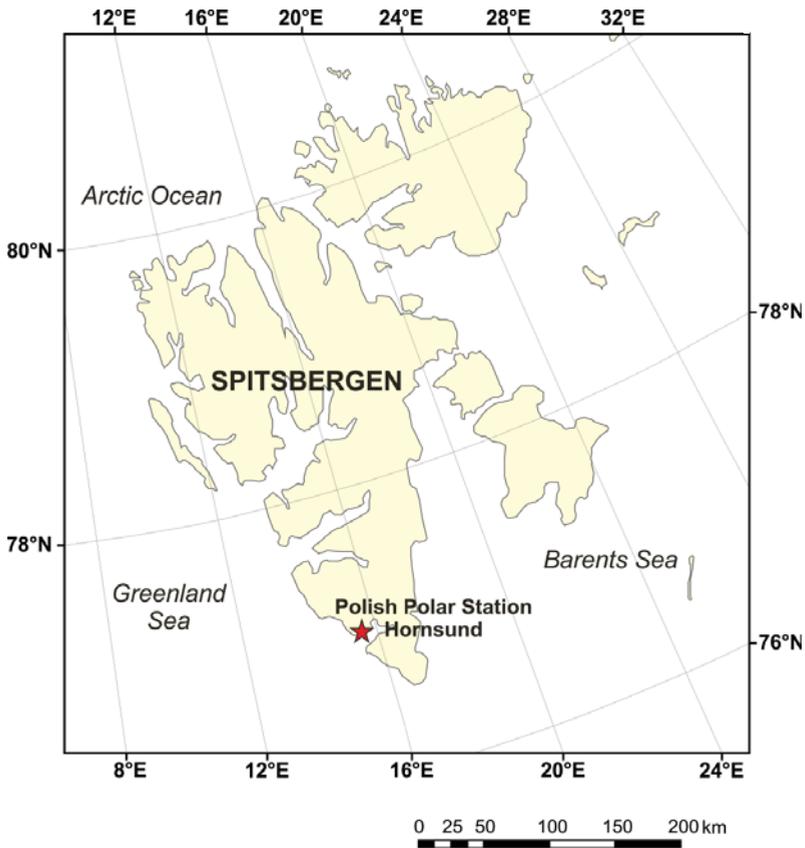


Fig. 1. The location of Polish Polar Station in Hornsund.

The first measurements of the ground thermal state in the vicinity of Polish Polar Station in Hornsund were initiated during the first Polish overwintering at Hornsund base in 1957/1958 and were continued in a few following summer seasons (Baranowski 1968). Since July 1978 year round observations were renewed and are continued until now. Variations of ground thermal conditions have been measured routinely at four depths: 5, 10, 20, 50 cm and since the 1980-s also at 100 cm. From the beginning of measurements, bent-stem soil thermometers at 5, 10, 20 and 50 cm and at 100 cm thermometer with a large time scale constant in plastic pipe were used. In 2001 those bent stem thermometers were replaced by thermistor sensors Pt-100, that are less fragile, more reliable and collect the data every 1 minute. The only thermometer that remained is the one measuring ground temperature at 100 cm.

In the second half of August 2013, two new boreholes were established using a drilling rig on raised marine terraces, composed of marine sediments

containing a mixture of sand and gravel with clay. Beneath 4–5 m thick marine deposits lies crystalline basement of metamorphic schists, paragneisses and marbles (Szymański *et al.* 2013). The first borehole was located at the Hornsund meteorological site, the second 300 m to the east in a small (1.3 km²), non-glaciated Fuglebekken catchment, around 150 m from the shore of Isbjørnhamna. The installed thermistor chains were STG-073 with eight sensors (Pt-100 cl. 1/3B DIN 43760 with accuracy of 0.1°C), measuring temperatures at depths: 5, 10, 20, 50, 100, 150, 175 and 200 cm. The data were collected on loggers in 60 minutes interval. The first logger was destroyed, so that in this study we present data from the logger in Fuglebekken.

The thermal state of the ground and thickness of the active layer in this area have been topic of broad studies. Jahn (1982, 1988) and Grześ (1984) described soil structures and permafrost-related geomorphological processes. Mięgała (1991, 1994) and Dolnicki (2005) investigated influence of snow cover melting on ground temperatures. Miętus (1988), Miętus and Filipiak (2001), Leszkiewicz and Caputa (2004), and Dolnicki (2010, 2013) discussed variability of the active layer depth. Dobiński and Leszkiewicz (2010) investigated the permafrost occurrence on the offshore terraces using electroresistivity soundings.

According to Humlum *et al.* (2003), permafrost in Svalbard is generally continuous with thickness ranging from 100 m near the coasts to more than 500 m in interior mountains. Spatial variations of permafrost state in Svalbard have been investigated by multiple researchers. Ground thermal conditions in Kapp Line, Svea, Ny-Ålesund and in Longyearbyen surroundings have been described by Christiansen *et al.* (2010), in Longyearbyen by Harada and Yoshikawa (1998), in Petuniabukta by Rachlewicz and Szczuciński (2008), in Bellsund by Marsz *et al.* (2013b), and in Kaffiøyra Plain by Sobota and Nowak (2014).

Climatic context and snow conditions. — The variability of air temperatures in Svalbard depends on the season. Summer temperatures are rather constant, with monthly means reaching slightly above 5°C. Average monthly temperatures during winter and early spring usually drop below -10°C. The annual fluctuations of air temperature during different seasons are reflected in a damped way a few meters below the ground surface. Although this fluctuation diminishes rapidly with depth, climate warming observed during last decades is detectable in changes of permafrost temperature at depths even up to 60 m (Isaksen *et al.* 2007). The maximal, minimal and mean daily air temperatures recorded at the Polish Polar Station in Hornsund from 1990 till 2014 are shown on Fig. 2a. The red line represents the air temperature in the period from September 2013 to June 2014, which we used as the upper boundary condition for the heat conduction model. After a warm September and average October and November, air temperature suddenly dropped in the end of November/beginning of December, but in following months, especially in February 2014, large positive anomalies

