



# **Evaluation of air-soil temperature relationships simulated by land surface models during winter across the permafrost region**

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1 **Abstract.** A realistic simulation of snow cover and its thermal properties are important for  
2 accurate modelling of permafrost. We analyze simulated relationships between air and near-  
3 surface (20 cm) soil temperatures in the Northern Hemisphere permafrost region during winter,  
4 with a particular focus on snow insulation effects in nine land surface models and compare  
5 them with observations from 268 Russian stations. There are large across-model differences ~~as~~  
6 ~~expressed by simulated differences~~ between near-surface soil and air temperatures, ( $\Delta T$ ), of 3 to  
7 14 K, in the gradients between soil and air temperatures (0.13 to 0.96 °C/°C), and in the  
8 relationship between  $\Delta T$  and snow depth. The observed relationship between  $\Delta T$  and snow  
9 depth can be used as a metric to evaluate the effects of each model's representation of snow  
10 insulation, and hence guide improvements to the model's conceptual structure and process  
11 parameterizations. Models with better performance apply multi-layer snow schemes and  
12 consider complex snow processes. Some models show poor performance in representing snow  
13 insulation due to underestimation of snow depth and/or overestimation of snow conductivity.  
14 Generally, models identified as most acceptable with respect to snow insulation simulate  
15 reasonable areas of near-surface permafrost (12-16 million km<sup>2</sup>). However, there is not a  
16 simple relationship between the quality of the snow insulation in the acceptable models and the  
17 simulated area of Northern Hemisphere near-surface permafrost, likely because several other  
18 factors such as differences in the treatment of soil organic matter, soil hydrology, surface  
19 energy calculations, and vegetation also provide important controls on simulated permafrost  
20 distribution.



4

## 1 1 Introduction

2 Present-day permafrost simulations by global climate models are limited and future  
 3 projections contain high, model-induced uncertainty (e.g., Slater and Lawrence, 2013; Koven  
 4 et al., 2013). Most of the model biases and across-model differences in simulating permafrost  
 5 area are due to biased simulations of the atmosphere (~~air~~ temperature, precipitation), land  
 6 (snow, soil temperature) and their coupling. In winter, the snow insulation effect is a key  
 7 process for the air-soil temperature coupling. Its strength depends on both the snow ~~quantity~~  
 8 (~~depth, areal coverage~~) and snow ~~quality~~ (density, conductivity) (see overview by Zhang,  
 9 2005). Many individual model studies have shown the strong impact of a few snow  
 10 parameters on soil temperature simulations (e.g., recently, Dutra et al., 2012; Gouttevin et al.,  
 11 2012; Essery et al., 2013; Wang et al., 2013; Jafarov et al., 2014). Most importantly,  
 12 consideration of wet snow metamorphism and snow compaction, improved snow thermal  
 13 conductivity and multi-layer snow schemes improve the simulation of snow dynamics and  
 14 soil temperature. Parameterizations that take into account snow compaction (e.g. related to  
 15 overburden pressure, thermal metamorphism and liquid water) work better than simpler  
 16 schemes such as an exponential increase of density with time (Dutra et al., 2010). The  
 17 influence of snow thermal conductivity on soil regime has been demonstrated by many model  
 18 studies (e.g., Bartlett et al., 2006; Saha et al., 2006; Vavrus, 2007; Nicolsky et al., 2007;  
 19 Dankers et al., 2011; Gouttevin et al., 2012). Winter soil temperature can change by up to 20  
 20 K simply by varying the snow thermal conductivity by 0.1-0.5 W m<sup>-1</sup> K<sup>-1</sup> (Cook et al., 2008).  
 21 The snow insulation effect also plays an important role for the Arctic soil temperature  
 22 response to climate change (e.g., reduced soil warming due to shallower snow versus  
 23 enhanced soil warming due to shorter snow season) (e.g., Lawrence and Slater, 2010), and  
 24 therefore for future near-surface permafrost thawing and soil carbon vulnerability (e.g.,  
 25 Schuur et al., 2008). The model skill in atmosphere-soil coupling with the concomitant snow  
 26 cover in the Arctic is an important factor in the assessment of limitations and uncertainty of  
 27 carbon mobility estimates (Schaefer et al., 2011).  
 28  
 29 The Snow Model Intercomparison Project (Snow MIP) (Essery et al., 2009) and the Project  
 30 for Intercomparison of Land-Surface Parameterization Schemes (PILPS) Phase 2e (Slater et  
 31 al., 2001) examined the snow simulations of an ensemble of land-surface schemes for the  
 32 mid-latitudes. However, until now there has been no attempt to evaluate the air-soil  
 33 temperature relationship in the Northern Hemisphere permafrost region and the detailed role



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1 of snow depth therein across an ensemble of models. In such an investigation, a first suitable  
 2 approach is the evaluation of stand-alone (off-line) land surface models (LSMs). The  
 3 retrospective (1960-2009) simulations from the model integration group of the Permafrost  
 4 Carbon Network ("PCN"; <http://www.permafrostcarbon.org>) provide an opportunity to  
 5 evaluate an ensemble of nine state-of-the-art LSMs. Here, the LSMs are run with observation-  
 6 based atmospheric forcing, meaning that snow depth is not influenced by biases in the  
 7 atmospheric forcing in a coupled model set-up. The evaluation of the offline modeled air  
 8 temperature - snow depth - near-surface soil temperature relationship in winter is therefore  
 9 important for revealing a model's skill in representing the effects of snow insulation.

10

11 Most of the LSMs participating in PCN are the land-surface modules of Earth System Models  
 12 (ESMs) participating in the Coupled Model Intercomparison Project (CMIP5; [http://cmip-  
 13 pcmdi.llnl.gov/cmip5/](http://cmip-pcmdi.llnl.gov/cmip5/)) although in some cases different versions were used for PCN and  
 14 CMIP5 simulations. Thus, the results we present can guide the corresponding evaluation of  
 15 these ESMs, though analysis of coupled model results requires consideration of couplings  
 16 between model components and is necessarily more complex.

17

18 The scope of the present study is to analyze the extent to which the ensemble of ~~PCN models~~  
 19 can reproduce the observed relationship between air and near-surface soil temperatures in the  
 20 Northern Hemisphere permafrost region during winter, with a particular focus on the snow  
 21 insulation effect. For the latter we analyze the impact of snow depth on the difference  
 22 between near-surface soil and air temperatures. Our related key questions are: How well do  
 23 the models represent the observed magnitude and spatial pattern of the snow depth control  
 24 over the air-soil temperature difference in winter? What is the range of these simulated  
 25 relationships across the model ensemble? To the extent possible, we try to relate the  
 26 performance level to the model's snow schemes. With this aim in mind, a simultaneous  
 27 analysis of simulated air and near-surface soil temperatures, and snow depth is presented and  
 28 compared with those from a novel set of Russian station observations. We focus here on a  
 29 comprehensive Russian station data set because this has been compiled within PCN, and it is  
 30 hard to find other station data sets which provide simultaneous observations of both air and  
 31 soil temperatures as well as snow depth over a long period.

32

33 In Sect. 2, we describe the model simulations, the station observations used for evaluation,  
 34 and the analysis methods. In Sect. 3, we present a detailed analysis of near-surface air



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1 temperature - snow depth - soil temperature relationships in winter. In Sect. 4, we discuss the  
 2 roles of atmospheric forcing and model processes. We summarize our findings and present  
 3 conclusions in Sect. 5.

## 4 **2 Data and Analysis**

### 5 **2.1 Models**

6 We use data from nine LSMs participating in the PCN, including CLM4.5, CoLM, ISBA,  
 7 JULES, LPJ-GUESS, MIROC-ESM, ORCHIDEE, UVic, and UW-VIC. For detailed  
 8 information about the models and simulations we refer to Rawlins et al. (2015) and Peng et al.  
 9 (2015). The total soil depth for soil thermal calculations ranges from 3 m (divided in 8 layers)  
 10 in LPJ-GUESS to 250 m (divided in 14 layers) in UVic. The soil physical properties differ  
 11 among the models as well, and four of them (CLM4.5, ISBA, UVic, UW-VIC) include  
 12 organic horizons. Three models (ISBA, LPJ-GUESS, UW-VIC) did not archive soil sub-grid  
 13 results and provide area-weighted ground temperature (i.e. averaged over wetlands and  
 14 vegetated areas, and in some cases lake fractions).

