

SNOW AND GLACIERS

THERMAL RESISTANCE OF SNOW: VARIATIONS IN SPACE AND TIME

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Space and time variations in the thermal resistance of snow have been studied in several areas of the Krasnoyarsk region. The depth of snow has been correlated with its thermal resistance and with frost depth depending on the properties of snow and soils. The dependence of thermal resistance on snow stratigraphy, with different snow grain sizes, ice crusts, and depth hoar has been illustrated by examples from Spitsbergen and Moscow area. Neglect of snow stratigraphy can lead to 1.5 times underestimation of thermal resistance.

*Depth hoar, frost depth, snow stratigraphy, snow thermal resistance*

INTRODUCTION

Snow is an important element of interactions in the system 'atmosphere–lithosphere', which affects the thermal state, formation conditions, and behavior of permafrost [Osokin *et al.*, 2000b, 2006]. Heat insulation properties of snow depend on its thermal resistance found as the ratio of snow depth to thermal conductivity. Thermal conductivity of snow (meant hereafter as its effective value), in turn, is estimated from its average density but may change as a result of variations in snow texture and structure induced by natural (weather) or man-caused factors.

This study focuses on space and time variations of thermal resistance and its dependence on snow stratigraphy.

INFLUENCE OF SNOW ON FROST DEPTH

Man-induced or natural changes to the conditions of snow accumulation in permafrost areas may trigger natural hazards and worsen the bearing capacity of foundations. Man-made redistribution and compaction of snow in cities and villages changes ground temperatures and seasonal thaw depth (active layer thickness). Namely, field observations show that compaction and removal of snow as part of engineering ground preparation (pioneering) for two or three years produces up to 2 m thick lenses of frozen soil [Maksimova *et al.*, 1977]. On the other hand, laying lawns with grass and shrubs maintains accumulation of soft snow and reduces considerably the depth to which the ground freezes beneath them in cold seasons.

The snow depth within small areas can vary strongly as a result of natural and man-made effects, while local air temperatures are commonly more sta-

ble. Heat flux changes in the system 'atmosphere–near-surface air–soil' due to warming can give rise to discontinuous permafrost and growth of taliks leading eventually to permafrost degradation. In some permafrost areas, the climatic temperature changes are compensated by changes in snow parameters [Izrael *et al.*, 2002]: while mean annual air temperatures become warmer, ground temperatures may become colder because of reduced depth and slower accumulation of snow in the first half of winter [Pavlov, 2008].

According to simulations, the depth to which the ground freezes in the cold season (frost depth) may become times larger or smaller as a result of variations in thermal properties of snow [Osokin *et al.*, 2001, 2007]. They are the effective thermal conductivity ( $\lambda_{sn}$ ) and thermal resistance ( $R_{sn} = h_{sn}/\lambda_{sn}$ ), and the related insulation capacity, which are defined by two basic characteristics: snow depth ( $h_{sn}$ ) and density ( $\rho_{sn}$ ).

The effect of thermal resistance on ground freezing patterns can be estimated from frost depth curves obtained at different values of snow parameters and air temperature, with the frost depths measured at different weather stations normalized to a fixed air temperature.

DATA

The model frost depth curves were obtained by forward modeling [Osokin *et al.*, 2000a, 2001] with soil parameters according to the Building Code (SNiP 2.02.04-88) [1997], for 1400 kg/m<sup>3</sup> clay silt. The mean-latitude values of incoming solar radiation to the horizontal surface in plainland Russia used in the calculations were borrowed from [Pivovarova, 1977]. The albedo at negative air temperatures was assumed

to be 0.8 in the presence of snow, 0.5 in the snow melting season, and 0.2 in the absence of snow.