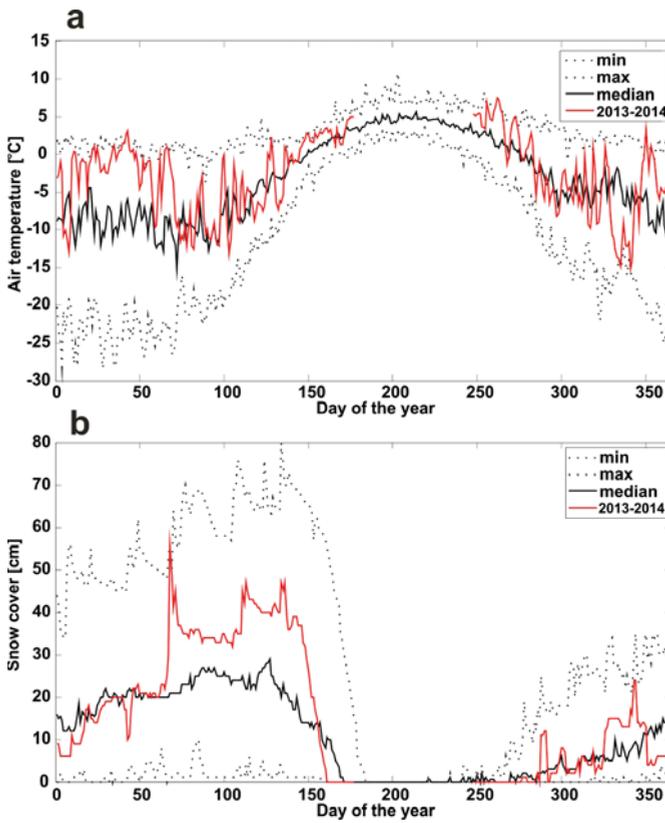


Fig. 2. Maximum, median and minimum daily air temperatures (a) and snow cover thickness (b) observed at Polish Polar Station in Hornsund in years 1990–2014. Red curve denotes observations in the period from September 2013 to June 2014.

were recorded. Anomaly events on the 17th of December 2013 and the 12th of February 2014 featured the highest daily air temperature ever recorded in these months at Hornsund reaching +4.7°C and +4.4°C, respectively. Such positive temperatures influenced the snow cover for short periods of time.

Snow thicknesses on SW Spitsbergen are highly variable in the accumulation period due to the strong wind drift and thaws that occur seasonally during winter time (Luks *et al.* 2011; Marsz *et al.* 2013b). From September 2013 to June 2014, snow covered the ground during 241 days. It appeared in the beginning of October but then lasted for a few days 4–6, 12–13 and 15–18th. Since the 21st of October 2013 until the 9th of June 2014, snow covered the ground continuously, but with changing thickness (Fig. 2b). The maximum thickness of the snow cover was observed on the 6th of March 2014, when it reached 56 cm. The average thickness in analyzed period of time was 21 cm, which is 4 cm higher than the multi-annual mean.

CryoGrid 2 model description. — Due to the limitations of the available data at depths lower than 1 m in Hornsund, numerical modelling of the most important processes is necessary to understand and describe the current changes of permafrost temperatures. Such approaches are important for impact assessment of the warming observed in the Arctic in recent decades, as well as the future warming projected by general circulation models (Etzelmüller *et al.* 2011).

CryoGrid 2 is a numerical framework to calculate the distribution of ground temperatures based on time series of air temperature and snow depth (Westermann *et al.* 2013). CryoGrid 2 is composed of two modules: the soil domain module and the snow domain module. The first module corresponds to the domain below the ground surface (positive vertical coordinate z) and the second to snow domain above the ground surface (negative z).

In the soil domain, variation of temperature T over time t in the given region is described by means of the 1-dimensional heat conduction equation (Jury and Horton 2004).

$$c_{eff}(z, T) \frac{\partial T}{\partial t} - \frac{\partial}{\partial z} \left(k(z, T) \frac{\partial T}{\partial z} \right) = 0 \quad (1)$$

where c_{eff} [J/m³K] denotes the effective volumetric heat capacity and $k(z, T)$ [W/mK] is the thermal conductivity.

The effective volumetric heat capacity c_{eff} accounts for the latent heat of melting/freezing of ice/water and depends on volumetric water content, the specific volumetric latent heat of fusion of water and the volumetric contents and the specific volumetric heat capacities (Hillel 1982) of the following constituents: water, ice, air, mineral and organic.

The thermal conductivity of the soil k is calculated from the volumetric fractions of the soil constituents: water, ice, air, mineral and organic.

In the snow domain, the only process of energy transfer considered in CryoGrid 2 is heat conduction (similar to Eq. 1). The thermal conductivity and volumetric heat capacity of the snowpack are assumed to be constant in time and space. The volumetric heat capacity is calculated from the densities of snow and ice.

As upper boundary condition, daily time series of air temperature (measured 2 m above the ground) and the snow cover depth are used. A geothermal heat flux is prescribed at the lower boundary.

The method of lines is used to solve the heat transfer equation (Eq. 1) numerically for a soil domain of 100 m depth. The partial differential Eq. 1 is replaced by a system of coupled ordinary differential equations (ODE) with time as only variable. This is achieved by using central finite differences to discretize the spatial derivatives on a grid, while the time t is left as a continuous variable. The system of ODE is solved numerically by means of ode45 Matlab solver.

To decrease the dimensionality of the problem but accurately describe the dynamic of the processes involved, and to account for larger temperature gra-

dients closer to the surface than in higher depths, the grid spacing increases with depth and is set in the following way:

- i) $\Delta z = 0.05$ m between ground surface and maximum snow depth,
- ii) $\Delta z = 0.01$ m from 0 to 1 m,
- iii) $\Delta z = 0.05$ m from 1 m to 3 m
- iv) $\Delta z = 0.2$ m from 3 m to 5 m
- v) $\Delta z = 0.5$ m from 5 m to 20 m
- vi) $\Delta z = 1.0$ m from 20 m to 30 m
- vii) $\Delta z = 5.0$ m from 30 m to 50 m
- viii) $\Delta z = 10.0$ m from 50 m to 100 m.

For each cell, only the heat flow in the vertical direction is considered and the lateral heat flow is discarded, so the change of internal energy and the temperature in the ground is described by means of 1D partial differential equation based on Fourier's law of heat conduction and latent heat from freezing or thawing of the ground. Movement of ground water is not considered in this model. In the model we assumed sediment stratigraphy that differs from marine sediments in the upper part — from the surface to the depth of 5 m, and crystalline bedrock underneath (Szymański *et al.* 2013).