15

16 Table 1 lists relevant snow model details. One model (UVic) uses an implicit snow scheme  
 17 which replaces the upper soil column with snow-like properties, i.e. the near-surface soil layer  
 18 takes the temperature of the air-snow interface. The other models use separate snow layers on  
 19 top of the ground, either a single bucket (LPJ-GUESS, UW-VIC) or multi-layer snow  
 20 schemes (CLM4.5, CoLM, ISBA, JULES, MIROC-ESM, ORCHIDEE). Snow insulation is  
 21 explicitly considered in all models; increasing snow depth increases the insulation effect.  
 22 Many models consider the effect of varying snow density on insulation (Table 1). This is  
 23 parameterized by a snow conductivity-density relationship that describes how, as snow  
 24 density increases, thermal conductivity increases, thereby reducing the snow insulation. Some  
 25 of the models (LPJ-GUESS, MIROC-ESM, ORCHIDEE, UVic) use a fixed snow density,  
 26 consider only dry snow and no compaction effects, while others represent liquid water in  
 27 snow and different processes for snow densification such as mechanical compaction, and  
 28 thermal and destructive metamorphism (Table 1).

29

30 The simulations were generally run for the period 1960-2009, although some simulations  
 31 were stopped a few years earlier. Each model team was free to choose appropriate driving  
 32 data sets for weather and climate, atmospheric CO<sub>2</sub>, nitrogen deposition, disturbance, land  
 33 cover, soil texture, etc. However, the climate forcing data (surface pressure, surface incident



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1 longwave and shortwave radiation, near-surface air temperature, wind and specific humidity,  
 2 rain and snowfall rates) are from gridded observational datasets (e.g. CRUNCEP, WATCH)  
 3 (SI Table 1). The exception is MIROC-ESM, which was run as a fully-coupled model, forced  
 4 by its own simulated climate. Mean annual temperature of the MIROC-ESM simulations for  
 5 the permafrost region were within the range (-7.2 to 2.2 °C) of the other forcing data sets used  
 6 in this study and the trend in near-surface air temperature (+0.03 °C yr<sup>-1</sup>) was the same for all  
 7 forcing data sets. However, MIROC-ESM had both the highest annual precipitation (range  
 8 433 to 686 mm) and the highest trend in annual precipitation (range -2.1 to +0.8 mm yr<sup>-1</sup>)  
 9 among the forcing data sets.

10

11 The spatial domain of interest is the Northern Hemisphere permafrost land regions. Our  
 12 analysis is based on the 0.5° × 0.5° resolution gridded driving and modeled data for winter  
 13 (DJF) 1980-2000.

## 14 2.2 Observations

15 A data set of monthly near-surface air temperature, 20 cm soil temperatures and snow depth  
 16 from Russian meteorological stations have been provided by the All-Russian Research  
 17 Institute of Hydrometeorological Information-World Data Centre (RIHMI-WDC;  
 18 <http://meteo.ru/>) (Sherstyukov, 2008). 579 stations report snow depth and 268 stations provide  
 19 simultaneous data of all three variables. Ground surface temperature data are not available.  
 20 Precipitation station data have been compiled from the Global Summary of the Day (GSOD)  
 21 data set produced by the National Climatic Data Center (NCDC; <http://www.ncdc.noaa.gov>)  
 22 for all of the stations that are included in the RIHMI-WDC data set. We also use gridded  
 23 snow water equivalent (SWE) data from the GlobSnow-2 product  
 24 (<http://www.globsnow.info/swe/>), which has been produced using a combination of passive  
 25 microwave radiometer and ground-based weather station data (Takala et al., 2011). **Snow**  
 26 **depth was then calculated from SWE using a snow density of 250 kg m<sup>-3</sup>.** Orographic  
 27 complexity, vegetation cover, and snow state (e.g. wet snow) affect the accuracy of this  
 28 product. When compared with ground measurements, the GlobSnow product can show  
 29 **regional differences (of ca. 0.5-5 cm) with biases increasing with increasing SWE** (e.g.,  
 30 Takala et al., 2011; Muskett, 2012). All these data have been compiled for winter (DJF) and  
 31 the same time period of 1980-2000. This period was chosen because soil temperature data are  
 32 sparse before 1980 and the JULES simulation stopped in the year 2000. Comparison of the  
 33 simulations with the station data was done using a weighted bilinear interpolation from the 4  
 34 surrounding model grid points onto the station locations.



### 2.3 Analysis Methods

Our analysis is focused on the common winter (DJF) condition, although snow can begin in November and end at the beginning of May, but we checked that a different winter definition (NDJFMA) does not substantially change the results. The focus in our study is on the evaluation of the simulated air-soil temperature relationships, modulated by snow depth. For this, we analyze the winter mean as well as the interannual variability (expressed as the standard deviation) of 4 variables: near-surface air temperature ( $T_{air}$ ), near-surface soil temperature (soil temperature at 20 cm depth;  $T_{soil}$ ), snow depth ( $d_{snow}$ ), and the difference between  $T_{soil}$  and  $T_{air}$ . This difference  $\Delta T$  ( $\Delta T = T_{soil} - T_{air}$ ) is called the air-soil temperature difference. By limiting our analysis to the winter only, we are able to attribute the across-model and model-to-observation differences in  $\Delta T$  primarily to snow insulation effects. We assume that there is relatively little impact due to soil moisture and texture between surface and 20 cm depth in winter. Although we recognize the difference between ground surface and 20 cm soil temperatures and that soil organic layer could play a role in certain locations (e.g., Romanovsky and Osterkamp, 1995), ground surface temperatures are not recorded in the Russian data set, while 20 cm soil depth temperatures are, hence our choice. However, we find that the results do not significantly change when the model simulated temperature differences between ground surface and near-surface air temperature are used instead of between 20 cm soil and near-surface air temperatures.

We use correlation analysis to investigate the co-variability between  $\Delta T$  and  $d_{snow}$  as well as between  $T_{soil}$  and its two forcing factors ( $T_{air}$  and  $d_{snow}$ ). The input consists of detrended time series of winter means at each grid point. The calculated correlation maps (i.e. spatial distributions of correlation coefficients) based on model and observation data, allow the comparison of the spatial patterns of these relationships. Significance of correlation coefficients is estimated by the Student t-test.

To further examine the functional behavior between different variables, we present relation diagrams between pairs of variables (e.g. variation of  $\Delta T$  with change of  $d_{snow}$ ). To evaluate the performance of the individual LSMs we calculate the root-mean-square error (RMSE) between the observed and modeled relationships. We illustrate the dependence on air temperatures by evaluating the models and observations for several different near-surface air temperature ranges. We split the data into 3 regimes: the coldest conditions ( $T_{air} \leq -25^\circ\text{C}$ ) represent 24% of observations, the intermediate temperature conditions ( $-25^\circ\text{C} < T_{air} \leq -15^\circ\text{C}$ )





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1 represent 42% of the observations, and the warmest conditions ( $-15\text{ °C} < T_{air} \leq -5\text{ °C}$ ) represent  
 2 34% of observations. The principal motivation for such classifications is to distinguish dry  
 3 snow pack regimes from those where sporadic melt may occur even in winter. Hence it is an  
 4 indirect separation of temperature-gradient metamorphosis regimes and density-gradient  
 5 metamorphosis snow pack regimes. Additionally, we present conditional probability density  
 6 functions (PDFs) of  $\Delta T$  for different snow depth and air temperature regimes and compare the  
 7 simulated with those obtained from station observations.

### 8 3 Results

#### 9 3.1 Relationship between air – soil temperature difference and snow depth

10 The air-soil temperature difference ( $\Delta T$ ) - snow depth ( $d_{snow}$ ) relationship in winter (Fig. 1)  
 11 shows that observations and all models produce a clear relationship between increase of  $\Delta T$   
 12 and increases of  $d_{snow}$ . However, Fig. 1 also shows a wide across-model spread in the  
 13 simulated relationships, and that some of the models are not consistent with the behavior in  
 14 the observations. There is also significant scatter in the observation-based relationship, the  
 15 inter-quartile range of  $\Delta T$  is 1.5-8.5 K at specific snow depth and air temperature regimes,  
 16 likely resulting from complicating factors such as snow pack density and moisture content  
 17 variability over the winter, as well as observational errors. Similar ranges of variability are  
 18 produced by several models (such as CLM4.5, CoLM and JULES), but other models (such as  
 19 ISBA and MIROC-ESM) produce noticeably smaller variations.

20

21 The Russian station data and some model results exhibit a linear relation between winter  $\Delta T$   
 22 and  $d_{snow}$  at relatively shallow snow depths with a trend towards asymptotic behavior at larger  
 23 snow depths (Fig. 1), which is in agreement with earlier findings (Zhang, 2005; Ge and Gong,  
 24 2010; Morse et al., 2011). However, only three models (CLM4.5, CoLM, JULES) reproduce  
 25 reasonably well the  $\Delta T$  -  $d_{snow}$  relationship seen in the observational station data (Fig. 1) using  
 26 a benchmark of RMSE  $< 5$  K for all temperature regimes. In particular LPJ-GUESS,  
 27 ORCHIDEE, UVic, UW-VIC, MIROC-ESM show large RMSE for cold air conditions. ISBA  
 28 stands out overall, with a RMSE of 7-18 K in all temperature ranges. We conclude that these  
 29 models do not adequately represent the features of the observed  $\Delta T/d_{snow}$  relationship.

