Editor's message:

After reading your revised manuscript, I realized you all have done a hard job. However, I agree with the two reviewers' comments that the manuscript needs further revision. The authors need to provide more details on differences between measurements and modeling results as the reviewer #1 pointed out.

Thank you for considering all the work behind the additions we made for the revised version of the MS. We fully agree on this comment by reviewer #1 and improve the MS again by restructuring results and discussions and by including additional plots showing in more detail the comparison of model results to observations.

The authors need to address each review questions thoroughly. Both reviewers pointed out AGAIN that snow modeling is a key issue, the authors need to pay more effort on these questions.

As you can see below and in the revised MS, we have payed additional attention to only this point and advanced the MS in this context. We have extended the introduction with a snow related paragraph, discussing the snow insulation effects on soil temperature:

"For the cold regions, one of the most important factors modifying soil temperature range is the surface snow cover. As discussed in many previous studies (Zhang, 2005; Koven et al., 2013; Scherler et al., 2013; Marmy et al., 2013; Langer et al, 2013; Boike et al., 2003; Gubler et al., 2013; Fiddes et al., 2013), snow dynamics are quite complex and its insulation effects can be extremely important for the soil thermal regime. Model representations of snow cover are lacking many fine-scale processes such as snow ablation, depth hoar formation, snow metamorphism, wind effects on snow distribution and explicit heat and water transfer within snow layers. These issues bring additional uncertainties to global projections."

More references to the effect of snow modelling are also mentioned in the Discussion section, with explanations of the surface temperatures and soil thermal regimes.

Both reviewers also pointed out that the structure and English need further work, I hope that this second review will really make all authors pay effort on this issue.

We thank the editor and reviewers for this suggestion. The MS has been edited again by a native speaking co-author. We hope that one can read it nicely now.

Non-public comments to the Author:
I realized that this is a piece of hard work, but the authors need to follow the standard and face reviewers' comments and sugestions. The overall rating is between "good" and "fair", which is quite low. Some key issues are not well answered and replied. I hope the manuscript will be accepted and published in

The Cryosphere after next revision.

We hope that this revision will be satisfactory.

Report 1

I read and reviewed the revised manuscript again. I feel they addressed my concerns reasonably well. I am glad that they provided some high quality figures I asked for.

Thanks again for providing these comments, which helped to improve the MS.

Since the paper already has 13 large figures, it may not be a good idea to add more figures. However, I still feel these figures contain primary information, which readers may like to see. They may be included as supplementary material. Two sites have multiple years. For clarity, authors may enlarge one or two years to show their differences.

We fully agree with the reviewer's comment and now provide timeseries plots of soil temperature in the supplementary material.

I also share some of the comments of another reviewer. Although the impacts of snow, moss and peat on ground temperature are well known, land surface models still did not treat them well. Such a site-specific reality check would be more convincing than spatial comparisons.

The idea of having several sites is to see, how different models perform in contrasting cold-regions ecosystems. However, we fully agree with the comment. See also our response to the other reviewer. In general, much more detailed model experiments could be done at each site in order to separate effects of different processes on the heat balance. However, as you already pointed out, it is a lot of more material and out of scope of this model intercomparison paper.

The text part seems short of readers' expectation (e.g., what really is the reason for the discrepancy?). The authors may have the same question as well, although they do not want to or unable to involve in too much details of the model schemes. In addition, the authors cannot be too criticizing of individual models. With that practical difficulty, I feel the text is reasonable although it is not fun to read. At least the figures are quite informative.

We agree with the reviewer that in general the reader likes to see more reasons behind differences in model results. We have addressed many discrepancies and tried to explain the reasons by attributing them to different snow modeling schemes, snow biases, site-specific conditions and soil internal functions. However, there are still many unknowns regarding the model mismatches. Thanks a lot for your understanding of the complications behind, which are evident in almost all model intercomparison papers.

The authors can do a bit of improvement in writing so it is more interesting to read. An efficient way probably is to ask a co-author to do a revision and polishing.

We thank the reviewer for this suggestion. We have improved the language of the revised version and also restructured the MS by separating results and discussions.

Report 2

The revision has certainly enhanced the paper. First, I want to give some sentences on the authors' comments:

Comment 5 about structure: The authors keep their structure, this is ok, but still this results in redundancies and other issues. In the end, it is up to the authors to evaluate this.

We agree with the reviewer and now separated results and discussions in a restructured MS.

Comment 7: What I wanted to say is that you can delete the ALT comparison to equilibrium models. You are not using a (semi)-empirical approach, so it makes no sense to compare your transient modelling with this. We are all aware of the advantages of transient modelling in terms of modelling ALT more precisely. My point is that the two schemes are virtually not comparable, and therefore it is not necessary in your paper.

We see the point of the reviewer and we also agree that it makes no sense to advance the paper by an additional comparison to equilibrium model results. These models are useful to obtain a first approximation of climate-driven gradients of ALT. However, our study clarifies some important shortcomings of equilibrium models when it comes to future projections, e.g. ice and water content and soil properties in the subsoil as well as permafrost temperature which can lead e.g. to lag effects. Therefore, we decided to mention this briefly in the beginning, so moved the following section from discussion to introduction:

"Also active layer thickness (ALT) estimates have improved in the current model versions. Simple relationships between surface temperature and ALT have been used in the early modeling studies (Lunardini, 1981; Kudryatsev et al., 1974; Romanovsky and Osterkamp, 1997; Shiklomanov and Nelson, 1999; Stendel et al., 2007, Anisimov et al., 1997). These approaches assume an equilibrium condition, whereas a transient numerical method is better suited within a climate change context. A good review of widely used analytical approximations and differences to numerical approaches is given by Riseborough et al. (2008). With the advanced soil physics in many models, these transient approaches are more widely used especially in long-term simulations. Such improvements highlight the need for an updated assessment of model performances in representing high latitude/altitude soil thermal dynamics."