The calculations were performed for the conditions of the Baikit weather station in the Krasnoyarsk region (61°40' N, 96°22' E), with the WMO code 23891 (WMO is the World Meteorological Organization). Daily air temperatures for periods of positive and negative temperatures were approximated as

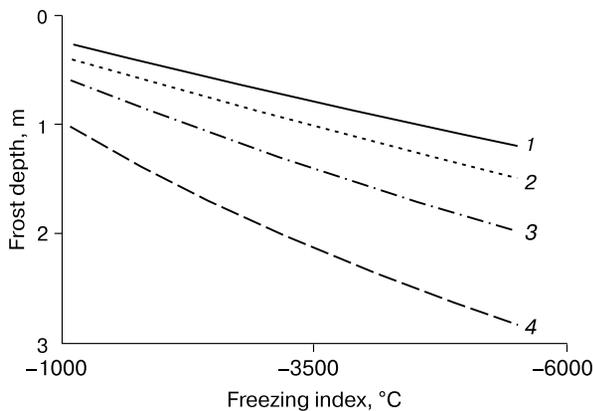
$$T_a = T_1 \sin(\pi\tau/\tau_{\max}) + 273,$$

where  $\tau$  is the current time;  $\tau_{\max}$  is the duration of the respective period;  $T_1 = T_{\max}$  or  $T_{\min}$ , while  $T_{\max} = \pi T_{th}/2$  and  $T_{\min} = \pi T_f/2$ ,  $T_{th}$  and  $T_f$  being the mean values for the periods of positive (thaw) and negative (frost) air temperatures, respectively. The Baikit site has the following parameters: freezing index (a total of negative air temperatures)  $\Sigma T_a = -4020$  °C;  $T_f = -19.0$ ,  $T_{th} = 10.3$  °C; length of cold season 220 days; the lag of the snow accumulation onset behind the onset of negative daily mean temperatures ( $\tau_{s0}$ ) is 8 days; the snowfall temperature equals the air temperature; wind speed is 7 m/s; air humidity is 70 %; cloud cover is 0.6.

The values of snow depth and density at the site were taken from the Climate reference book [Borisenkova, 1990]. The water content in unfrozen soil and the unfrozen water content at the frost/thaw boundary in clay silt were assumed to be, respectively, 25 % and 11 %. The ground temperature pattern by the onset of freezing, obtained by preliminary calculations, was approximated by a parabola with the maximum 2 °C at the depth 2.5 m, and 0 °C at 0 and 5 m below the surface (1.3 °C on average).

## RESULTS

Calculations for the frost depth as a function of freezing index, at different snow depths (snow densi-



**Fig. 1. Predicted frost depth as a function of freezing index at different thermal resistances and depths of snow of the same density:**

1 – 6.64 m<sup>2</sup>·K/W (1.0 m); 2 – 4.93 (0.75); 3 – 3.29 (0.50); 4 – 1.64 m<sup>2</sup>·K/W (0.25 m).