30

31 Figure 2 views the  $\Delta T/d_{snow}$  relationship in the complementary form of the PDFs of  $\Delta T$  for  
 32 different snow depth and air temperature regimes. Since the Russian snow depths are clearly  
 33 non-Normal in distribution (SI Fig. 1, with a median  $d_{snow}$  of 30 cm), we divide the data into



10

1 "shallow" ( $d_{snow} \leq 20$  cm) and "thick" ( $d_{snow} \geq 45$  cm) regimes. The modal value of the station  
 2 data  $\Delta T$  PDF is 5 K for "shallow" snow and 14 K for "thick" snow - that is thick snow is a  
 3 better insulator than thin snow. Based on the  $\Delta T$  PDFs, five models (CoLM, CLM4.5, JULES,  
 4 ORCHIDEE, MIROC-ESM) successfully separate the  $\Delta T$  regimes under different snow depth  
 5 conditions, while the other models clearly fail for at least one of these snow depth regimes.  
 6 However, even for the better models, both the shapes and the modal values of the simulated  
 7 PDFs differ from the observed PDF.

8

9 Both Figs. 1 and 2 indicate that air-soil temperature differences are related to air temperature  
 10 conditions. This is due to snow pack properties, particularly its density and moisture content,  
 11 that affect the air-soil temperature difference. For example, the density of fresh fallen snow  
 12 tends to be much lower under cold air temperatures than warm (Anderson, 1976), leading to  
 13 increased insulation (larger  $\Delta T$ ). Snow densification is also a function of air temperature, for  
 14 example, depth hoar metamorphosis of the snow pack, which produces more insulation  
 15 (loosely packed depth-hoar crystals have very low thermal conductivity), is promoted by  
 16 strong thermal gradients in the snow pack, and is typical of continental climates (e.g., Zhang  
 17 et al., 1996).

18

19 The observations in Figs. 1 and 2 indicate that snow under colder climates have greater  
 20 insulation than under warmer climates. This is shown by a larger  $\Delta T$  for colder  $T_{air}$  than for  
 21 warmer  $T_{air}$  (for a certain snow depth) and a greater sensitivity of  $\Delta T$  to changes in  $d_{snow}$  (Fig.  
 22 1), and by the larger modal value of the  $\Delta T$  PDF for colder  $T_{air}$  than for warmer  $T_{air}$  (21 K for  
 23  $T_{air} \leq -25$  °C and 9 K for  $-15$  °C  $< T_{air} \leq -5$  °C; Fig. 2). This is consistent with colder climates  
 24 having lower density snow packs, and the differences are in line with measurements of snow  
 25 density variability (Zhong et al., 2013). Additionally, both the inter-quartile range in Fig. 1  
 26 and the width of the PDFs in Fig. 2 become larger as air temperatures cool. This may be  
 27 related to the formation of depth hoar, which is a very good insulator and its varying presence  
 28 in the snow pack decouples  $\Delta T$  from  $d_{snow}$ . Cold, thin snow packs tend to contain much more  
 29 low density depth hoar than warmer snow packs (e.g., Zhang et al., 1996; Singh et al., 2011).  
 30 Continental regions have large annual temperature cycles, with greater interannual variability  
 31 and thinner snow packs, than maritime ones. This variability leads to greater scatter and  
 32 greater sensitivity of the  $\Delta T/d_{snow}$  relationship in the cold winter regions. An additional cause  
 33 of scatter is that the density of fresh-fallen snow decreases with falling temperature.  
 34 Accordingly, we find in the cold air temperature regime ( $T_{air} \leq -25$  °C) a larger  $\Delta T$  in early



11

1 winter (November-December) when the snow pack is composed of thin, low density fresh  
 2 snow (and depth hoar) than in late winter (January-February) (SI Fig. 2). Under warm  
 3 conditions ( $-15\text{ °C} < T_{\text{air}} \leq -5\text{ °C}$ ) such a separation is not observed.

4  
 5 Our analysis (Fig. 1) indicates that some models (CLM4.5, CoLM, JULES) are better able to  
 6 replicate the observed effect of air temperature on the  $\Delta T / d_{\text{snow}}$  relationship than others (LPJ-  
 7 GUESS, MIROC-ESM, ORCHIDEE, UW-VIC). The latter do not fully replicate the larger  
 8  $\Delta T$  under cold air temperature conditions. CLM4.5, CoLM and JULES capture a larger  $\Delta T$  for  
 9 colder air temperatures for a given  $d_{\text{snow}}$  in agreement with the observations. However, for  
 10 shallow snow JULES simulates twice as large increase of  $\Delta T$  with increasing  $d_{\text{snow}}$  for all  
 11 temperature ranges, compared with observations. Two models (ISBA, UVic) clearly fail in  
 12 this evaluation. Poor model performance in reflecting air temperature influence on the  $\Delta T$   
 13  $/d_{\text{snow}}$  also manifests itself in regime separation of the PDFs (Fig. 2). Some models do not  
 14 separate the  $\Delta T$  regimes under different air temperature conditions well or at all (ISBA, LPJ-  
 15 GUESS, MIROC-ESM, UVic), while others cannot capture the observed cold temperature  
 16 regime features (i.e., too broad PDFs and shifts towards smaller modal values; ORCHIDEE,  
 17 UW-VIC). The three models with reasonable inter-variable relations (CLM4.5, CoLM,  
 18 JULES) also capture the regime separation in the PDFs. These three models as well as LPJ-  
 19 GUESS and ORCHIDEE also represent the observed greater insulation of early winter snow  
 20 packs under cold conditions (SI Fig. 2).

21  
 22 The maps of the  $\Delta T / d_{\text{snow}}$  correlations in winter (Fig. 3) demonstrate the strong spatial  
 23 variability in the  $\Delta T / d_{\text{snow}}$  relationship, but indicate that most models agree on the general  
 24 large-scale pattern. Some models (CLM4.5, CoLM, ORCHIDEE, UW-VIC) show a  
 25 reasonable pattern correlation coefficient ( $r \geq 0.4$ ) with observations, while the others do not.  
 26 Most models simulate the highest positive correlation in the region of the East Siberian Plain  
 27 and Siberian High lands. In some regions, namely in Scandinavia, West Russian Arctic, West  
 28 and Central Siberian Plains, the correlation is much weaker and often not statistically  
 29 significant. These are the regions of large winter snow depth (Sect. 4.1.2) which are  
 30 influenced by North Atlantic cyclonic activity which brings relatively warm moist air and  
 31 heavy precipitation in winter (and a positive correlation between snow depth and air  
 32 temperature), leading to relatively small mean  $\Delta T$ . Obvious outliers in the  $\Delta T / d_{\text{snow}}$  correlation  
 33 map are the LPJ-GUESS and UVic models, which do not reproduce the observed pattern of  
 34 correlation. UVic calculates a reverse pattern correlation than observations for many regions



12

1 (e.g. significant positive correlation in West Siberian Plain and Central Siberian Highlands).  
 2 LPJ-GUESS produces very few statistically significant correlations. The model correlations  
 3 are likely highly sensitive to the quality of the snowfall forcing data, which is uncertain across  
 4 much of the region due to limited station data that go into most global snowfall products  
 5 (Hancock et al., 2014; Drobot et al., 2006).

### 6 **3.2 Variability of soil temperature with air temperature and snow depth**

7 Next we assess whether or not the models can correctly reproduce the interannual near-  
 8 surface soil temperature ( $T_{soil}$ ) variability in relation to snow depth ( $d_{snow}$ ) and near-surface air  
 9 temperature ( $T_{air}$ ) variability. ~~Previous authors~~ (Smith and Riseborough, 2002; Sokratov and  
 10 Barry, 2002; Zhang, 2005; Lawrence and Slater, 2010) have noted that the strength of  
 11 relationship between  $T_{soil}$  and  $T_{air}$  is modulated by  $d_{snow}$  and the snow insulation effect  
 12 increases only up to a limiting depth beyond which extra snow makes little difference to soil  
 13 temperatures. Zhang (2005) reports that the limiting snow depth is approximately 40 cm.