Comment 9: We know what "spin-up" is. But your explanation is interesting. Of

course you need longer spin-up if the model soil layers is deeper. However, there is absolutely no correlation between spin-up time and depth of model domain. E.g. Orchidee has 43 m domain and 10000 years spin up. "Overkill"? You have 500 years for LPJ (2 m domain) and 10 year COUP (10 m domain), for the same domain you have 50 years for JSBACH. So, no relation. Maybe there are other reason, then you should give them.

We are sorry about the confusion from the too short description in the previous version. The aim of the spin-up procedure is to bring the state variables in equilibrium with forcing data that represent the environment before the transient period starts. Therefore, the spin-up duration depends on process formulations, and discretization schemes like soil layering. Hence, we cannot standardize this but the spin-up procedure can be seen as part of the individual model itself.
Then, this paper is about physical state variables and processes. Temperature, moisture, freezing/thawing, heat conduction, and the respective state variables will typically reach equilibrium after a few decades. However, most of the models also include a carbon cycle, and carbon pools need usually much longer to reach equilibrium. Some of the modeling groups used the existing long spin-up procedure for the biogeochemical models instead of shorter periods because that was faster than a new technical implementation, and the aim of the spin-up did not change.

We have included the following sentences into the manuscript:

"To bring the state variables into equilibrium with climate, models are spun up with climate forcing. Spin-up procedure is part of the model structure, in some cases a full biogeochemical and physical spin up is implemented, whereas in some models a simpler physical spin up is possible. This brings different requirements for the spin up time length, so each model was independently spun-up depending on its model formulations and discretization scheme and the details are given in Table 4. "

Regarding all these discussion, if the reviewers think that Table 4 is causing some confusion, we will gladly remove it.

Still a major comment is the following: I cannot see any comment on the last paragraph of my first review:

"(...) One improvement could be to evaluate really the effect of the different uncertainties, so what is the isolated effect of the snow scheme for the ground temperatures, similar to the effect of the soil scheme. Which of these effects should be improved, maybe a complicated soil scheme is not that important than a more sophisticated snow scheme. Maybe in different climate settings the relative effect is different. I could imagine that a good soil scheme is more important in sedimentary tundra environments than in barren ground and rock? Maybe the snow representation can be more simple in environment A than B? Such questions would certainly increase the merit of this study as they would give useful recommendations. "

My point is that you in in general do not evaluate systematically the performance of the models, and this is related how you designed the study.

You are testing three elements:
1. How is the snow build-up and melt modelled?
2. How is the thermal effect of snow modelled?
3. How is the soil thermal regime modelled?

As far as I can see in your study design, you force the model mostly with the same observations/data for each site. This is fine. Then you run the model and look on the performance. Of course, there will be deviations, starting with that the models in a varying degree manage to re-produce snow cover. This is the first deviation. Then you model the snow (thermal) processes, fine, and here deviations can be attributed to the snow model formulation and of course how snow development is modeled. To isolated look on the performance of the snow scheme, the model should be forced with the snow observations, at least with the same series. The same applies for the sub-soil scheme, the deviations you show are caused by how good the snow is modeled, not how the soil model scheme is formulated. To evaluate the latter, you should force the model with the same ground surface temperature series (under the snow or without snow), for instance. Then deviations are due to the model formulation. Maybe, this is impossible to achieve with you models, I do not know. But then you should comment on that at least.

We fully agree with the reviewer and appreciate your deep thinking for improving the MS. In general, such factorial design would be ideal in order to isolate the effects of certain processes on the soil temperature regime. This is true even when the practical implementation cannot be done as suggested for process-oriented land surface schemes. For example, snow depth is an emergent result of many interacting processes including the interaction of heat balance (snow insulation) and snowmelt (temperature dynamics). It is not easily possible to force such model with snow depth or surface temperature and interpret results of factorial experiments. Nevertheless, experiments with having several processes represented in a cascade could be done, and the first author has done such study in his recent PhD thesis using only the JSBACH model (see Fig. 1 below). These results are documented in a separate section of the thesis, which should go into a separate paper. For a model intercomparison study such experiments and all the results are far too much.

However, in order to show the isolated effects of the soil internal processes on soil temperature dynamics, we now provide a comparison of topsoil and subsoil temperatures for each model and site observation in Figure 2 below. With this figure, one can comment solely on soil internal processes without any snow insulation factor. From this figure, it is seen that models have different rates of transferring heat between top- and sub-soil layers. As an example, HYBRID8 and LPJ-GUESS are mentioned to have faster heat transfer rates in the MS. This can be verified from Figure 2 here, as these two model values are closer to the 1:1 line especially for the snow periods of Nuuk, Schilthorn and Bayelva. This means their soil scheme is transferring heat faster than the other models. However,

during snow-free periods in general and at Samoylov, where the soil profile is more moist than other sites, this pattern is less visible due to model internal functions on how they couple heat and hydrology schemes. If the reviewer wishes, we can include this figure in the supplementary material of the MS.

We have added the following text at the beginning of section 4.2:
"The ideal way to assess the soil internal process would be to use the same snow forcing or under snow temperature for all models, however most of the land models used in this study are not that modular. Hence, intertwined effects of surface and soil internal processes need to be discussed together here."

So, in my opinion, your paper is certainly hard work, to a certain degree also valuable for the community, especially those developing and applying ESM to permafrost etc.

We thank the reviewer for all the constructive ideas and hope that our MS will support future model development works.

Other comments:

p 13, l 10: Schilthorn site. This is a steep mountain site, so there are certainly various 3D effects influencing the ground temperature profile. Should be mentioned at least.

The potential 3D effect was included and clarified for this site – the main part of this discussion was moved to the "Discussion" chapter (see editor and reviewer comments above). The following sentences are added to section 4.2:

"In reality, there are almost isothermal conditions of about -0.7°C between 7 m and at least 100 m depth at this site (PERMOS, 2013), which are partly caused by the 3-dimensional thermal effects due to steep topography (Noetzli et al. 2008). Temperatures near the surface will not be strongly affected by 3-dimensional effects, as the monitoring station is situated on a small but flat plateau (Scherler et al. 2013), but larger depths get additional heat input from the opposite southern slope, causing slightly warmer temperatures at depth than for completely flat topography (Noetzli et al. 2008)."

p. 14, l 1..: I think for decadal effects, you need certainly more than 30 m.