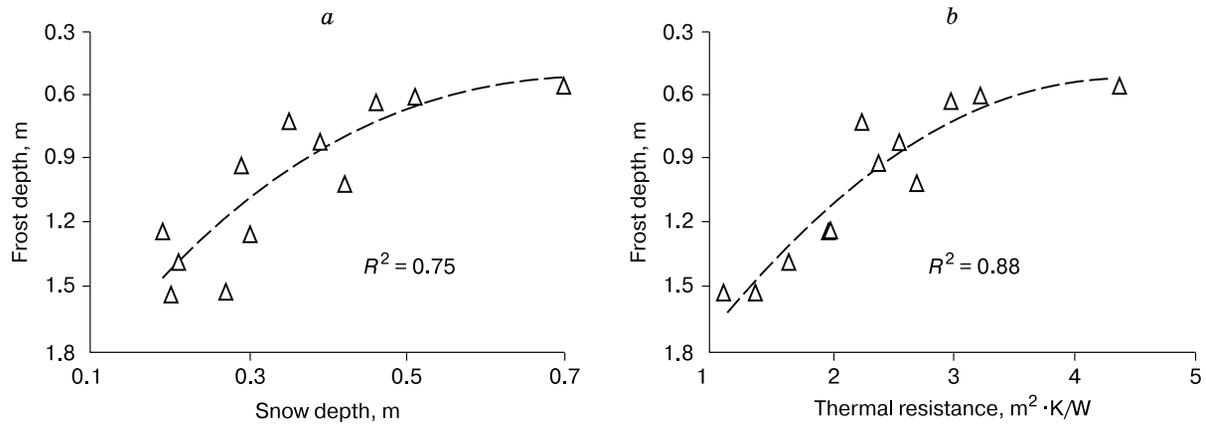
ty being constant) and thermal resistances (Fig. 1) predict that the frost depth becomes 2.2 times greater when the freezing index increases by a factor of 3.3 (from -1000 to -3300 °C), at the snow depth 0.5 m and the thermal resistance 3.29 m<sup>2</sup>·K/W. At three times lower thermal resistance (1.64 m<sup>2</sup>·K/W instead of 4.93 m<sup>2</sup>·K/W) and snow depth (0.25 m instead of 0.75 m), this frost depth increase remains almost the same, about two times (Fig. 1).

Field evidence on the influence of snow cover and its thermal resistance on ground freezing in different areas were obtained as follows. First, normal (long-term average) values of frost depth ( $\xi$ , m), freezing index, and snow depth and density [Osokin et al., 2013] were estimated from agroclimatic reference data [Bakhtin, 1961]. The norm of freezing index for the weather stations in the Krasnoyarsk region used for reference was found to be about -2200 °C. The frost depths were normalized to this value using the curves of Fig. 1, and the normalized values ( $\xi_T$ ) were plotted as a function of snow depth and thermal resistance (Fig. 2). Note that the normalized frost depths showed a better correlation with thermal resistance than with snow depth ( $R^2 = 0.88$  against  $R^2 = 0.75$ ).

The model and field frost depths as a function of snow thermal resistance are compared in Fig. 3. The model values were calculated using the above parameters, at soil water contents 25 and 35 %, and at different snow depths, and were normalized to  $\Sigma T_a = -2200$  °C; the field values were obtained at weather stations in the Komi Republic and the Krasnoyarsk region. The thermal resistance of snow is generally lower at weather stations in Komi than in the Krasnoyarsk region (Fig. 3), while the empirical frost depths for both regions are within the predicted range at 25 and 35 % water contents.

The frost depths, and their variations in field data (Fig. 3), have also other controls besides the thermal resistance and depth of snow. Calculations with the model parameters as above and with reference to data from the Baikit weather station show that frost becomes 0.02 m deeper as the lag of the snow accumulation onset behind the onset of negative daily mean temperatures ( $\tau_{s0}$ ) becomes one day longer; as this difference increases to 5 days, the frost depth increase reaches 0.1 m. On the other hand, the frost depth becomes 0.13 m shallower as the mean ground temperature by the time of frost onset rises for 1 °C in response to air temperature warming. Note that the daily means of positive air temperatures for the weather stations of Komi and Krasnoyarsk areas range within 8–10 and 11–14 °C, respectively. As these values grow from 10 to 14 °C, the respective ground temperature mean increases from 1.9 to 3.2 °C, which reduces the frost depth from 1.66 to 1.48 m (11 %).