14

15 To inspect the difference of the insulation effects on both sides of such a limiting snow depth,  
 16 we investigate the  $T_{soil}/T_{air}$  relationship under shallow ( $d_{snow} \leq 20$  cm) and thick ( $d_{snow} \geq 45$  cm)  
 17 snow conditions. Observations showed that the slope of this relationship is higher when the  
 18 snow cover is thin, compared with thicker snow conditions (e.g., for Yukon Territory in  
 19 Canada; Karunaratne and Burn, 2003). Indeed, the Russian observations (Fig. 4, Table 2)  
 20 indicate a three times stronger  $T_{soil}/T_{air}$  relationship ( $0.62$  °C/°C,  $R^2=0.8$ ) under shallow snow  
 21 pack than thicker snow conditions ( $0.21$  °C/°C,  $R^2=0.4$ ). The average  $d_{snow}$  in the shallow snow  
 22 regime is 13.7 cm and that for the thick snow regime is 58.5 cm, so we would expect, if near-  
 23 surface air temperature and conductivities were equal in both snow depth classes, a ratio of  
 24 4.3 in the slopes of Fig. 4. The models that better reproduce the observed  $\Delta T / d_{snow}$   
 25 relationship (CLM4.5, CoLM, JULES) reproduce the observed variation in the  $T_{soil} / T_{air}$   
 26 relation better than others. JULES and CoLM indicate a factor of 4 change, while CLM4.5  
 27 indicates a factor of 2 change. Other models (LPJ-GUESS, MIROC-ESM, ORCHIDEE)  
 28 strongly underestimate the increase of the slope for decreasing snow depth (they simulate a  
 29 factor change of about 1.5). The two models that had unrealistic  $\Delta T / d_{snow}$  relationships (ISBA,  
 30 UVic) also fail in this evaluation of their  $T_{soil} / T_{air}$  relationship. They simulate too strong  $T_{soil}$   
 31  $/T_{air}$  relationships (gradients larger than  $0.9$  °C/°C,  $R^2>0.7$ ; Table 2) that are almost completely  
 32 independent of the snow depth regimes, particularly in ISBA, which is not consistent with  
 33 observations. These models' spatial correlation patterns between  $T_{soil}$  and  $T_{air}$  also differ  
 34 greatly from the observations and the other models (SI Fig. 3) and show very high positive



13

1 correlation ( $r > 0.8$ ) in most regions, as may be expected from the strong relationship shown  
 2 in Fig. 4. The RMSE of their modeled  $T_{soil}/T_{air}$  relationships from observations reaches ca.  
 3  $10\text{ }^{\circ}\text{C}$ .

4  
 5 The  $T_{soil}/d_{snow}$  relationship (Fig. 5) displays the variation of  $T_{soil}$  with changing snow depth  
 6 and emphasizes the weakening role of snow depth for  $T_{soil}$  under thick snow conditions. With  
 7 increasing  $d_{snow}$ , soil temperatures asymptotically converge towards a value of around  $0\text{ }^{\circ}\text{C}$ .  
 8 Overall, the Russian observations indicate that snow depth above about 80-90 cm has very  
 9 little additional insulation effect on  $T_{soil}$ . Most of the models show consistent results with  
 10 regard to this aspect, although the inter-quartile range of  $T_{soil}$  for a certain snow depth is quite  
 11 large in some models (ISBA, ORCHIDEE, UVic, UW-VIC) (Fig. 5). The figure further points  
 12 to the air temperature dependency of the relation. On average, for a given  $d_{snow}$ , a colder  $T_{soil}$   
 13 is observed for colder near-surface air temperatures, compared with warmer air temperatures.  
 14 Most models can replicate this effect of  $T_{air}$  on the  $T_{soil}/d_{snow}$  relationship, though with  
 15 differing accuracy. The RMSE between the observed and modeled relationships can reach ca.  
 16  $10\text{ }^{\circ}\text{C}$  and more (in ISBA, UVic, UW-VIC), particularly under cold conditions.

17  
 18 The spatial patterns of the correlation coefficients between  $T_{soil}$  and  $T_{air}$  (SI Fig. 3) and  
 19 between  $T_{soil}$  and  $d_{snow}$  (SI Fig. 4) show a relatively large across-model scatter in the specific  
 20 regions. Obvious outliers in the  $T_{soil}/T_{air}$  correlation maps (SI Fig. 3) are ISBA and UVic  
 21 which strongly overestimate the correlation ( $r > 0.9$ ) over most of the Arctic. Other models  
 22 (LPJ-GUESS, ORCHIDEE, UW-VIC) also overestimate the correlation in some regions (e.g.  
 23 western Russian Arctic,  $r > 0.7$ ). Most of the simulated maps of  $T_{soil}/d_{snow}$  correlation (SI Fig.  
 24 4) agree with the observations on a strong positive correlation in East Siberia. This is a region  
 25 of relatively shallow snow (10-40 cm; Fig. 6) and there  $T_{soil}$  is very sensitive to variations in  
 26 snow depth (e.g., Romanovsky et al., 2007). Comparing both simulated correlation maps, it is  
 27 obvious that in this region,  $T_{soil}$  correlates more strongly with  $d_{snow}$  than with  $T_{air}$ , in agreement  
 28 with the Russian data and earlier studies (Romanovsky et al., 2007; Sherstyukov, 2008).

#### 29 4 Roles of atmospheric forcing and model processes

30 The across-model differences in the snow insulation effect, presented by the air temperature -  
 31 snow depth - soil temperature relationships described above, are partially due to differences in  
 32 the atmospheric forcing data and also due to differences in the snow and soil physics used in  
 33 the LSMs. However, because the climate forcing data sets utilized with each model are



14

1 observation-based (except for MIROC-ESM), obvious outliers in individual model  
 2 performance likely mainly indicate poor or deficient physical descriptions of the air/snow/soil  
 3 relations in that specific LSM.

#### 4 **4.1 Atmospheric forcing and snow depth**

##### 5 **4.1.1 Air temperature and precipitation**

6 Both near-surface air temperature and precipitation are given by the climate forcing data sets  
 7 (SI Table 1) for all models, except for MIROC-ESM which simulates both. The across-model  
 8 differences in forcing air temperature used are relatively small and the simulated spatial  
 9 patterns of temperature are very similar (SI Fig. 5). All forcing datasets are somewhat colder  
 10 than Russian station data in their grid cells. The biases range from -0.8 K to -4.7 K (SI Table  
 11 2), reflecting biases in the climate forcing data used by the models. In contrast, MIROC-ESM  
 12 has a positive (mean) air temperatures bias of +2.7 K.

13

14 The large-scale patterns of precipitation are similar across the models, but regional differences  
 15 can be large (SI Fig. 6). The individual differences in precipitation range from -0.2 mm/day  
 16 to +0.5 mm/day (SI Table 2) relative to the average of the Russian station data. Unfortunately,  
 17 snowfall was archived in only a few models, however large-scale spatial patterns are similar  
 18 across these models (SI Fig. 7).

##### 19 **4.1.2 Snow depth**

20 The broad-scale spatial snow depth ( $d_{snow}$ ) patterns are similar across the models and show  
 21 general agreement with the observed patterns (Fig. 6). The well-pronounced areas of  
 22 maximum winter  $d_{snow}$  (50-100 cm) are in Scandinavia, the Urals, the West Siberian Plain,  
 23 Central Siberian Highlands, the Far East, Alaskan Rocky mountains, and Labrador Peninsula  
 24 and isle of Newfoundland. However, large regional across-model variability is obvious. Some  
 25 models (JULES, LPJ-GUESS, ORCHIDEE, UVic) underestimate  $d_{snow}$ , while others  
 26 (CLM4.5, CoLM, ISBA, UW-VIC) overestimate it (Fig. 6; SI Table 3). The model biases are  
 27 quite similar with respect to station observations and GlobSnow data. The evaluation of the  
 28 model performance for SWE compared to GlobSnow indicates the same bias characteristics  
 29 as described here for snow depth (not shown). It should be noted, that the models do not  
 30 account for snowdrift. However, redistribution of snow due to wind is an important aspect,  
 31 which makes comparison between in-situ measured and modeled snow depths difficult (e.g.,  
 32 Vionnet et al., 2013; Sturm and Stuefer, 2013; Gispnas et al., 2014).



15

1  
 2 Precipitation/snowfall across-model differences cannot be the primary explanation of these  
 3  $d_{snow}$  differences since some models (JULES, MIROC-ESM, ORCHIDEE) have positive bias  
 4 in precipitation ( $> 0.2$  mm/d, SI Table 2) but simulate much lower  $d_{snow}$  compared to other  
 5 models (Fig. 6, SI Figs. 6, 7, SI Table 3). Across-model differences in the interannual  
 6 variability of winter precipitation do not translate simply to corresponding differences in the  
 7 interannual  $d_{snow}$  variability (not shown). For example, UVic calculates the (unrealistically)  
 8 largest interannual  $d_{snow}$  variability in the boreal Europe permafrost region which is not  
 9 reflected in the precipitation variability. These results indicate that the simulated snow depth  
 10 is a function of both, the prescribed winter precipitation, and the model's snow energy and  
 11 water balance.

## 12 4.2 Model processes

13 We have shown that the across-model spread in the representation of snow insulation effects  
 14 (Sects. 3.1, 3.2) can not predominantly be explained by differences in the forcing data (Sect.  
 15 4.1), but to a large extent is due to the representation of snow processes in the models. By  
 16 considering the relationship plots and the conditional PDFs (Figs. 1, 2, 4, and 5) we were  
 17 able to sort the models in terms of their snow insulation performance. In this section we  
 18 discuss the influence of the different snow parameterizations in the models.