We agree with the reviewer, however we wanted to cite the Alexeev et al. (2007) paper, where he specifically mentions 30 m.

p. 15, l 31: See above, delete the whole paragraph, I do not understand the point.

Thanks again for this point. We agree that our MS did not specifically investigate this ALT estimation topic. So we have removed the paragraph and mentioned the point in the introduction (see above response to comment 7).

p. 16, l. 30: In your study your virtually do not show point 7, so delete.

We also agree on this point and removed the conclusion point from the MS.

Figures and Tables

Table 1: Selected depths of observed and modeled soil temperatures referred as "topsoil temperature" in Figure 1.

	Nuuk	Schilthorn	Samoylov	Bayelva
OBSERVATION	5 cm	20 cm	6 cm	6 cm
JSBACH	3.25 cm	18.5 cm	3.25 cm	3.25 cm
ORCHIDEE	6.5 cm	18.5 cm	6.5 cm	6.5 cm
JULES	5 cm	22.5 cm	5 cm	5 cm
COUP	5.5 cm	20 cm	2.5 cm	5.5 cm
HYBRID8	3.5 cm	22 cm	3.5 cm	3.5 cm
LPJ-GUESS	5 cm	25 cm	5 cm	5 cm

Table 2: Selected depths of observed and modeled soil temperatures referred as "subsoil temperature" in Figure 1.

	Nuuk	Schilthorn	Samoylov	Bayelva
OBSERVATION	30 cm	200 cm	71 cm	112 cm
JSBACH	268 cm	268 cm	268 cm	268 cm
ORCHIDEE	217 cm	217 cm	217 cm	217 cm
JULES	150 cm	150 cm	150 cm	150 cm
COUP	200 cm	200 cm	200 cm	200 cm
HYBRID8	215 cm	220 cm	355 cm	267 cm
LPJ-GUESS	195 cm	195 cm	195 cm	195 cm
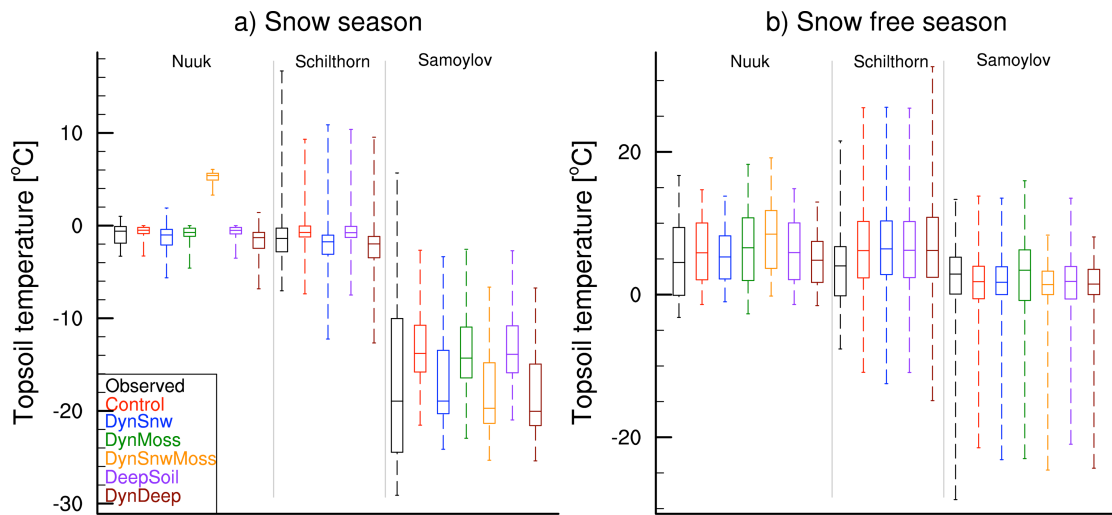
Figure 1: Factorial process experiments with JSBACH model at three different sites. Observed temperatures and the default 10m model version (Control) with constant snow and moss heat transfer parameters are compared to model versions with dynamically changing snow heat parameters (DynSnw), dynamically changing moss heat parameters (DynMoss), dynamic snow and moss parameters (DynSnwMoss), 50 m soil depth (DeepSoil), 50 m soil depth with dynamic snow and moss parameters (DynDeep). This comparison is included in the PhD thesis by Altug Ekici.
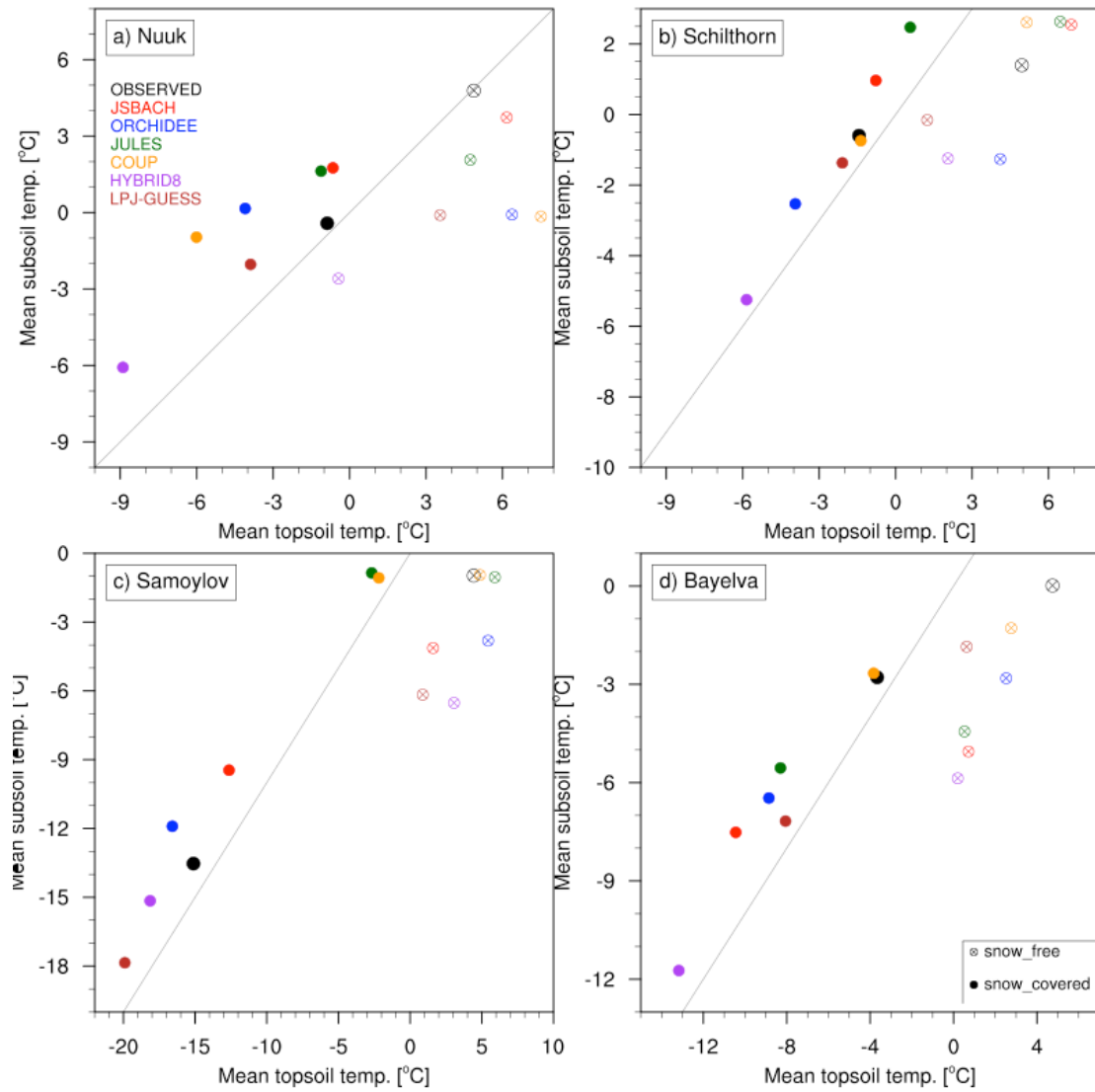
Figure 2: Scatter plots showing topsoil/subsoil temperature relation from observations and models at each site for snow and snow-free seasons. Snow season is defined separately for observations and each model, by taking snow depth values over 5 cm to represent the snow-covered period (observed snow season is used for HYBRID8). The average temperature of all snow covered (or snow free) days of the simulation period is used in the plots. Markers distinguish snow and snow free seasons and colors distinguish models. Gray lines represent the 1:1 line. See Tables 1-2 for the exact soil depths used in this plot.