In silt (instead of clay silt), the thermal conductivity of unfrozen and frozen soil parts becomes 20 % and 12 % higher, respectively, while the frost depth

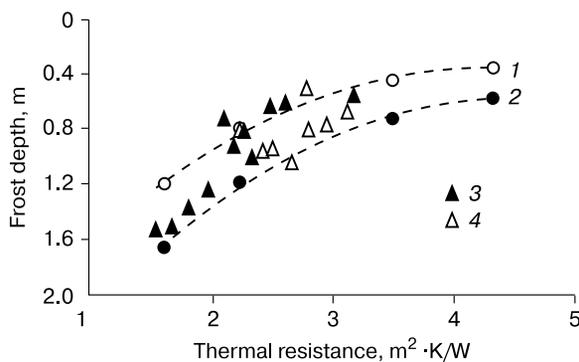


**Fig. 2. Normalized frost depth as a function of snow depth (a) and thermal resistance (b), for weather stations of Krasnoyarsk region.**

Dash line shows the trend.

increase is as low as 2 %. The change in clay silt density from 1400 to 1600 kg/m<sup>3</sup> decreases the frost depth for 10 % at the account of greater volumetric percentage of water, though its weight percentage remains the same.

Vegetation (moss and plants) is another factor that can influence frost depth by increasing thermal resistance. The thermal conductivity of some moss species reported in [Tishkov *et al.*, 2013] ranges within 0.15–0.20 W/(m·K) and corresponds to that of snow over a large part of permafrost [Balobaev, 1991]. Therefore, moss slows down ground warming in the warm season, but it does not prevent ground from freezing in the cold season. The reason is that the thermal conductivity of moss becomes 2–3 times greater in winter, specifically because the thermal conductivity of ice is almost 4 times that of water [Osokin *et al.*, 2011; Osokin and Sosnovskiy, 2012].



**Fig. 3. Predicted and measured normalized frost depth as a function of snow thermal resistance.**

1, 2 – predicted values, for soil water contents 35 and 25 %, respectively; 3, 4 – measured field data from Krasnoyarsk region and Komi Republic.

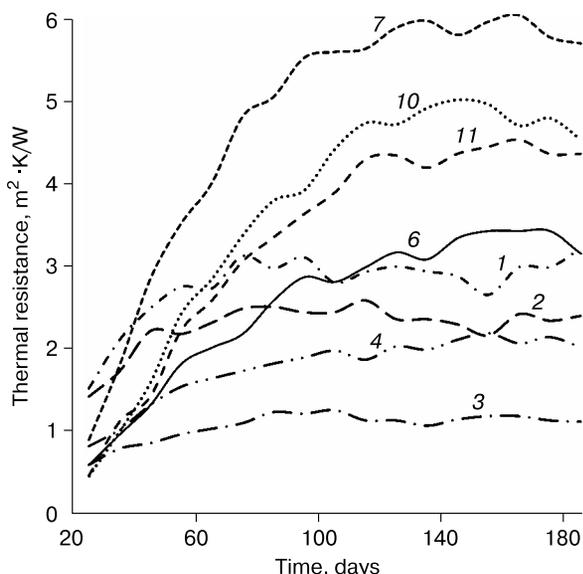
Note that heat capacity of moss is commensurate to that of ice, about 2 kJ/(kg·K).

#### SPACE AND TIME PATTERNS OF SNOW THERMAL RESISTANCE

The spatial patterns of snow thermal resistance were obtained by comparing its values at different weather stations of the Krasnoyarsk region. Thermal resistance was estimated with reference to published values of snow depth and density. The latter is an important parameter that controls both the depth and thermal conductivity of snow: e.g., as snow density increases from 200 to 300 kg/m<sup>3</sup>, its depth reduces by a factor of 1.5 and thermal conductivity becomes 1.9 times greater; correspondingly, thermal resistance decreases by a factor of 2.8.

Figure 4 shows variations in snow thermal resistance at weather stations in the northern Krasnoyarsk region, with maximum snow depths from 22 cm (Norilsk) to 75 cm (Baikit) and densities from 170 to 260 kg/m<sup>3</sup>. The weather stations are numbered according to their numbers in the Climate reference book [Borisenkov, 1990]. The thermal resistances of snow at the stations Agata (WMO code 23383), Ust'-Kamo (23992), Vanavara (24908), and Igarka (23274), numbered as 5, 8, 9, and 12, respectively, are within the ranges for the stations Taimba (no code) and Chemdalsk (30014), numbers 10 and 11, respectively (Fig. 4).