The selected version of the CryoGrid 2 model includes seven parameters: volumetric water content zone 1 (VWC1) from the ground surface to 2 m below, volumetric water content zone 2 (VWC2) from 2 to 5 m, volumetric water content zone 3 (VWC3) from 5 to 100 m, heat flux at lower boundary (Q), density of the snow [kg/m³], thermal conductivity of the snow [W/mK] and thermal conductivity of the mineral component of the soil [W/mK].

The CryoGrid 2 model was calibrated based on ground temperature data measured at the new borehole in the Fuglebekken catchment. The data set consists of temperatures measured by eight sensors at depths $z = [5, 10, 20, 50, 100, 150, 175, 200$ cm] in the period from the 5th of September 2013 to the 26th of June 2014. As the objective function, a sum of Nash–Sutcliffe (NS) coefficients calculated for these eight selected depths was used (Nash and Sutcliffe 1970).

$$J = \sum_{i=1}^8 NS_i \quad (2)$$

$$NS_i = 1 - \frac{\sum_{t=1}^{t_{\max}} (T_{io}^t - T_{im}^t)^2}{\sum_{t=1}^{t_{\max}} (T_{io}^t - \bar{T}_{io}^t)^2} \quad (3)$$

where \bar{T}_{io} is the mean of observed temperatures, and T_{im} is modeled temperature at the depth z [i], T_{io} is observed temperature at time t . For any depth, the Nash–Sutcliffe efficiency can range from $-\infty$ to 1, so an efficiency of 8 corresponds to a perfect match of modeled temperature to the observed data at 8 depths.

The CryoGrid 2 model was calibrated using Differential Evolution with Global and Local neighbours (DEGL) (Das *et al.* 2009; Piotrowski and Napiórkowski 2012).

Results

Estimation of initial conditions. — The results of CryoGrid 2 model are highly influenced by initial conditions, *i.e.* ground temperatures at entire analyzed space domain from 0 to 100 m below the surface. There is no such deep borehole in southern Spitsbergen area, so that the initial ground temperature profile in Hornsund has to be estimated.

On basis of the preliminary analysis of the ground temperature variation in the analysed domain, the profile was divided into three layers:

- a) the closest to surface (0–2 m) — with high seasonal fluctuations (active layer) where temperatures were measured;
- b) 2–30 m — with seasonal temperature fluctuations; no measurements;
- c) 30–100 m — no significant fluctuations in short time scale (amplitude less than 0.1°C), no measurements.

Afterwards, the initial ground temperature was formed by the combination of three temperature profiles corresponding to the three layers defined above.

Estimation of profile temperature in the deepest layer (closest to downstream boundary, *i.e.* from 30 to 100) meters was conducted using steady state simulations. The equilibrium conditions were calculated by solving Eq. 1 for partial derivative with respect to $t = 0$, imposed flux at the upstream boundary that corresponded to geothermal heat flux Q :

$$-k \frac{dT}{dz} = Q \quad (4)$$

and for assumed values of the thermal conductivity k , the average snow cover in period 1983–2014, and the average air temperature in Hornsund in the period 1979–2014 (upper boundary condition). The obtained solution for equilibrium conditions is represented as straight line in Fig. 3. The obtained temperature profile has similar course as measured ground temperatures in the Janssonhaugen borehole (Isaksen *et al.* 2007).

The initial ground temperature in the layer closest to the surface, *i.e.* from 0 to 2 m was formed by recorded temperatures taken from the first day of observations the 5th of September 2013.

The layer between 2 and 30 m is affected by seasonal fluctuations, so that the temperature profile cannot be considered steady throughout the year and the initial boundary condition could not be estimated directly from the steady state simulations. For that reason it was calculated by cubic interpolation between simulated and observed data (Fig. 3).

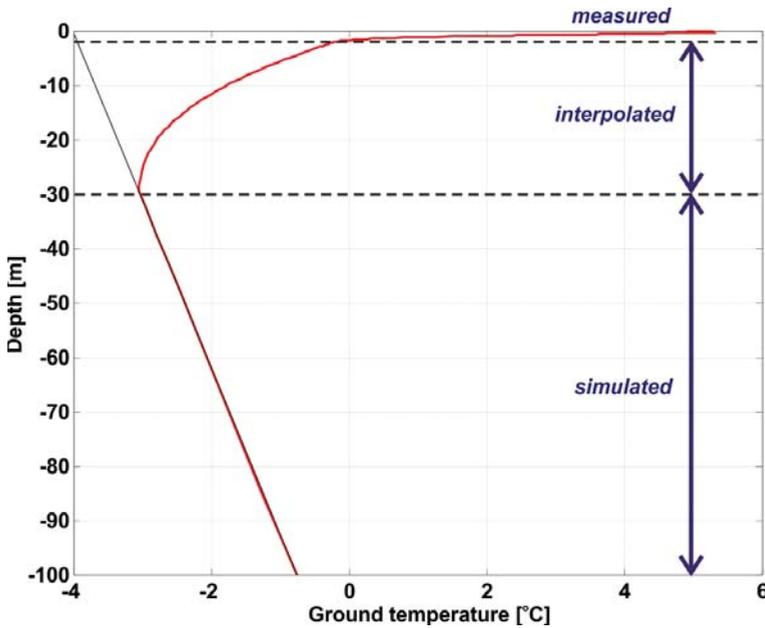


Fig. 3. Assumed initial conditions of the ground temperature profile in Hornsund – measured and simulated ground temperatures on the 5th of September 2013.

Then the objective function (Eq. 2) was minimized with respect to Cryo-Grid 2 model parameters (see next sections for details). Based on the obtained solution, the coefficients of the effective volumetric heat capacity and the thermal conductivity were corrected and the procedure was repeated until the estimated initial condition was approved for further calculations.