19

20 Models with better performance (CLM4.5, CoLM, JULES) apply multi-layer snow schemes.  
 21 This allows them to simulate more realistic (stronger) insulation because they consider the  
 22 snowpack's vertical structure and variability. They calculate the energy and mass balance in  
 23 each snow layer, are able to capture nonlinear profiles of snow temperature, and can also  
 24 account for thermal insulation within the snowpack such as when the upper layer thermally  
 25 insulates the lower layers (e.g., Dutra et al., 2012). These models also incorporate storage and  
 26 refreezing of liquid water within the snow, parameterize wet snow metamorphism, snow  
 27 compaction, and snow thermal conductivity (Table 1), which have been found to be among  
 28 the most important processes for good snow depth and surface soil temperature simulation  
 29 (e.g., Wang et al., 2013).

30

31 An underestimated snow depth directly leads to insulation that is too weak in JULES, LPJ-  
 32 GUESS, ORCHIDEE, and UVic (Fig. 6, SI Table 3). However only in ORCHIDEE and UVic  
 33 does this lead to a significant underestimation of  $\Delta T$  (SI Table 3, SI Fig. 8) indicating bias





16

1 compensation in the two other models. Thus, compensating error effects occur due to snow  
 2 density and conductivity (SI Fig. 9, Table 1), which impact snow thermal insulation.

3  
 4 Our analysis showed that two models (ISBA, UVic) have  $T_{soil}/T_{air}$  correlation that are too  
 5 high indicating that they do not represent the modulation of the  $T_{soil}/T_{air}$  relationship by snow  
 6 depth (Fig. 4). This is consistent with their underestimation of  $\Delta T$  (Figs. 1 and 2, SI Fig. 8, SI  
 7 Table 3). In UVic, the snowpack is treated not as a separate layer but as an extension of the  
 8 top soil layer and a combined surface-to-soil thermal conductivity is calculated (Table 1).  
 9 Such a scheme largely negates or reduces the insulating capacity of snow (Slater et al., 2001).  
 10 Koven et al. (2013) noted that such a scheme simulates very little warming of soil, and  
 11 sometimes even cooling. The slightly underestimated snow depth (SI Table 3, Fig. 6)  
 12 contributes (but not as the primary factor) to reduced snow insulation, as reported for UVic  
 13 (Avis, 2012).

14  
 15 ISBA strongly underestimates  $\Delta T$ , while strongly overestimating  $d_{snow}$ , compared with  
 16 observations (SI Table 3, Fig. 6). However, ISBA uses the same atmospheric forcing data as  
 17 JULES (accordingly the air temperature and precipitation are quite similar; SI Table 2). Also,  
 18 the model's snow density ( $150\text{--}250\text{ kg m}^{-3}$ ) is similar to other models (CLM45, CoLM,  
 19 JULES) (SI Fig. 9) and in agreement with Zhong et al. (2013) who report snow density values  
 20 of on  $180\text{--}250\text{ kg m}^{-3}$  for tundra/taiga and  $156\text{--}193\text{ kg m}^{-3}$  for alpine snow classes in winter.  
 21 This apparent contradiction comes from the parameterization of snow cover fraction within  
 22 each grid cell (SCF). The version of ISBA used here calculates a unique superficial soil  
 23 temperature whether or not the soil is covered by snow and all the energy and radiative fluxes  
 24 are area-weighted by SCF (equations 7 and 20 in Douville et al., 1995). In order to get  
 25 reasonable albedos in snow-covered forests, as is necessary when ISBA is coupled to the  
 26 CNRM-CM climate model, the parameterization gives very low SCF in the boreal forest  
 27 (between 0.2 and 0.5). Hence, snow insulates only 20% to 50% of the grid cell, despite fairly  
 28 high snow depths. The heat fluxes from the snow-covered fraction are averaged with the  
 29 fluxes from the snow-free surface, strongly concealing the actual insulating effect of snow  
 30 and underestimating it over the grid cell. Using the detailed snow model Crocus (Brun et al.,  
 31 1992; Vionnet et al., 2012) with a SCF equal to 100% leads to an almost perfect simulation of  
 32 near-surface soil temperature over Northern Eurasia (Brun et al., 2013). A similar experiment  
 33 with ISBA and a SCF equal to 100% (Decharme et al., 2015) leads to good performances  
 34 showing that the low  $\Delta T$  in ISBA despite high snow depth in the present study is mostly due





17

1 to this sub-grid snow fraction. The results are further improved by updating the snow albedo  
 2 and snow densification parameterization.

3

4 Interestingly, the ORCHIDEE performance in simulating snow depth and  $\Delta T$  is similar to  
 5 UVic (underestimation of  $d_{snow}$  and  $\Delta T$ ; SI Table 3). However, ORCHIDEE can better  
 6 represent the observed  $T_{soil}/T_{air}$  relationship and its modulation due to snow pack.  
 7 ORCHIDEE employs, similarly to UVic, a fixed snow density and thermal conductivity.  
 8 However, in contrast with UVic, ORCHIDEE applies a multi-layer scheme and simulates heat  
 9 diffusion in the snowpack in up to 7 discrete layers (Table 1; Koven et al., 2009). This helps  
 10 resolving the snow thermal gradients between the top and the base of the snow cover, and  
 11 might explain how some of the snow insulation effects are reasonably represented in  
 12 ORCHIDEE, despite the simpler treatment of temperature diffusion.

## 13 5 Summary and conclusions

14 The aim of this work was to evaluate how state-of-the-art LSMs capture the observed  
 15 relationship between winter near-surface soil and air temperatures ( $T_{soil}$ ,  $T_{air}$ ) and their  
 16 modulation by snow depth ( $d_{snow}$ ) and climate regime. We presented some benchmarks to  
 17 evaluate model performance. The results are based on the comparison of LSMs with a novel  
 18 comprehensive Russian station data set.

19

20 We see large differences across the models in their mean air-soil temperature difference ( $\Delta T$ )  
 21 of 3 to 14 K, in the gradient between near-surface soil and air temperatures ( $T_{soil}/T_{air}$ ) (0.49 to  
 22 0.96 °C/°C for shallow snow, 0.13 to 0.93 °C/°C for thick snow), and in the increase of  $\Delta T$   
 23 with increasing snow depth (modal value of  $\Delta T$  PDF: 0 to 10 K for shallow snow, 5 to 21 K  
 24 for thick snow). Most of the nine models compare to the observations reasonably well  
 25 (observations:  $\Delta T = 12$  K, modal  $\Delta T$  values of 5 K for shallow snow and of 14 K for thick  
 26 snow,  $T_{soil}/T_{air} = 0.62$  °C/°C for shallow snow,  $T_{soil}/T_{air} = 0.21$  °C/°C for thick snow). Several  
 27 models also capture the modulation by air temperature condition (larger increase in  $\Delta T$  with  
 28 increasing  $d_{snow}$  under colder conditions) and display the control of snow depth on  $T_{soil}$   
 29 (weaker  $T_{soil}/T_{air}$  relationship under thicker snow). However, while they generally capture  
 30 these observed relationships, their strength can differ in the individual models. Two models  
 31 (ISBA, UVic) show the largest deficits in snow insulation effects and cannot separate the  $\Delta T$   
 32 regimes neither for different snow depths nor for different air temperature conditions.

33



18

1 The primary aim of this study was to use this ensemble of models to document model  
 2 performance with respect to  $T_{soil}$  versus  $T_{air}$  relationships, and to identify those with better  
 3 performance, rather than to quantify the best model. We were able to attribute performance  
 4 strength/weakness to snow model features and complexity. Models with better performance  
 5 apply multi-layer snow schemes and consider complex snow processes (e.g. storage and  
 6 refreezing of liquid water within the snow, wet snow metamorphism, snow compaction).  
 7 Those models which show limited skill in snow insulation representation (underestimated  $\Delta T$ ,  
 8 very weak dependency of  $\Delta T$  on  $d_{snow}$ , almost unity ratio of  $T_{soil}/T_{air}$ ) have some deficiencies  
 9 in the simulation of heat transfer in snow and soil layer, particularly in the representation of  
 10 snow depth and density (conductivity). We also emphasize that compensating errors in snow  
 11 depth and conductivity can occur. For example, an excessive correlation between  $T_{soil}$  and  $T_{air}$   
 12 can be attributed to excessively high thermal conductivity even when the snow depth is  
 13 correctly (or over) simulated. This finding underscores the need for detailed model  
 14 evaluations using multiple, independent performance metrics to establish that the models get  
 15 the right functionality for the right reason. It should be noted that the treatment of ground  
 16 properties, particularly soil organic matter and soil moisture/ice content, also affect the  
 17 simulated winter ground temperatures. The specific evaluation of these individual processes is  
 18 more robustly investigated with experiments conducted for individual models (e.g. recently,  
 19 Wang et al., 2013; Gubler et al., 2013; Decharme et al., 2015).