Changes to the manuscript

The manuscript has been improved substantially. We have
- separated the Results and Discussion sections,
- improved the language with the help of a native speaking coauthor,
- shortened the text in Discussion,
- moved some paragraphs from Discussion to the Introduction,
- shortened the Conclusions section,
- included a Supplementary Material section.

Please see the following marked up version for the full list of changes.

Site-level model intercomparison of high latitude and high altitude soil thermal dynamics in tundra and barren landscapes

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Abstract

Modeling soil thermal dynamics at high latitudes and altitudes requires representations of physical processes such as snow insulation, soil freezing and thawing, and subsurface conditions like soil water/ice content and soil texture. We have compared six different land models: JSBACH, ORCHIDEE, JULES, COUP, HYBRID8 and LPJ-GUESS at four different sites with distinct cold region landscape types to identify the importance of physical processes in capturing observed temperature dynamics in soils. The sites include alpine, high Arctic, wet polygonal tundra and non-permafrost Arctic, thus showing how a range of models can represent distinct soil temperature regimes. For all sites, snow insulation is of major importance for estimating topsoil conditions. However, soil physics is essential for the subsoil temperature dynamics and thus the active layer

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thicknesses. This analysis shows that land models need more realistic surface processes, such as detailed snow dynamics and moss cover with changing thickness and wetness, along with better representations of subsoil thermal dynamics.

1   Introduction

Recent atmospheric warming trends are affecting terrestrial systems by increasing soil temperatures and causing changes in the hydrological cycle. Especially in high latitudes and altitudes, clear signs of change have been observed (Serreze et al., 2000; ACIA, 2005; IPCC AR5, 2013). These relatively colder regions are characterized by the frozen state of terrestrial water, which brings additional risks associated with shifting soils into an unfrozen state. Such changes will have broad implications for the physical (Romanovsky, 2010), biogeochemical (Schuur et al., 2008) and structural (Larsen et al., 2008) conditions of the local, regional and global climate system. Therefore, predicting the future state of the soil thermal regime at high latitudes and altitudes holds major importance for Earth system modeling.

There are increasing concerns as to how land models perform at capturing high latitude soil thermal dynamics, in particular in permafrost regions. Recent studies (Koven et al., 2013; Slater and Lawrence, 2013) have provided detailed assessments of commonly used Earth System Models (ESMs) in simulating soil temperatures of present and future state of the Arctic. By using the Coupled Model Intercomparison Project phase 5 - CMIP5 (Taylor et al., 2009) results, Koven et al. (2013) have shown a broad range of model outputs in simulated soil temperature. They attributed most of the inter-model discrepancies to air-land surface coupling and snow representations in the models. Similar to those findings, Slater and Lawrence (2013) confirmed the high uncertainty of CMIP5 models in predicting the permafrost state and its future trajectories. They concluded that these model versions are not appropriate for such experiments, since they lack critical processes for cold region soils. Snow insulation, land model physics, and vertical model resolutions were identified as the major sources of uncertainty.

For the cold regions, one of the most important factors modifying soil temperature range is the surface snow cover. As discussed in many previous studies (Zhang, 2005; Koven et al., 2013; Scherler et al., 2013; Marmy et al., 2013; Langer et al, 2013; Boike et al., 2003; Gubler et al., 2013; Fiddes et al., 2013), snow dynamics are quite complex and its insulation effects can be extremely important for the soil thermal regime. Model representations of snow cover are lacking many fine-scale processes such as snow ablation, depth hoar formation, snow metamorphism, wind effects on snow distribution and explicit heat and water transfer within snow layers. These issues bring additional uncertainties to global projections.