As the depth of snow increases, its density and, correspondingly, thermal conductivity increase as well [Balobaev, 1991]. As a result, the ratios of these parameters (thermal resistance of snow) do not change much with time over the greatest part of the year (Fig. 4). Thermal resistance characterizes insulation properties of snow at a given geographic point. However, weather changes may cause changes to



**Fig. 4. Thermal resistance of snow from weather records of Krasnoyarsk region.**

1 – Volochanka (WMO code 20982); 2 – Dudinka (23074); 3 – Norilsk (23078); 4 – Essei (24105); 6 – Tura (24507); 7 – Baikit (23891); 10 – Taimba (no code); 11 – Chemdalsk (WMO code 30014). Weather stations are coded according to Climate reference book [Borisenkov, 1990].

stratigraphy of snow, which will influence its thermal properties: warming and cooling excursions produce, respectively, crusts and layers of ice inside snow cover or depth hoar.

### WEATHER CONTROL OF SNOW STRATIGRAPHY

The weather control of snow stratigraphy can be considered with the example of the Barentsburg site in the western Svalbard archipelago (Spitsbergen Island). The snow cover in the Svalbard area has a layered structure with ice crusts produced by winds, partial thaw, and rainfall in the cold season.

The snow structure is additionally affected by cooling excursions which increase the temperature gradient and mass transport rate across the snow cover and thus give rise to depth hoar. The morphology of snow crystals can change under the effect of temperature gradient, while the latter has bearing on the evaporation-condensation rate ( $v$ , mm/day) [Yoshida, 1963]:

$$v = 0.24 \text{ grad } T,$$

where  $\text{grad } T$  is the temperature gradient, °C/cm. At the gradient 20 °C per 0.5 m,  $v \approx 0.1$  mm/day.

High gradients of vapor concentration across the snow depth, which arise upon 10–20 °C cooling for 3–4 days, accelerate softening of snow to a depth controlled by the mean winter air temperature and by the

frequency of temperature oscillations. Softened snow layers form in 6 to 30 days if the gradient exceeds 20 °C/m, at the mean air temperature –7 to –12 °C [Golubev et al., 2010]. The conditions for depth hoar formation are: negative snow temperature and temperature gradient above 20 °C/m holding continuously for more than seven days [Shmakin et al., 2009].

Snow surveys in southern Moscow area showed that depth hoar occupied more than 65 % of snow cover in 2011 [Chernov, 2013], and the percentage was still higher in 2012.

### EFFECT OF SNOW STRATIGRAPHY ON THERMAL RESISTANCE

Ice crusts formed by frequent winter thaw episodes in Spitsbergen (see above) have quite a low thermal resistance but contribute considerably to average snow depth. Therefore, thermal resistance estimated without due regard for snow stratigraphy may be 1.5 times underestimated. In addition to the ice crusts, thermal resistance of snow depends on its vertical density variations. The effect of snow stratigraphy on thermal resistance was assessed using snow density and depth profiles (Fig. 5) obtained at different elevations above sea level (asl) at the Barentsburg weather station (Spitsbergen). The stratigraphy shown by measurements in trench 1 and the respective thermal resistance values are listed in Table 1. Effective thermal conductivity ( $\lambda_{\text{sn}}$ ) was calculated using a regression equation based on 20 published empirical relationships [Osokin et al., 1999]. The estimates have been obtained as average values for each density at every 10 kg/m<sup>3</sup>, the resulting curve being approximated by

$$\lambda_{\text{sn}} = 9.165 \cdot 10^{-2} - 3.814 \cdot 10^{-4} \rho_{\text{sn}} + 2.905 \cdot 10^{-6} \rho_{\text{sn}}^2.$$