The results (initial conditions and optimal parameter values) allow us to obtain a rough estimate for the permafrost base in close surroundings of Polish Polar Station in Hornsund for chosen boundary conditions. In the case of $Q=0.04$ [W/m²] and the average air temperature of -4°C, the depth of permafrost equals around 120 m. These estimates are strongly influenced by the assumed initial conditions and the choice of model parameters. We also estimated the uncertainty of depth of the permafrost by Monte Carlo analysis taking into account the variability of values of the geothermal heat flux at lower boundary from 0.04 to 0.07 [W/m²], volumetric water content from 3 to 6%, and thermal conductivity of the mineral component of the soil from 1.0 to 1.8 [W/mK] in the layer located between 5 and 100 m from the surface. As a result the median permafrost depth equal to 104 m was estimated. As expected, the highest depth of permafrost, around 182 m was estimated for $Q=0.04$ W/m², highest thermal conductivity 1.8 W/mK and volumetric water content reaching 6%. In turn, the smallest thickness of permafrost, almost 58 m was estimated for $Q=0.07$ [W/m²],

lowest thermal conductivity 1.0 W/mK and volumetric water content 3%. These analyses indicate that the estimated permafrost thickness in the Hornsund area might vary up to 50% due to assumed variation of model parameter values. We furthermore emphasize that the location of the permafrost base is influenced by the climate conditions of a much larger period than considered here, possibly the entire period since deglaciation (when permafrost-free conditions can be assumed and permafrost began to form). Our treatment implicitly assumes that the permafrost base is already in equilibrium with the Holocene climate, so that the steady-state assumption can be applied. While our rough estimate is in agreement with previous estimates in coastal areas in Svalbard (Humlum *et al.* 2003), only in situ measurements in a deep borehole may provide the true permafrost depth.

Sensitivity analysis by Morris method. — A sensitivity analysis allows deriving diagnostic insights from models by identifying the key input parameters controlling the performance of a model (Herman *et al.* 2013).

To assess influence of CryoGrid 2 model parameters on the modelled output, a sensitivity analysis using the Morris method was applied (Morris 1991). In this method, the sensitivity is estimated on the basis of a number of local changes of the objective function J at different points of the possible range of model parameters (see Table 1), called elementary effects. The global results are obtained by regional exploration of the input space (a chosen number of trajectories).

Each parameter x_i is perturbed along a grid of size Δ_i in the chosen parameter space (between selected limits defined in Table 1). It is assumed that for a model with p parameters each consist of a sequence of p perturbation. Each trajectory allows estimating one elementary effect for each parameter. The elementary effect for parameter i is calculated according to the following equation:

$$EE_i = \frac{J(x_1, \dots, x_i + \Delta_i, \dots, x_p) - J(x)}{\Delta_i} \quad (5)$$

where $J(x)$ denotes the prior point in the trajectory. The analysis with a single trajectory is strongly dependent on the location of the initial point in the parameter space. Therefore in the Morris method, the analyses are repeated for a chosen number of trajectories sampled according to the selected approach. In the approach originally proposed by Morris, the trajectories are randomly sampled over the parameter space. More accurate estimates of global sensitivity are obtained following Campolongo *et al.* (2007, 2011) and Ruano *et al.* (2012) where trajectories that maximize coverage of parameter space are selected.

The results of this method are presented in the form of three measures: μ , μ^* and σ . The first two measures represent the overall influence of the CryoGrid 2 model parameter on the output. The first measure, μ is the original measure proposed by Morris (1991) and it is calculated according to the equation:

$$\mu_i = \frac{\sum_{i=1}^r EE_i}{r} \quad (6)$$

The second measure, μ^* , was introduced by Campolongo *et al.* (2007) and it is computed from the mean of the absolute values of the elementary effects which allows the problem of effects of an opposite sign to be solved according to the equation:

$$\mu_i^* = \frac{\sum_{i=1}^r |EE_i|}{r} \quad (7)$$

The third measure, σ , estimates the ensemble of the factor's higher order effects (*i.e.* nonlinearity or interaction with other parameters) according to the equation:

$$\sigma_i = \frac{\sum_{i=1}^r (EE_i - \mu_i)^2}{r} \quad (8)$$

These three measures allow for a classification of the CryoGrid 2 model parameters in two groups, taking into account their effect on the analysed model output (objective function J) defined by equations 2 and 3 as a sum of NS coefficients at eight depths). The first group consists of parameters with a significant influence on an objective function. Other parameters without a significant impact on the value of an objective function are in the second group.

The results of sensitivity analysis by Morris method are shown on Fig. 4 in the form of the influence of seven model parameters on the applied objective function. To make the chart more readable, the scale was limited, and the parameter Q that has the least influence is not presented.

Table 1

Description of CryoGrid 2 parameters and their limits used in sensitivity analysis and calibration.

Nr	Parameter	Unit	Abbreviation	min	max
1	Volumetric water content zone 1	[%]	VWC1	1.00	30
2	Volumetric water content zone 2	[%]	VWC2	1.00	30
3	Volumetric water content zone 3	[%]	VWC3	1.00	30
4	Heat flux at lower boundary	[W/m ²]	Q	0.04	0.07
5	Density of the snow	[kg/m ³]	ρ_{snow}	100	600
6	Thermal conductivity of the snow	[W/mK]	k_{snow}	0.05	0.5
7	Thermal conductivity of the mineral component of the soil	[W/mK]	k_{bedrock}	0.01	5.0

The highest values indicate the most significant influence on the objective function value, while the values closer to 0 indicate a lack of significant influence. The VWC1 parameter was found to have the highest influence on the objective function. Two additional parameters (k_{snow} , and k_{bedrock}) are also very significant. The influence of the volumetric water content in zone 2 and 3 on the analyzed model output is less important. The snow density ρ_{snow} shows a much smaller significance, while the parameter Q does not have influence on model output.

As a result of sensitivity analysis the identification of the model parameters was simplified, as only three parameters influenced the Nash–Sutcliffe coefficient defined in Eq. 2.

The second group of four parameters (VWC2, VWC3, ρ_{snow} , and Q) without significant influence on the model output was considered constant and did not take part in the model calibration.

The influence of parameter values on model fit at different depths was also tested. In accordance to our expectations, the influence of VWC3 is most significant in the deeper layers (Fig. 5a). Figure 5b shows the growing significance of k_{bedrock} with depth. According to the estimated μ^* values, the influence of k_{bedrock} is twenty times higher at 200 cm than at 5 cm.