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1 Current land surface schemes, and most vegetation and soil models, represent energy and mass
2 exchange between the land surface and atmosphere in one dimension. Using a grid cell approach,
3 such exchanges are estimated for the entire land surface or specific regions. However, comparing
4 simulated and observed time series of states or fluxes at point scale rather than grid averaging is an
5 important component of model evaluation, for understanding remaining limitations of models
6 (Ekici et al., 2014; Mahecha et al., 2010). In such "site-level runs", we assume that lateral processes
7 can be ignored and that the ground thermal dynamics are mainly controlled by vertical processes.
8 Then, models are driven by observed climate and variables of interest can be compared to
9 observations at different temporal scales. Even though such idealized field conditions never exist, a
10 careful interpretation of site-level runs can identify major gaps in process representations in models.
11 In recent years, land models have improved their representations of the soil physical environment in
12 cold regions. Model enhancements include the addition of soil freezing and thawing, detailed snow
13 representations, prescribed moss cover, extended soil columns, and coupling of soil heat transfer
14 with hydrology (Ekici et al., 2014; Gouttevin et al., 2012a; Dankers et al., 2011; Lawrence et al.,
15 2008; Wania et al., 2009a). Also active layer thickness (ALT) estimates have improved in the
16 current model versions. Simple relationships between surface temperature and ALT have been used
17 in the early modeling studies (Lunardini, 1981; Kudryatsev et al., 1974; Romanovsky and
18 Osterkamp, 1997; Shiklomanov and Nelson, 1999; Stendel et al., 2007, Anisimov et al., 1997).
19 These approaches assume an equilibrium condition, whereas a transient numerical method is better
20 suited within a climate change context. A good review of widely used analytical approximations
21 and differences to numerical approaches is given by Riseborough et al. (2008). With the advanced
22 soil physics in many models, these transient approaches are more widely used especially in long-
23 term simulations. Such improvements highlight the need for an updated assessment of model
24 performances in representing high latitude/altitude soil thermal dynamics.
25 We have compared the performances of six different land models in simulating soil thermal
26 dynamics at four contrasting sites. In contrast to previous work (Koven et al., 2013; Slater and
27 Lawrence, 2013), we used advanced model versions specifically improved for cold regions and our
28 model simulations are driven by (and evaluated with) site observations. To represent a wider range
29 of assessment and model structures, we used both land components of ESMs (JSBACH,
30 ORCHIDEE, JULES) and stand-alone models (COUP, HYBRID8, LPJ-GUESS), and compared
31 them at Arctic permafrost, Alpine permafrost and Arctic non-permafrost sites. By doing so, we
32 aimed to quantify the importance of different processes, to determine the general shortcomings of
33 current model versions and finally to highlight the key processes for future model developments.
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2 Methods

2.1 Model descriptions

2.1.1 JSBACH

Jena Scheme for Biosphere-Atmosphere Coupling in Hamburg (JSBACH) is the land surface component of the Max Planck Institute Earth System Model (MPI-ESM), which comprises ECHAM6 for the atmosphere (Stevens et al., 2012) and MPIOM for the ocean (Jungclaus et al., 2013). JSBACH provides the land surface boundary for the atmosphere in coupled simulations; however, it can also be used offline driven by atmospheric forcing. The current version of JSBACH (Ekici et al., 2014) employs soil heat transfer coupled to hydrology with freezing and thawing processes included. The soil model is discretized as five layers with increasing thicknesses up to 10 meters depth. There are up to 5 snow layers with constant density and heat transfer parameters. JSBACH also simulates a simple moss/organic matter insulation layer again with constant parameters.

2.1.2 ORCHIDEE

ORCHIDEE is a global land surface model, which can be used coupled to the Institut Pierre Simon Laplace (IPSL) climate model or driven offline by prescribed atmospheric forcing (Krinner et al., 2005). ORCHIDEE computes all the soil-atmosphere-vegetation relevant energy and water exchange processes in 30-minute time steps. It combines a soil-vegetation-atmosphere transfer model with a carbon cycle module, computing vertically detailed soil carbon dynamics. The high latitude version of ORCHIDEE includes a dynamic three-layer snow module (Wang et al., 2013), soil freeze-thaw processes (Gouttevin et al., 2012a), and a vertical permafrost soil thermal and carbon module (Koven et al., 2011). The soil hydrology is vertically discretized as 11 numerical nodes with 2m depth (Gouttevin et al., 2012a), and soil thermal and carbon modules are vertically discretized as 32 layers with ~47m depth (Koven et al., 2011). A one-dimensional Fourier equation was applied to calculate soil thermal dynamics, and both soil thermal conductivity and heat capacity are functions of the frozen and unfrozen soil water content and of dry and saturated soil thermal properties (Gouttevin et al., 2012b).

2.1.3 JULES

JULES (Joint UK Land Environment Simulator) is the land-surface scheme used in the Hadley Centre climate model (Best et al., 2011; Clark et al., 2011), which can also be run offline, driven by atmospheric forcing data. It is based on the Met Office Surface Exchange Scheme, MOSES (Cox et al., 1999). JULES simulates surface exchange, vegetation dynamics and soil physical processes. It can be run at a single point, or as a set of points representing a 2D grid. In each grid cell, the surface is tiled into different surface types, and the soil is treated as a single column, discretized vertically into layers (4 in the standard set-up). JULES simulates fluxes of moisture and energy between the

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atmosphere, surface and soil, and the soil freezing and thawing. It includes a carbon cycle that can simulate carbon exchange between the atmosphere, vegetation and soil. It also includes a multi-layer snow model (Best et al., 2011), with layers that have variable thickness, density and thermal properties. The snow scheme significantly improves the soil thermal regime in comparison with the old, single-layer scheme (Burke et al., 2013). The model can be run with a timestep of between 30 minutes and 3 hours, depending on user preference.