20

21 A realistic simulation of the snow is a key pre-requisite for accurate modeling of the soil  
 22 thermal dynamics across the permafrost region. The areal cover of Northern Hemisphere near-  
 23 surface permafrost varies greatly across the nine models in the hindcast simulation (1960-  
 24 2009). Some of the better performing snow insulation effect models (CLM4.5, JULES)  
 25 simulate a near-surface permafrost area of 12 to 16 million  $\text{km}^2$ , which is comparable with the  
 26 IPA map estimate (16 million  $\text{km}^2$ ) (Brown et al., 1997; Slater and Lawrence, 2013). CoLM  
 27 and ORCHIDEE, identified as reasonable models with respect to snow insulation, simulate  
 28 much lower (7 million  $\text{km}^2$ ) and higher (20 million  $\text{km}^2$ ) areas, respectively. However, ISBA,  
 29 one of the two models that showed rather limited skill in representing snow insulation effects,  
 30 also simulates the highest permafrost area (20 million  $\text{km}^2$ ). This is consistent with previous  
 31 studies (e.g., Vavrus, 2007; Koven et al., 2013) which concluded that first-order control on  
 32 modeled near-surface permafrost distribution is the representation of the air-to-surface soil  
 33 temperature difference. When the models with poor snow models are eliminated, there is no  
 34 clear relationship between the quality of the snow insulation in the models and the simulated



19

1 area of permafrost, likely because several other factors such as differences in the treatment of  
 2 soil organic matter, soil hydrology, surface energy calculations, model soil column depth, and  
 3 vegetation also provide important controls on simulated permafrost distribution (Marchenko  
 4 and Etzelmüller, 2013).

5  
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## 1 Tables

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3 **Table 1. PCN snow model details.**

| Model<br>Reference for snow<br>scheme                          | Snow<br>scheme <sup>1</sup> | Snow<br>layers      | Water<br>phases          | Liquid water<br>treatment <sup>2</sup>                  | Snow density <sup>3</sup>                                      | Snow thermal<br>conductivity <sup>4</sup>  |
|--|-----------------------------|---------------------|--------------------------|---|--|--|
| CLM4.5<br>Swenson and<br>Lawrence, 2012<br>Oleson et al., 2013 | ML                          | Dynamic<br>(max. 5) | Liquid,<br>Ice           | Bucket-type<br>prognostic in<br>each layer              | depends on<br>snow depth;<br>compaction<br><sup>3) a,b,c</sup> | quadratic<br>equation on $\rho$  |
| CoLM<br>Dai et al., 2003<br>Ji et al. 2014                     | ML                          | Dynamic<br>(max. 5) | Liquid,<br>Ice           | Bucket-type<br>prognostic in<br>each layer              | depends on<br>snow depth;<br>compaction<br><sup>3) a,b,c</sup> | quadratic<br>equation on $\rho$  |
| ISBA<br>Boone and Etchevers,<br>2001                           | ML                          | Static<br>(3)       | Liquid,<br>Ice,<br>Vapor | Diagnosed<br>from snow<br>temperature,<br>mass, density | compaction<br><sup>3) a,b</sup>                                | quadratic<br>equation on $\rho$ ,<br>contribution due<br>to vapor transfer   |
| JULES<br>Best et al., 2011                                     | ML                          | Dynamic<br>(max. 3) | Liquid,<br>Ice,<br>Vapor | Bucket-type<br>prognostic in<br>each layer              | compaction<br><sup>3) a</sup>                                  | power equation<br>on $\rho$  |
| LPJ-GUESS<br>Gerten et al., 2004<br>Wania et al., 2009         | BL                          | Static<br>(1)       | Ice                      | Not<br>represented                                      | fixed<br>362 kg m <sup>-3</sup>                                | fixed<br>0.196 Wm <sup>-1</sup> K <sup>-1</sup>  |
| MIROC-ESM<br>Takata et al., 2003                               | ML                          | Dynamic<br>(max. 3) | Ice                      | Not<br>represented                                      | fixed<br>300 kg m <sup>-3</sup>                                | fixed<br>0.3 Wm <sup>-1</sup> K <sup>-1</sup>  |
| ORCHIDEE<br>Gouttevin et al., 2012                             | ML                          | Dynamic<br>(max. 7) | Ice                      | Not<br>represented                                      | fixed<br>330 kg m <sup>-3</sup>                                | fixed<br>0.25 Wm <sup>-1</sup> K <sup>-1</sup><br>for tundra,<br>0.042 Wm <sup>-1</sup> K <sup>-1</sup><br>for taiga |
| UVic<br>Meissner et al., 2003<br>Avis, 2012                    | I                           | Static<br>(1)       | Ice                      | Not<br>represented                                      | fixed<br>330 kg m <sup>-3</sup>                                | bulk<br>conductivity   |
| UW-VIC<br>Andreadis et al., 2009                               | BL                          | Dynamic<br>(max. 2) | Liquid,<br>Ice,<br>Vapor | Constant<br>liquid water<br>holding<br>capacity         | compaction<br><sup>3) a,b</sup>                                | fixed<br>0.7 Wm <sup>-1</sup> K <sup>-1</sup>  |

4 <sup>1</sup> ML: Multi-layer, BL: Bulk-layer, I: Implicit; according to Slater et al. (2001)

5 <sup>2</sup> Not represented means dry snow

6 <sup>3</sup> Processes for densification of the snow: a) mechanical compaction (due to the weight of the overburden), b)  
 7 thermal metamorphosis (via the melting–refreezing process), c) destructive metamorphism (crystal breakdown  
 8 due to wind, thermodynamic stress); Anderson (1976), Jordan (1991), Kojima (1967)

9 <sup>4</sup> quadratic equation on  $\rho$  according to Jordan (1991), Anderson (1976); contribution due to vapor transfer  
 10 according to Sun et al.(1999)

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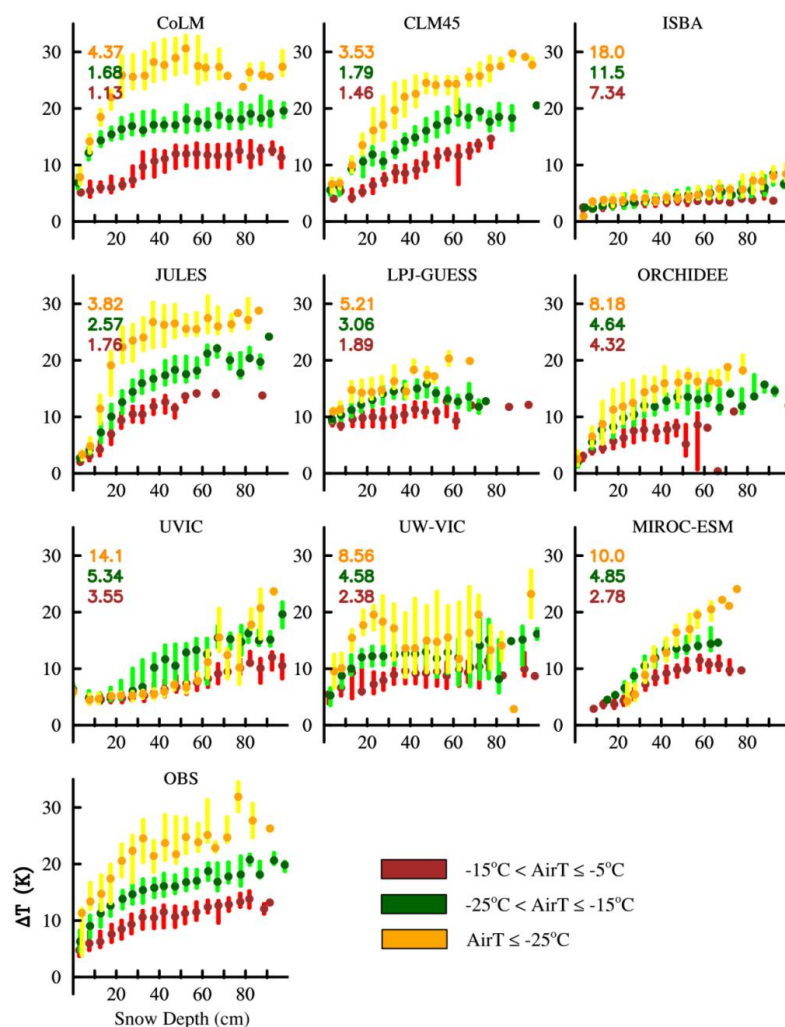
**Table 2.** Sensitivity of near-surface soil temperature ( $T_{soil}$ ) to air temperature ( $T_{air}$ ) in winter (DJF) calculated by the slopes of the linear regression between  $T_{soil}$  (°C) and  $T_{air}$  (°C) for different regimes of snow depth ( $d_{snow}$ ), using data from all Russian station grid points and 21 individual winter 1980-2000. All relationships are statistically significant at  $p \leq 0.01$ .