The results of this simplified sensitivity analysis were confirmed by the more computationally demanding Sobol method. The Sobol method, based on the analysis of variance output from the model (Archer *et al.* 1997) is the second approach applied. In this method, it is assumed that the total variance of the output from the model, calculated using the Monte Carlo approach, can be decomposed into the sum of the variance of each input factor explained by

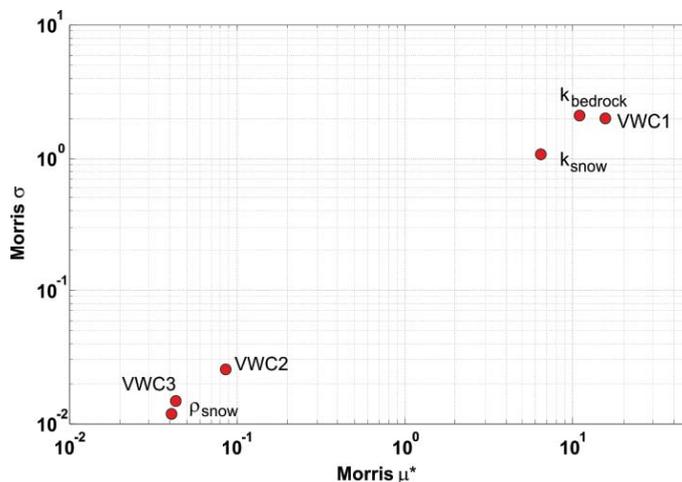


Fig. 4. The results of sensitivity analysis by the Morris method. The influence of seven model parameters on the applied objective function (J).

the fractional contribution (CryoGrid 2 model parameters) and by interactions between them by the equation:

$$V(J) = \sum_i V_i + \sum_{i < j} V_{ij} + \sum_{i < j < m} V_{ijm} + \dots + V_{12\dots(l-1)} \quad (9)$$

where: J – analysed model output (objective function defined by Eq. 2); $V(J)$ – the total variance of the analysed model output; i, j, m – model parameters; l – number of model parameters; V_i – the model output variance explained by influence of the i^{th} parameter; V_{ij} – the model output variance explained by influence of the interactions between parameters i and j ; V_{ijm} – the model output variance explained by influence of the interactions between parameters i, j and m ; $V_{12\dots(l-1)}$ – the model output variance explained by influence of the interactions between parameters up to $l-1$ order. The decomposition is unique if the input factors are independent from each other.

The outcomes of sensitivity study by the Sobol method are usually presented as the first order, higher order and total order sensitivity measures. The first order sensitivity index (S_i) describes the influence of the parameter i on the model output.

$$S_i = \frac{V_i}{V(J)} \quad (10)$$

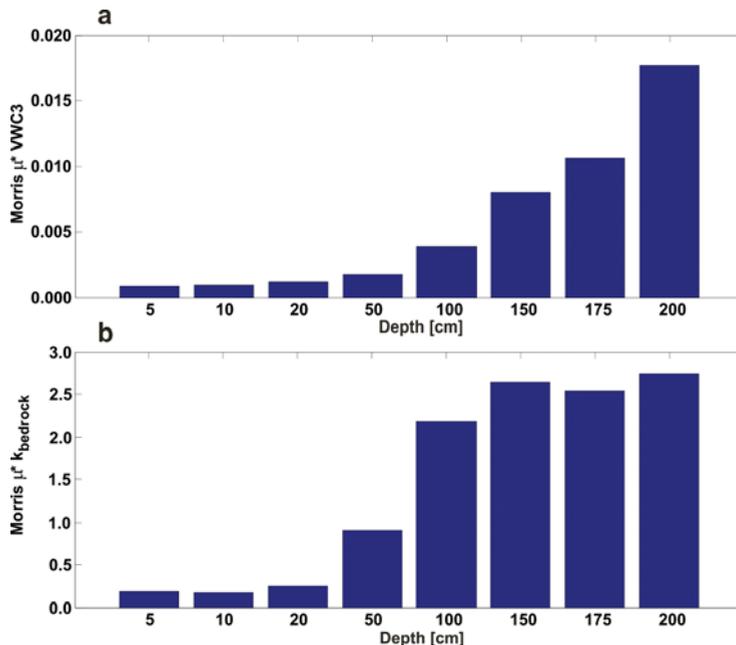


Fig. 5. The influence of the parameter VWC3 (a) and parameter k_{bedrock} (b) on model fit (NS) at different depths.

Higher order sensitivity measures describe the influence of interactions between model parameters on model output. For example, the second order sensitivity measure, between the parameters i and j can be written as:

$$S_{ij} = \frac{V_{ij}}{V(J)} \quad (11)$$

The model sensitivity to the interactions among subsets of parameters, so-called higher order effects, is investigated using the Sobol total sensitivity measure: S_{Ti} . This measure represents all interactions which involve parameter i .

In this study, the total sensitivity measure was approximated using numerical integration in a Monte Carlo framework by the so-called ‘freezing unessential variables’ method (Sobol 2001; Saltelli 2002; Saltelli *et al.* 2008). According to Homma and Saltelli (1996), the approximation method gives accurate and informative results at reasonable computational cost.

The Sobol first and total order sensitivity indices were calculated. Six CryoGrid 2 model parameters (except Q) were sampled separately using a quasi-random Sobol sequences from uniform distribution within the ranges presented in Table 1.

The outcomes of Sobol methods (Table 3) provide a similar ranking of model parameters to Morris method. The parameters VWC1, k_{bedrock} and k_{snow} have significant influence on the model output. The other parameters are unimportant

Table 2

Description of the CryoGrid 2 model parameters.

Nr	Parameter	Units	Abbreviation	optimum	constant
1	Volumetric water content 1	[%]	VWC1	1.18	-
2	Volumetric water content 2	[%]	VWC2	-	5.00
3	Volumetric water content 3	[%]	VWC3	-	5.00
4	Heat flux at lower boundary	[W/m ²]	Q	-	0.058
5	Density of the snow	[kg/m ³]	ρ_{snow}	-	330
6	Thermal conductivity of the snow	[W/mK]	k_{snow}	0.40	-
7	Thermal conductivity of the mineral component of the soil	[W/mK]	k_{bedrock}	1.18	-

Table 3

The results of sensitivity analysis by Sobol method.

	VWC1	VWC2	VWC3	ρ_{snow}	k_{snow}	k_{bedrock}
S_i	0.4520	0.0000	0.0000	0.0000	0.1460	0.2960
S_{Ti}	0.5624	0.0000	0.0000	0.0000	0.1858	0.3992

for the appropriate model calibration using the analysed model output. These findings are also presented in Fig. 6, where six panels present the achieved values of goodness of fit at eight levels when the model parameters are varying. The result of each of 7000 simulations is shown as dot. Just in the case of three parameters ($VWC1$, $k_{bedrock}$ and k_{snow}) it can be seen that different values of these parameters lead to different values of tested model output. Also the ranges of parameter values that give the highest values of model output could be selected. In the case of other parameters, the change of parameters values does not influence the model output.