2.1.4 COUP

COUP is a stand-alone, one-dimensional heat and mass transfer model for the soil–snow–atmosphere system (Jansson and Karlberg, 2011) and is capable of simulating transient hydrothermal processes in the subsurface including seasonal or perennial frozen ground (see e.g. Hollesen et al. 2011; Scherler et al., 2010, 2013). Two coupled partial differential equations for water and heat flow are the core of the COUP Model. They are calculated over up to 50 vertical layers of arbitrary depth. Processes that are important for permafrost simulations, such as freezing and thawing of the soil as well as the accumulation, metamorphosis, and melt of a snow cover are included in the model (Lundin, 1990, Gustafsson et al., 2001). Freezing processes in the soil are based on a function of freezing point depression and on an analogy of freezing-thawing and wetting-drying (Harlan, 1973; Jansson and Karlberg, 2011). Snow cover is simulated as one layer of variable height, density, and water content.

The upper boundary condition is given by a surface energy balance at the soil–snow–atmosphere boundary layer, driven by climatic variables. The lower boundary condition at the bottom of the soil column is usually given by the geothermal heat flux (or zero heat flux) and a seepage flow of percolating water. Water transfer in the soil depends on texture, porosity, water, and ice content. Bypass flow through macropores, lateral runoff and rapid lateral drainage due to steep terrain can also be considered (e.g. Scherler et al. 2013). A detailed description of the model including all its equations and parameters is given in Jansson and Karlberg (2011) and Jansson (2012).

2.1.5 HYBRID8

HYBRID8 is a stand-alone land surface model, which computes the carbon and water cycling within the biosphere and between the biosphere and atmosphere. It is driven by the daily/sub-daily climate variables above the canopy, and the atmospheric CO_2 concentration. Computations are performed on a 30-minute timestep for the energy fluxes, and exchanges of carbon and water with the atmosphere and the soil. Litter production and soil decomposition are calculated at a daily timestep. HYBRID8 uses the surface physics and the latest parameterization of turbulent surface fluxes from the GISS ModelE (Schmidt et al., 2006), but has no representation of vegetation dynamics. The snow dynamics from modelE are also not yet fully incorporated. Heat dynamics are described in Rosenzweig et al. (1997) and moisture dynamics in Abramopoulos et al. (1998).

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In HYBRID8 the prognostic variable for the heat transfer is the heat in the different soil layers, and from that the model evaluates the soil temperature. The processes governing this are diffusion from the surface to the sub-surface layers, and conduction and advection between the soil layers. The bottom boundary layer in HYBRID8 is impermeable, resulting in zero heat flux from the soil layers below. The version used in this project has no representation of the snow dynamics and has no insulating vegetation cover. However, the canopy provides a simple heat buffer due its separate heat capacity calculations.

2.1.6 LPJ-GUESSS

Lund-Potsdam-Jena General Ecosystem Simulator (LPJ-GUESS) is a process-based model of vegetation dynamics and biogeochemistry optimized for regional and global applications (Smith et al., 2001). Mechanistic representations of biophysical and biogeochemical processes are shared with those in the Lund-Potsdam-Jena dynamic global vegetation model LPJ-DGVM (Sitch et al. 2003; Gerten et al. 2004). However, LPJ-GUESS replaces the large area parameterization scheme in LPJ-DGVM, whereby vegetation is averaged out over a larger area, allowing several state variables to be calculated in a simpler and faster manner, with more robust and mechanistic schemes of individual- and patch-based resource competition and woody plant population dynamics. Detailed descriptions are given by Smith et al. (2001), Sitch et al. (2003), Wolf et al., (2008), Miller and Smith (2012), and Zhang et al. (2013).

LPJ-GUESS has recently been updated to simulate Arctic upland and peatland ecosystems (McGuire et al., 2012; Zhang et al., 2013). It shares the numerical soil thawing-freezing processes, peatland hydrology and the model of wetland methane emission with LPJ-DGVM WHyMe, as described by Wania et al. (2009a, 2009b, 2010). To simulate soil temperatures and active layer depths, the soil column in LPJ-GUESS is divided into a single snow layer of fixed density and variable thickness, a litter layer of fixed thickness (10 cm for these simulations, except for Schilthorn where it is set to 2.5 cm), a soil column of 2 m depth (with sublayers of thickness 0.1 m, each with a prescribed fraction of mineral and organic material, but with fractions of soil water and air that are updated daily), and finally a "padding" column of depth 48 m (with thicker sublayers), to simulate soil thermal dynamics. Insulation effects of snow, phase changes in soil water, daily precipitation input and air temperature forcing are important determinants of daily soil temperature dynamics at different sub-layers.

2.2 Study sites

2.2.1 Nuuk

The Nuuk observational site is located in southwestern Greenland. The site is situated in a valley in Kobbefjord at 500 m altitude above sea level, and ambient conditions show Arctic climate properties, with a mean annual temperature of -1.5 °C in 2008 and -1.3 °C in 2009 (Jensen and

6

Rasch, 2009, 2010). Vegetation types consist of *Empetrum nigrum* with *Betula nana* and *Ledum groenlandicum*, with a vegetation height of 3-5 cm. The study site soil lacks mineral soil horizons due to cryoturbation and lack of podsol development, as it is situated in a dry location. The soil is composed of 43% sand, 34% loam, 13% clay and 10% organic materials. No soil ice or permafrost formations have been observed within the drainage basin. Snow cover is measured at the Climate Basic station, 1.65 km from the soil station but at the same altitude. At the time of the annual Nuuk Basic snow survey in mid-April, the snow depth at the soil station was very similar to the snow depth at the Climate Basic station: +/- 0.1 meter when the snow depth is high (near 1 meter). Strong winds (>20 m/s) have a strong influence on the redistribution of newly fallen snow, especially in the beginning of the snow season, so the formation of a permanent snow cover at the soil station can be delayed as much as one week, while the end of the snow cover season is similar to that at the Climate Basic station (Birger Ulf Hansen, personal communication, 2013).