|             | Snow depth regimes            |       |                               |       |
|-------------|-------------------------------|-------|-------------------------------|-------|
|             | Shallow                       |       | Thick                         |       |
|             | $d_{snow} \leq 20$ cm         |       | $d_{snow} \geq 45$ cm         |       |
|             | $T_{soil}/T_{air}$<br>(°C/°C) | $R^2$ | $T_{soil}/T_{air}$<br>(°C/°C) | $R^2$ |
| Observation | 0.62                          | 0.79  | 0.21                          | 0.41  |
| CLM4.5      | 0.69                          | 0.89  | 0.33                          | 0.56  |
| CoLM        | 0.49                          | 0.73  | 0.13                          | 0.44  |
| ISBA        | 0.93                          | 0.98  | 0.93                          | 0.94  |
| JULES       | 0.68                          | 0.77  | 0.19                          | 0.46  |
| LPI-GUESS   | 0.73                          | 0.89  | 0.52                          | 0.75  |
| MIROC-ESM   | 0.78                          | 0.98  | 0.49                          | 0.67  |
| ORCHIDEE    | 0.86                          | 0.83  | 0.56                          | 0.64  |
| UVic        | 0.96                          | 0.97  | 0.81                          | 0.68  |
| UW-VIC      | 0.54                          | 0.74  | 0.76                          | 0.65  |

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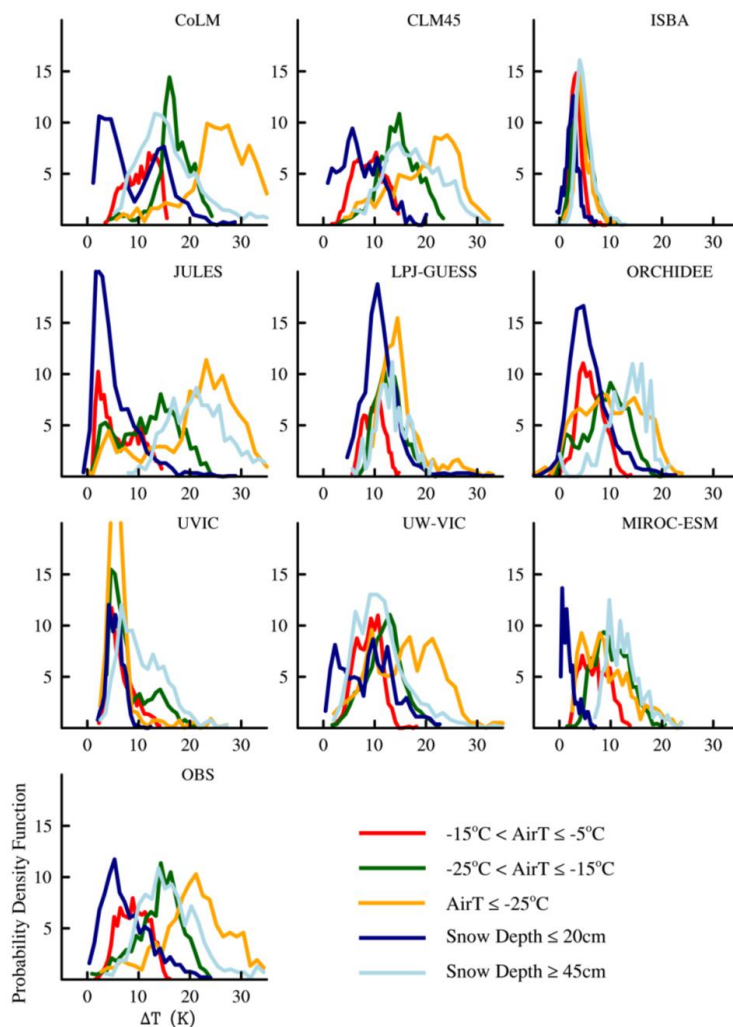
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3 **Figure 1.** Variation of  $\Delta T$  (K), the difference between soil temperature at 20 cm depth and air  
 4 temperature) with snow depth (cm) for winter 1980-2000. The **dots** represent the medians of 5  
 5 cm snow depth bins and the upper and lower bars indicate the 25<sup>th</sup> and 75<sup>th</sup> percentiles,  
 6 calculated from all Russian station grid points (n=268) and 21 individual winters. The  
 7 numbers in each model panel indicate the RMSE between the observed and modeled  
 8 relationship. Color represents different air temperature regimes.



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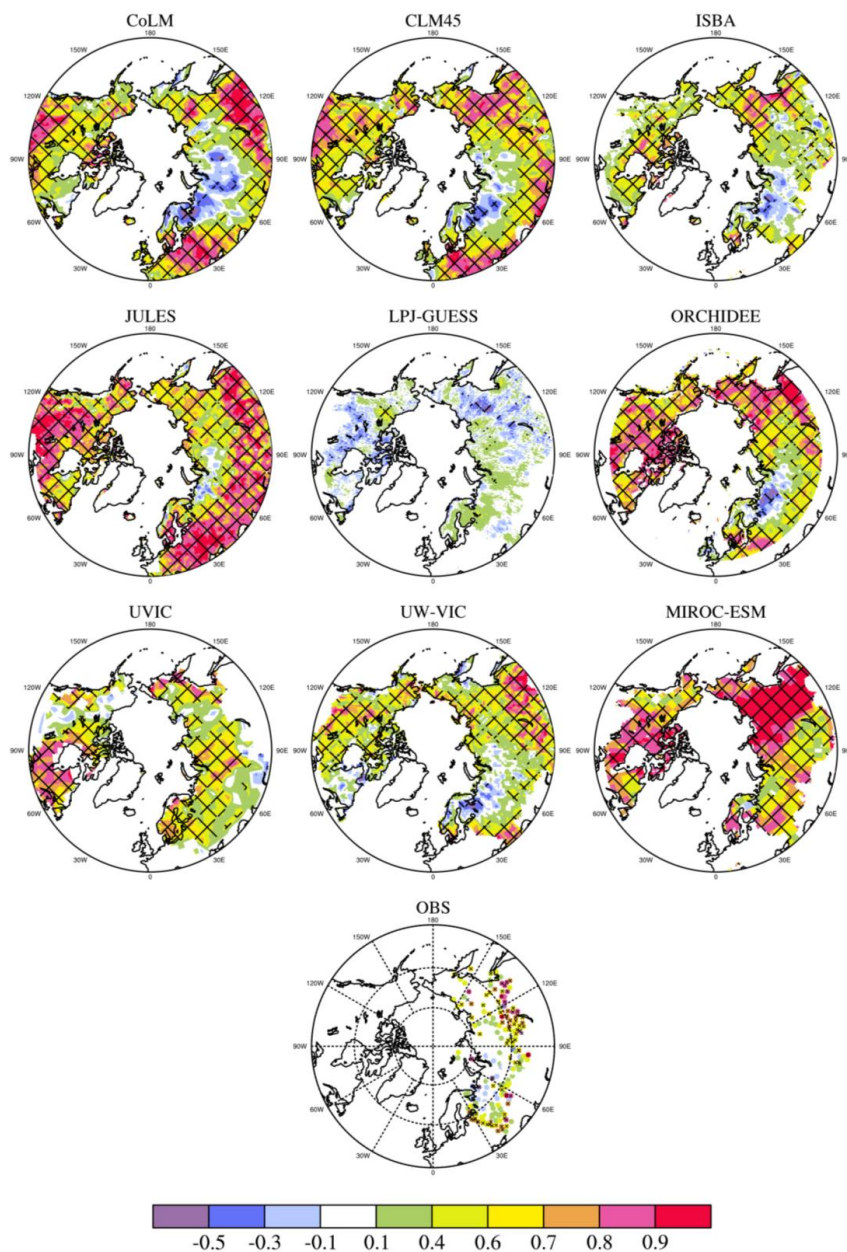
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3 **Figure 2.** Conditional probability density functions (PDFs) of  $\Delta T$  (K), the difference between  
 4 soil temperature at 20 cm depth and air temperature for different snow depth and air  
 5 temperature regimes (color) for winter 1980-2000.