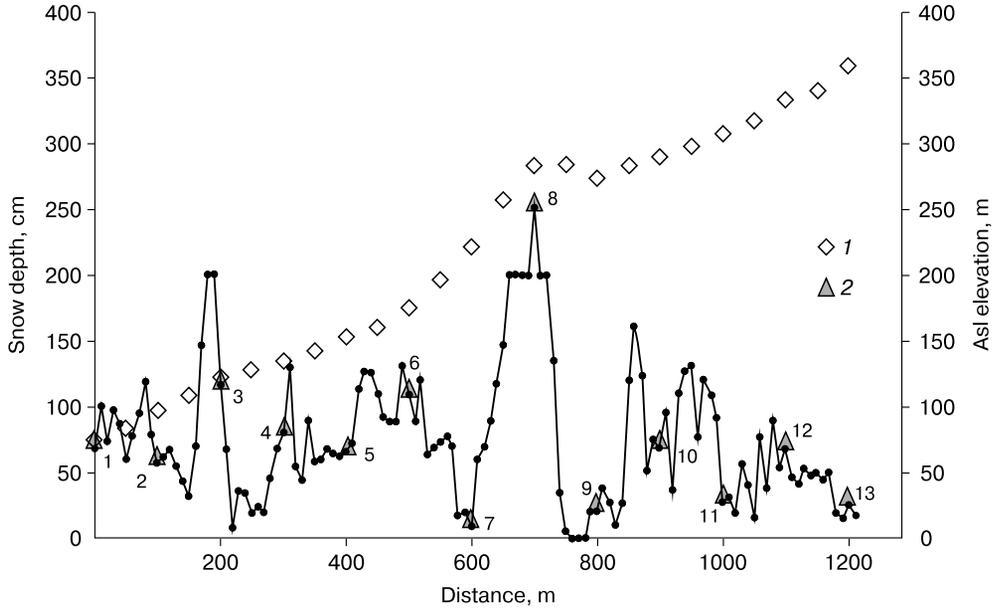
The thermal conductivities  $\lambda_{\text{sn}}$  calculated with this equation almost coincide with those obtained using the equation of Pavlov [1979] at a snow temperature –10 °C.

Knowing that thermal resistance of a layered medium is the sum of layer values, the total (actual) thermal resistance of snow was found as a sum of  $R_i$  for each layer:  $R_{\text{sn}} = \sum R_i$ . Correspondingly, the actual thermal conductivity of snow  $\lambda_{\text{sn}}$  will be

$$\lambda_{\text{sn}} = h_{\text{sn}} / \sum_i \frac{h_i}{\lambda_i},$$

where  $\lambda_i$  is the thermal conductivity of the  $i$ -th snow layer;  $h_{\text{sn}}$  and  $h_i$  are the total depth (thickness) of snow and the thickness of the  $i$ -th layer, respectively.

Average thermal resistances of snow ( $R^{\text{av}}$ ) found from its average density are compared with the actual ( $R^{\text{ac}}$ ) values for some trenches cut at different asl elevations (Fig. 5) in Table 2. The  $R^{\text{av}}$  and  $R^{\text{ac}}$  values, and their ratio ( $K_R = R^{\text{av}}/R^{\text{ac}}$ ) show that the average



**Fig. 5. Snow depths at different asl elevations and in different trenches near Barentsburg (Svalbard archipelago, Spitsbergen).**

1 – asl elevation; 2 – trench number (1–13).

is lower than the actual thermal resistance based on snow stratigraphy (layer values). The average and maximum differences of  $R^{av}$  and  $R^{ac}$  are 13 and 25 %, respectively, and reached 29 % in 2012 (at  $R^{av} = 5.4 \text{ m}^2\cdot\text{K}/\text{W}$  and  $R^{ac} = 7.6 \text{ m}^2\cdot\text{K}/\text{W}$ ), according to our measurements.