2.2.2 Schilthorn

The Schilthorn massif (Bernese Alps, Switzerland) is situated at 2970m altitude in the north central part of the European Alps. Its non-vegetated lithology is dominated by deeply weathered limestone schists, forming a surface layer of mainly sandy and gravelly debris up to 5m thick, which lies over presumably strongly-jointed bedrock. Following the first indications of permafrost (ice lenses) during the construction of the summit station between 1965 and 1967, the site was chosen for long-term permafrost observation within the framework of the European PACE project and consequently integrated into the Swiss permafrost monitoring network PERMOS as one of its reference sites (PERMOS, 2013).

The measurements at the monitoring station at 2900m altitude are located on a flat plateau on the north-facing slope and comprise a meteorological station and three boreholes (14m vertical, 100m vertical and 100m inclined), with continuous ground temperature measurements since 1999 (Vonder Mühll et al., 2000; Hoelzle and Gruber, 2008; Harris et al., 2009). Borehole data indicate permafrost of at least 100m thickness, which is characterized by ice-poor conditions close to the melting point. Maximum active-layer depths recorded since the start of measurements in 1999 are generally around 4-6m, but during the exceptionally warm summer of the year 2003 the active-layer depth increased to 8.6 m, reflecting the potential for degradation of permafrost at this site (Hilbich et al., 2008).

The monitoring station has been complemented by soil moisture measurements since 2007 and geophysical (mainly geoelectrical) monitoring since 1999 (Hauck 2002, Hilbich et al. 2011). The snow cover at Schilthorn can reach maximum depths of about 2-3m and usually lasts from October through to June/July. One dimensional soil model sensitivity studies showed that impacts of long-term atmospheric changes would be strongest in summer and autumn, due to this late snowmelt and

1   the long decoupling of the atmosphere from the surface. So, increasing air temperatures could lead

2   to a severe increase in active-layer thickness (Engelhardt et al. 2010, Marmy et al. 2013, Scherler et

3   al. 2013).

2.2.3 Samoylov

5   Samoylov Island belongs to an alluvial river terrace of the Lena River Delta. The island is elevated

6   about 20 m above the normal river water level and covers an area of about 3.4 km² (Boike et al.

7   2013). The western part of the island constitutes a modern floodplain, which is lowered compared

8   with the rest of the island and is often flooded during ice break-up of the Lena River in spring. The

9   eastern part of the island belongs to the elevated river terrace, which is mainly characterized by

10 moss, and sedge vegetated tundra (Kutzbach et al. 2007). In addition, several lakes and ponds

11 occur, which make up about 25% of the surface area of Samoylov (Muster et al. 2012).

12 The land surface of the island is characterized by the typical micro-relief of polygonal patterned

13 ground, caused by frost cracking and subsequent ice-wedge formation. The polygonal structures

14 usually consist of depressed centers, surrounded by elevated rims, which can be found in a partly or

15 completely collapsed state (Kutzbach et al. 2007). The soil in the polygonal centers usually consists

16 of water-saturated sandy peat, with the water table standing a few centimeters above or below the

17 surface. The elevated rims are usually covered with a dry moss layer, underlain by wet sandy soils,

18 with massive ice wedges underneath. The cryogenic soil complex of the river terrace reaches depths

19 of 10 to 15 m and is underlain by sandy to silty river deposits. These river deposits reach depths of

20 at least 1 km in the delta region (Langer et al. 2013).

21 There are strong spatial differences in surface energy balance due to heterogeneous surface and

22 subsurface properties. Due to thermo-erosion, there is an ongoing expansion of thermokarst lakes

23 and small ponds (Abnizova et al. 2012). Soil water drainage is strongly related to active layer

24 dynamics, with lateral water flow occuring from late summer to autumn (Helbig et al. 2012). Site

25 conditions include strong snow-micro-topography, and snow-vegetation interactions due to wind

26 drift (Boike et al. 2013).

2.2.4 Bayelva

28 The Bayelva climate and soil-monitoring site is located in the Kongsfjord region on the west coast

29 of the Svalbard Island. The North Atlantic Current warms this area to an average air temperature of

30 about −13 °C in January and +5 °C in July, and provides about 400 mm precipitation annually,

31 falling mostly as snow between September and May. The annual mean temperature of 1994 to 2010

32 in the village of Ny-Ålesund has been increasing by +1.3 K per decade (Maturilli et al., 2013). The

33 observation site is located in the Bayelva River catchment on the Brøgger peninsula, about 3 km

34 from Ny-Ålesund. The Bayelva catchment is bordered by two mountains, the Zeppelinfjellet and

35 the Scheteligfjellet, between which the glacial Bayelva River originates from the two branches of

the Brøggerbreen glacier moraine rubble. To the north of the study site, the terrain flattens, and after about 1 km the Bayelva River reaches the shoreline of the Kongsfjorden (Arctic Ocean). In the catchment area, sparse vegetation alternates with exposed soil and sand and rock fields. Typical permafrost features, such as mud boils and non-sorted circles, are found in many parts of the study area. The Bayelva permafrost site itself is located at 25 m a.s.l., on top of the small Leirhaugen hill. The dominant ground pattern at the study site consists of non-sorted soil circles. The bare soil circle centers are about 1 m in diameter and are surrounded by a vegetated rim, consisting of a mixture of low vascular plants of different species of grass and sedges (*Carex spec., Deschampsia spec., Eriophorum spec., Festuca spec., Luzula spec.*), catchfly, saxifrage, willow and some other local common species (*Dryas octopetala, Oxyria digyna, Polegonum viviparum*) and unclassified species of mosses and lichens. The vegetation cover at the measurement site was estimated to be approximately 60%, with the remainder being bare soil with a small proportion of stones. The silty clay soil has a high mineral content, while the organic content is low, with organic fractions below 10% (Boike et al., 2007). In the study period, the permafrost at Leirhaugen hill had a mean annual temperature of about −2 °C at the top of the permafrost at 1.5 m depth.

Over the past decade, the Bayelva catchment has been the focus of intensive investigations into soil and permafrost conditions (Roth and Boike, 2001; Boike et al., 2007; Westermann et al., 2010; Westermann et al., 2011), the winter surface energy balance (Boike et al., 2003), and the annual balance of energy, H_2O and CO_2, and micrometeorological processes controlling these fluxes (Westermann et al. 2009; Lüers et al., 2014).