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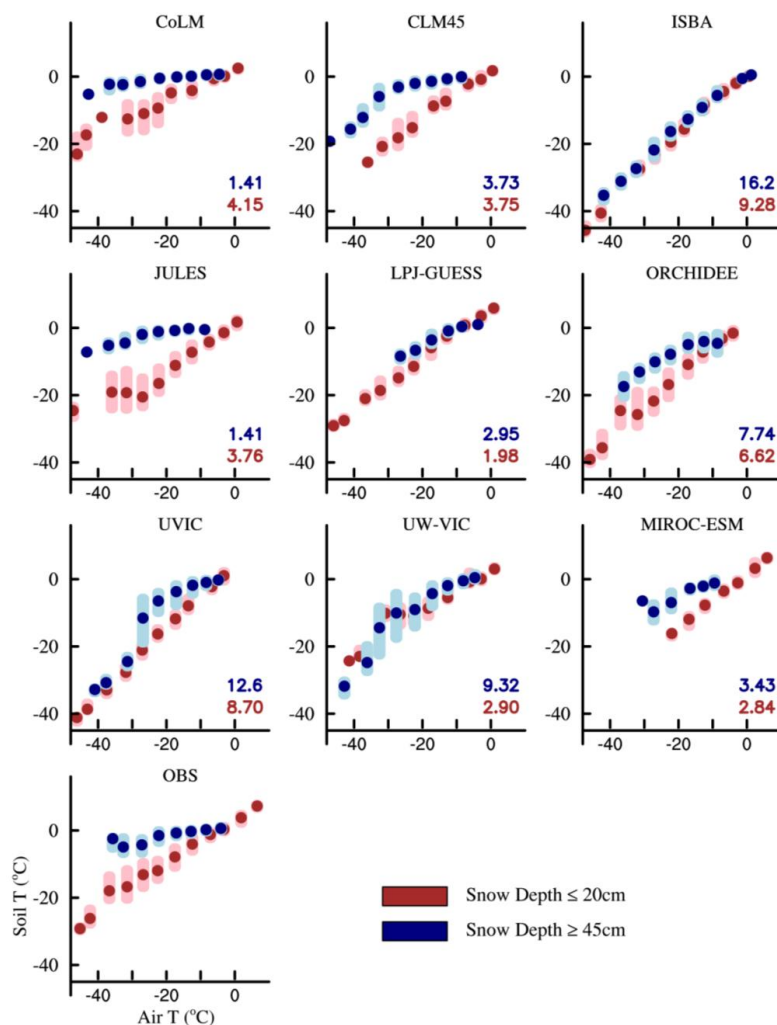


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3 **Figure 3.** Spatial maps of the correlation coefficients between snow depth and  $\Delta T$ , the  
 4 difference between soil temperature at 20 cm depth and air temperature for winter 1980-2000.  
 5 Regions with greater than 95% significance are hashed.



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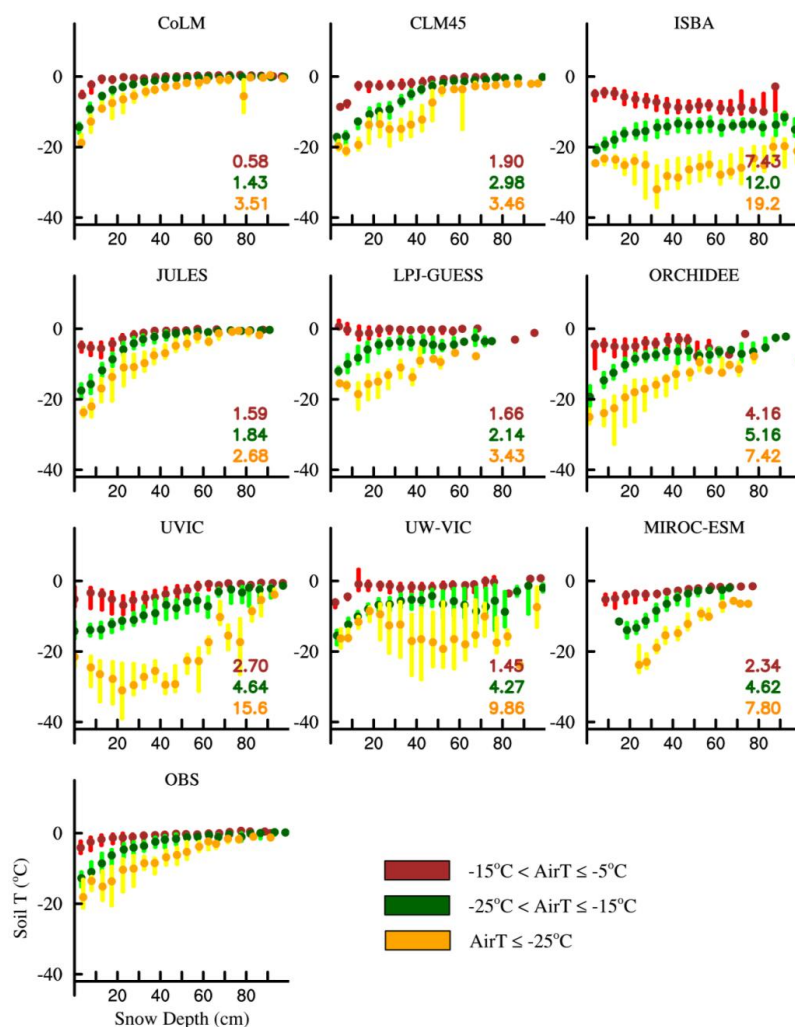
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3 **Figure 4.** Variation of soil temperature at 20 cm depth (°C) with air temperature (°C) for  
 4 winter 1980-2000. The dots represent the medians of 5 °C air temperature bins and the upper  
 5 and lower bars indicate the 25<sup>th</sup> and 75<sup>th</sup> percentiles, calculated from all Russian station grid  
 6 points (n=268) and 21 individual winters. The numbers in each model panel indicate the  
 7 RMSE between the observed and modeled relationship. Color represents different snow depth  
 8 regimes.





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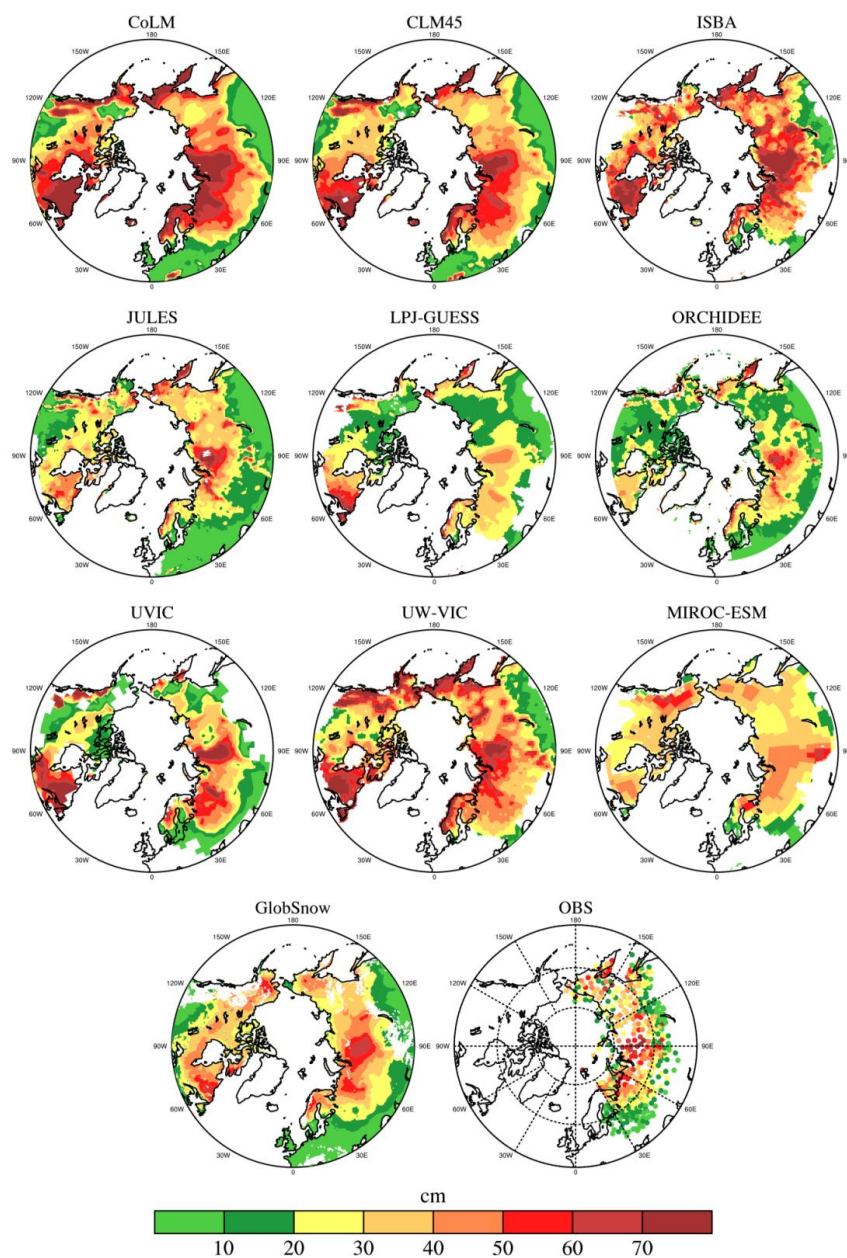
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3 **Figure 5.** Variation of soil temperature at 20 cm depth (°C; y axis) with snow depth (cm) for  
 4 winter 1980-2000. The dots represent the medians of 5 cm snow depth bins and the upper and  
 5 lower bars indicate the 25<sup>th</sup> and 75<sup>th</sup> percentiles, calculated from all Russian station grid points  
 6 (n=268) and 21 individual winters. The numbers in each model panel indicate the RMSE  
 7 between the observed and modeled relationship. Color represents different air temperature  
 8 regimes.





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 3 **Figure 6.** Spatial maps of snow depth (cm) for winter 1980-2000.