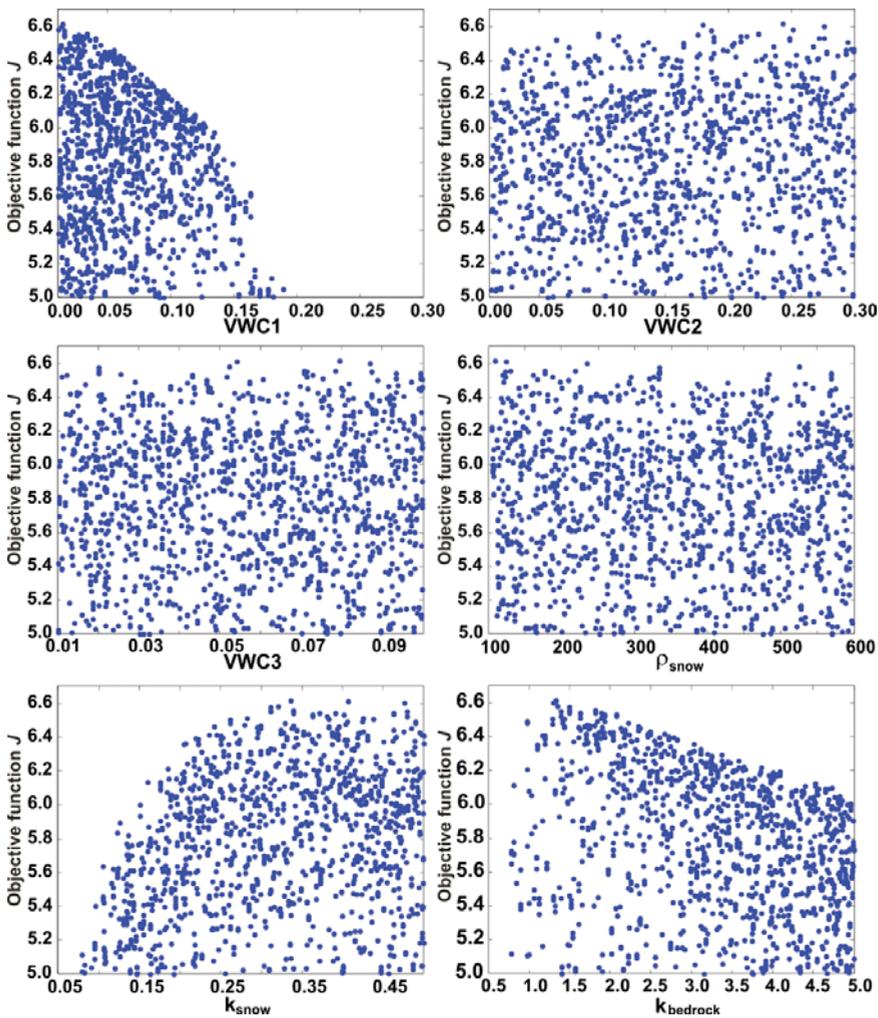


Fig. 6. The simulation results of the sensitivity analysis by Sobol method for the objective function $J > 5$. Each dot represents outcome of one simulation.

Model calibration. — Taking into account the results of sensitivity analysis, only three parameters of CryoGrid 2 model were calibrated, namely VWC1, k_{snow} and k_{bedrock} . As mentioned before, for the final calibration we used ground temperature data measured at the borehole in the Fuglebekken catchment. The data set consists of temperatures measured by eight sensors at depths: 5, 10, 20, 50, 100, 150, 175 and 200 cm in the period from the 5th of September 2013 to the 26th of June 2014. The model was calibrated using Differential Evolution with Global and Local neighbours (DEGL) (Das *et al.* 2009; Piotrowski and Napiórkowski 2012).

As the objective function defined by Eq. 2, a sum of Nash–Sutcliffe (*NS*) coefficients estimated for the different layers was applied. The optimum of the objective function is 6.56. For each analyzed depth, the optimum solution, *i.e.* the *NS* values, are presented in Table 4. These values vary slightly – from 0.7520 at 20 cm to 0.9072 at 200 cm.

The optimal parameters values represent simplified modelled conditions and are listed in the Table 2. According to our calculation the volumetric water content in the first layer is very small. This value represents dry conditions throughout the year, which are more complex in the environmental systems than in the model. The calibrated thermal conductivity of the snow amounts to 0.40 [W/mK], while the thermal conductivity of the mineral component of the soil is 1.18 [W/mK]. These values correspond well to the values presented by Westermann *et al.* (2013).

The results of model calibration are presented in Fig. 7. Each panel presents the comparison of observed (blue lines) and simulated (red lines) ground temperatures at analyzed depths. As can be seen the results of the model fit closely to observed data. In layers closer to the ground surface the simulated temperatures are highly dependent on forcing temperatures and the daily ampli-

Table 4

Calibration and validation results.

nr	Depth [cm]	Calibration (2013–2014) Nash–Sutcliffe (<i>NS</i>)	Validation (1990–2014) Nash–Sutcliffe (<i>NS</i>)
1	5	0.7580	0.9036
2	10	0.7735	0.9076
3	20	0.7520	0.8968
4	50	0.8009	0.8940
5	100	0.8272	0.8968
6	150	0.8467	–
7	175	0.8955	–
8	200	0.9072	–