Snow density variations in the Moscow area are relatively small, but the actual thermal resistance ex-

ceeds that derived from average density neglecting the snow stratigraphy because of depth hoar present over the greatest part of the snow cover [Chernov, 2013]. Thermal conductivity of depth hoar measured near Moscow (200–400  $\text{kg}/\text{m}^3$ ) turned out to be 1.6–1.8 times lower than in granular snow [Chernov, 2013]. The ratios  $K_R$  found assuming the thermal conductivity of depth hoar to be 2 (1.5) times as low as in granular snow were 0.71 (0.89) in 2007, 0.68 (0.80) in 2011, and 0.56 (0.71) in 2012. Thus, the actual (true) thermal resistance of snow obtained with due regard for its stratigraphy can be much greater than the values based on average density.

**Table 1. Snow stratigraphy (trench 1), Barentsburg site**

Snow layers	$h_{sn}$ , m	$\rho_{sn}$ , $\text{kg}/\text{m}^3$	$\lambda_{sn}$ , $\text{W}/(\text{m}\cdot^\circ\text{C})$	$R_{sn}$ , $\text{m}^2\cdot^\circ\text{C}/\text{W}$
Soil				
Ice crust	0.010	900	2.10	0.005
Fine soft snow	0.085	260	0.19	0.450
Ice crust	0.020	900	2.10	0.010
Medium dense snow	0.025	420	0.44	0.056
Ice crust	0.005	900	2.10	0.002
Coarse dense snow	0.070	420	0.44	0.158
Coarse soft snow	0.020	260	0.19	0.106
Ice crust	0.005	900	2.10	0.002
Coarse dense snow	0.150	320	0.27	0.562
Ice crust	0.005	900	2.10	0.002
Coarse dense snow	0.150	410	0.42	0.354
Ice crust	0.025	900	2.10	0.012
Moss				
Average	0.57	350–430		

**Table 2. Thermal resistance of snow**

Trench number	$h_{sn}$ , m	Actual $R_{sn}^{ac}$ , $\text{m}^2\cdot^\circ\text{C}/\text{W}$	Average			$K_R$
			$\rho_{sn}$ , $\text{kg}/\text{m}^3$	$\lambda_{sn}$ , $\text{W}/(\text{m}\cdot^\circ\text{C})$	$R_{sn}^{av}$ , $\text{m}^2\cdot^\circ\text{C}/\text{W}$	
1	0.57	1.72	421	0.44	1.28	0.75
2	0.55	1.65	395	0.39	1.40	0.85
3	1.16	4.79	333	0.29	4.03	0.84
4	0.675	1.73	424	0.45	1.49	0.86
5	0.87	2.28	404	0.41	2.12	0.93
6	0.88	3.65	309	0.25	3.51	0.96
9	0.67	2.13	383	0.37	1.80	0.85
10	0.28	0.86	372	0.35	0.78	0.91
11	0.54	2.38	321	0.27	2.01	0.84

## CONCLUSIONS

The thermal insulation properties of snow depend on its thermal resistance. Calculations show that decrease in snow thermal resistance and the corresponding increase of freezing index (a total of negative air temperatures) provide a nearly equal increase in frost depth (10 %).

Analysis of experimental data based on real snow density shows that seasonal frost depth better correlates with thermal resistance of snow than with its depth ( $R^2 = 0.88$  against  $R^2 = 0.75$ ).

The true thermal resistance estimated with regard to snow stratigraphy, including ice crusts and depth hoar, is much higher than that derived from average density neglecting the snow texture and structure. The presence of ice crusts, which have low thermal resistance, contributes more to the average density of snow than to its thermal resistance. Depth hoar has lower effective thermal conductivity, and hence higher thermal resistance, than coarse granular snow of the same density.

Thus, the stratigraphy of snow has to be taken into account in the estimation of its insulation properties not to overlook the onset of permafrost degradation and the ensuing hazardous decrease of soil strength.

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