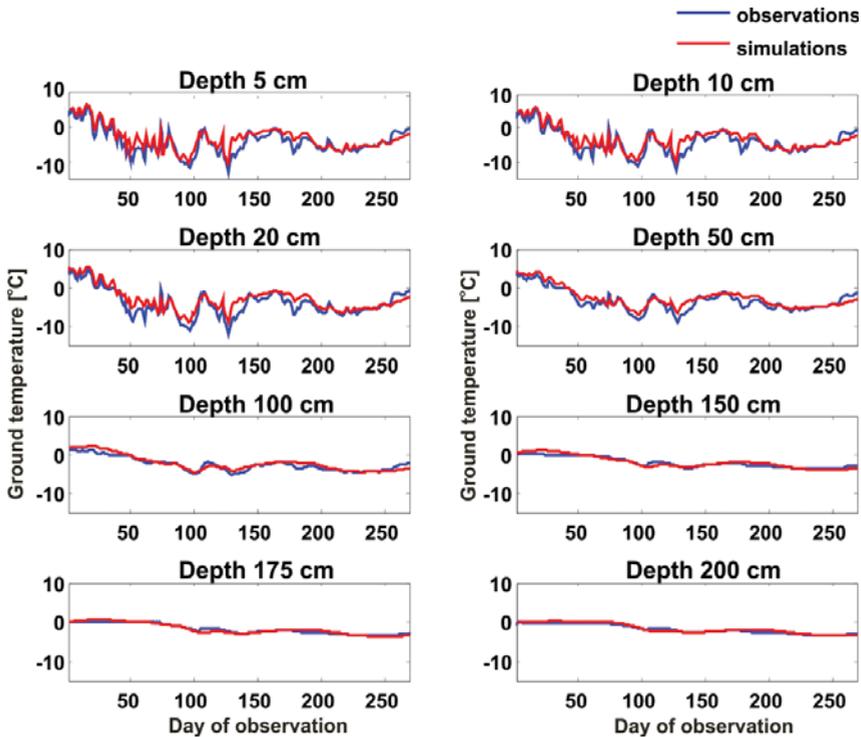


Fig. 7. The results of CryoGRID 2 model calibration. Comparison of observed and simulated ground temperatures in Hornsund at eight depths. X axis denotes days in the period September 2013 – June 2014, Y axis ground temperatures.

tude can be clearly seen. Going deeper with the profile, the reaction to forcing temperatures from above gets lower and the amplitude gets lower accordingly. The zero curtain effect is presumed approximately by CryoGrid 2.

Model validation. — To verify the model, simulated and observed data of ground temperatures at the Hornsund Station in the years 1990–2014 are compared. The results of the model validation presented as Nash–Sutcliffe values are shown in Table 3. The Nash–Sutcliffe values vary from 0.8940 for 50 cm up to 0.9076 at 10 cm, which means that very good results were achieved.

Fig. 8 displays a comparison of observed and simulated ground temperatures at 10 and 100 cm. Each blue dot is a result of single day comparison. Note the different scales of these two figures, resulting from smaller temperature amplitude at higher depths. For 10 cm depth, the model fit gets worse for the extreme values. It is due to the scope of variability of the thermal regime. In most cases, the model works better at higher depths.

The results agree particularly well with observed mean conditions of ground temperatures. The curve fits of calibrated values of ground temperatures fit less

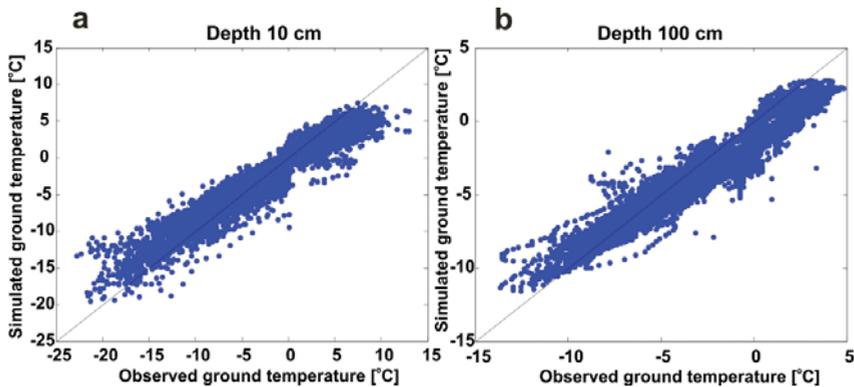


Fig. 8. The results of CryoGRID 2 model validation. Comparison of observed and simulated ground temperatures in Hornsund at 10 and 100 cm in years 1990–2014.

accurately to the extremes, in particular close to the surface. During the summer months, temperatures are slightly underestimated. We suppose that is due to the changing values of ground water content in the ablation period, when the active layer appears.

Charts with full data series would be unclear for readers, therefore we present the results for the last five years 2010–2014 in Fig. 9. On the left side of the figure, there is an example of simulation in the upper layer (close to the ground surface) at 10 cm. The right side presents ground temperatures at 100 cm. Generally in both cases, simulations fit quite well to the measurements. The best match is estimated for the cold seasons. The highest differences between simulations and observations occur for the ablation seasons, for which the temperatures are underestimated. We assume that those errors are due to differences of the water content in the active layer. When it appears, unfrozen ground is permeable and allows water to infiltrate. The volumetric water content is higher than during cold season, which changes the thermal regime of the ground. In CryoGrid 2 it is assumed that the volumetric water content is constant throughout the year, while in reality it changes when the ground thaws.

Inter-daily fluctuations are visible at small depths from 5 to 50 cm, but diminish with depth, so that the fit is better further down from the ground surface. The closer to the surface, the more meteorological elements influence the thermal regime (*e.g.* total sunshine, cloud cover, wind action, rain) resulting in higher variability (Harris *et al.* 2009; Westermann *et al.* 2011).

Simulation results. — The satisfactory results of the model calibration and validation presented in the previous sections allow us to estimate the magnitude of changes of the ground thermal regime for the period 1990–2014. In Fig. 10, we present simulated subsurface temperature–depth profile in the vicinity of Polish

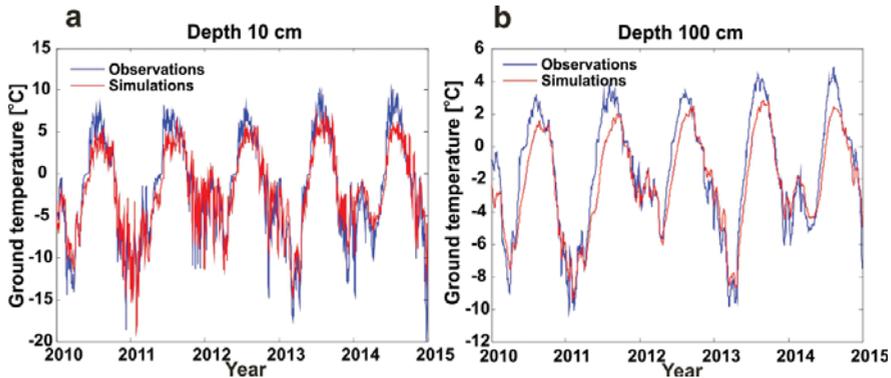


Fig. 9. The results of CryoGRID 2 model validation. Comparison of observed and simulated time series of the ground temperatures in Hornsund at (a) 10 cm and (b) 100 cm in years 2010–2014.

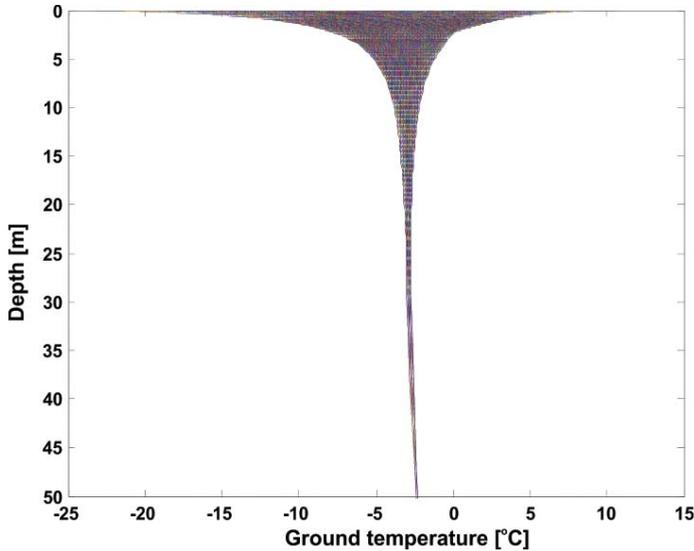


Fig. 10. Modelled temperature-depth profile at Polish Polar Station in Hornsund.

Polar Station in Hornsund. The temperature trumpet consists of modelled results from each day of the period 1990–2014. The shape of this trumpet curve is the result of seasonal and multiannual amplitude of temperatures and model parameters. The results are comparable to those obtained from the deep, instrumented boreholes in other sites of Spitsbergen (Christiansen *et al.* 2010). In general, three zones of temperature variability can be distinguished as a result of simulations. The near surface thermal response to thermal anomalies is clearly evident in Hornsund. According to the model results, the multiannual variability reaches 25 m, while at lower depths is not significant (less than 0.1°C) in the analyzed period of time.

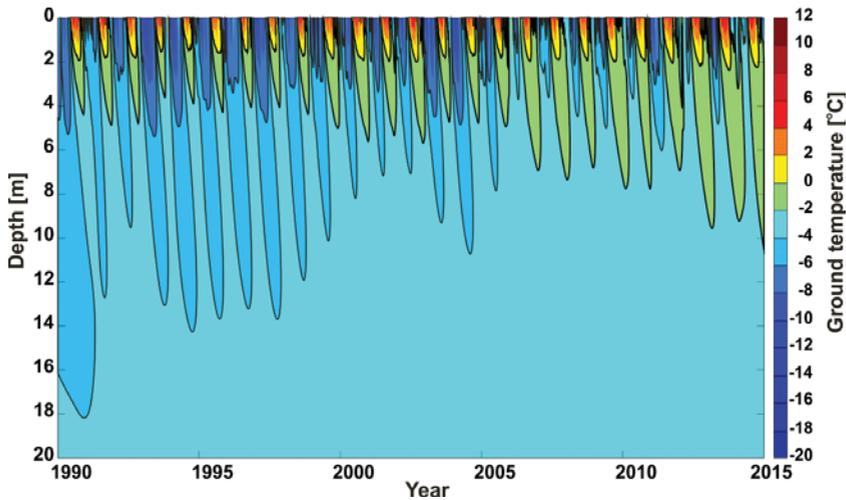


Fig. 11. The distribution and evolution of subsurface temperatures in Hornsund in the period 1990–2014.

The evolution of ground temperatures in the uppermost 20 m is presented in Fig. 11. Navy blue and green colors represent temperatures below zero ($^{\circ}\text{C}$), while yellow and red temperatures above zero. The upper layer, down to 18 m is characterized by seasonal changes of the ground temperatures in the analyzed period of time 1990–2014.

The simulated ground thermal regime shows an increase of the ground temperatures with time. In the first decade (1990–2000) winters were colder and lower temperatures could influence the ground. In the next period of time, the temperatures of the ground were higher. In the last decade, winter cooling effects were not as strong as in the 1990s. These differences of ground temperatures correspond to an increasing trend of air temperatures observed in Hornsund and general warming across Svalbard.

Conclusions

This work presents the results of permafrost modelling in Hornsund area using the CryoGrid 2 model. Probably it is the first attempt to describe the permafrost thermal conditions in deeper soil layers in southern Spitsbergen.

We started with the estimation of initial conditions. In second step, a sensitivity analysis of the model parameters by Morris method was performed. The model was calibrated with measured ground temperatures in a newly established borehole in the Fuglebekken catchment, with temperature sensors at eight depths: 5, 10, 20, 50, 100, 150, 175 and 200 cm. The results of the calibration process

were satisfactory, with Nash–Sutcliffe (*NS*) values varying from 0.7520 at 20 cm to 0.9072 at 200 cm. The model was validated with the ground temperature data from the period 1990–2014 from five depths: 5, 10, 20, 50, 100 cm measured routinely at Hornsund station. At the validation stage, we also received a satisfactory match between simulations and observations. *NS* values varied from 0.8940 for 50 cm up to 0.9076 at 10 cm. The seasonal variations were well reproduced in the fit of the observed and simulated time series. During the summer months, the ground temperatures are slightly underestimated, probably as the results of changing ground water content in the active layer. Taking into account errors of the assessment in the ablation period, we did not describe the evolution of the active layer thickness.

The results of modelling allowed us to estimate the subsurface temperature trumpet and the evolution of the ground thermal regime. We also estimated the thickness of permafrost in the surroundings of Polish Polar Station in Hornsund to more than 100 m, but a large margin of error must be assumed due to shortcomings of the method. These estimates are strongly influenced by assumed initial conditions and values of parameters. We also estimated uncertainty of depth of permafrost by analysis of geothermal gradient for Monte Carlo analysis. In that case, the median permafrost depth equal to 104 m was estimated together with standard deviation 28 m. These analysis indicate that the estimated depth of permafrost in the Hornsund area vary about 50% due to small variation of model parameter values. Only in situ measurements in deep borehole may provide the true permafrost thickness. The outcome of this paper agrees well with other studies carried out on Svalbard.

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