2.3   Intercomparison set-up and simulation protocol

In order solely to compare model representations of physical processes and to eliminate any other source of uncertainty (e.g. climate forcing, spatial resolution, soil parameters etc.), model simulations were driven by the same atmospheric forcing and soil properties at site-scale. Driving data for all site simulations were prepared and distributed uniformly. Site observations were converted into continuous time series with minor gap filling. Where the observed variable set lacked the variable needed by the models, extended WATCH reanalysis data (Weedon et al., 2010; Beer et al., 2014) was used to complement the data sets. Soil thermal properties are based on the sand, silt, and clay fractions of the Harmonized World Soil database v1.1 (FAO et al., 2009). All model simulations were forced with these datasets. Table 3 summarizes the details of site driving data preparation together with soil static parameters.

To bring the state variables into equilibrium with climate, models are spun up with climate forcing. Spin-up procedure is part of the model structure, in some cases a full biogeochemical and physical spin up is implemented, whereas in some models a simpler physical spin up is possible. This brings

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1 different requirements for the spin up time length, so each model was independently spun-up
2 depending on its model formulations and discretization scheme and the details are given in Table 4.
3 Most of the analysis focuses on the upper part of the soil. The term "topsoil" is used from now on to
4 indicate the chosen upper soil layer in each model, and the first depth of soil temperature
5 observations. The details of layer selection are given in Table A1 of Appendix-A.

6
7 ## 3   Results
8
9 ### 3.1   Topsoil temperature and surface insulation effects
10 As all our study sites are located in cold climate zones (Fig. 1), there is significant seasonality,
11 which necessitates a separate analysis for each season. Figure 2 shows average seasonal topsoil
12 temperature distributions (see Table A1 for layer depths) extracted from the six models, along with
13 the observed values at the four different sites. In this figure, observed and simulated temperatures
14 show a wide range of values depending on site-specific conditions and model formulations.
15 Observations show that during winter and spring Samoylov is much colder than the other sites (Fig.
16 2a, 2b). Observed summer and autumn temperatures are similar at all sites (Fig. 2c, 2d), with Nuuk
17 being the warmest site in general. For the modeled values, the greatest inconsistency with
18 observations is in matching the observed winter temperatures, especially at Samoylov and
19 Schilthorn (Fig. 2a). The modeled temperature range increases in spring (Fig. 2b), and even though
20 the mean modeled temperatures in summer are closer to observed means, the maximum and
21 minimum values show a wide range during this season (Fig. 2c). Autumn, shows a more uniform
22 distribution of modeled temperatures compared with the other seasons (Fig. 2d).
23 A proper assessment of critical processes entails examining seasonal changes in surface cover and
24 the consequent insulation effects for the topsoil temperature. To investigate these effects, Figure 3
25 shows the seasonal relations between air and topsoil temperature at each study site. Air temperature
26 values are the same for all models, as they are driven with the same atmospheric forcing.
27 Observations show that topsoil temperatures are warmer than the air during autumn, winter, and
28 spring at all sites, but the summer conditions are dependent on the site (Fig. 3). In the models,
29 winter topsoil temperatures are warmer than the air in most cases, as observed. However, the
30 models show a wide range of values, especially at Samoylov (Fig. 3c), where the topsoil
31 temperatures differ by up to 25°C between models. In summer, the models do not show consistent
32 relationships between soil and air temperatures, and the model range is highest at the Nuuk and
33 Schilthorn sites.
34 To analyze the difference in modeled and observed snow isolation effect in more detail, Figure 4
35 shows the changes in snow depth from observed and modeled values. Schilthorn has the highest
36 snow depth values (>1.5m), while all other sites have a maximum snow height between 0.5-1 m

(Fig. 4). Compared with observations, the models usually overestimate the snow depth at Schilthorn and Samoylov (Fig 4b, 4c) and underestimate it at Nuuk and Bayelva (Fig. 4a, 4d).

For our study sites, the amount of modeled snow depth bias is correlated with the amount of modeled topsoil temperature bias (Fig. 5). With overestimated (underestimated) snow depth, models generally simulate warmer (colder) topsoil temperatures. As seen in Figure 5a, almost all models underestimate the snow depth at Nuuk and Bayelva, and this creates colder topsoil temperatures. The opposite is seen for Samoylov and Schilthorn, where higher snow depth bias is accompanied by higher topsoil temperature bias (except for ORCHIDEE and LPJ-GUESS models).

As snow can be persistent over spring and summer seasons in cold regions (Fig. 4), it is worthwhile to separate snow and snow-free seasons for these comparisons. Figure 6 shows the same atmosphere/topsoil temperature comparison as in Figure 3 but using individual (for each model and site) snow and snow-free seasons instead of conventional seasons. In this figure, all site observations show a warmer topsoil temperature than air, except for the snow-free season at Samoylov. Models, however, show different patterns at each site. For the snow season, models underestimate the observed values at Nuuk and Bayelva, whereas they overestimate it at Schilthorn and Samoylov except for the previously mentioned ORCHIDEE and LPJ-GUESS models. Modeled snow-free season values, however, do not show consistent patterns.

3.2 Subsurface thermal regime

Assessing soil thermal dynamics necessitates scrutinizing subsoil temperature dynamics as well as surface conditions. Soil temperature evolutions of simulated soil layers are plotted for each model at each site in Fig. 7-10. Strong seasonal temperature changes are observed close to the surface, whereas temperature amplitudes are reduced in deeper layers and eventually a constant temperature is simulated at depths with zero annual amplitude (DZAA).

Although Nuuk is a non-permafrost site, most of the models simulate subzero temperatures below 2-3 meters at this site (Fig. 7). Here, only ORCHIDEE and COUP simulate a true DZAA at around 2.5-3 meters, while all other models show a minor temperature change even at their deepest layers. At the high altitude Schilthorn site (Fig. 8), JSBACH and JULES simulate above 0°C temperatures (non-permafrost conditions) in deeper layers. Compared with other models with snow representation, ORCHIDEE and LPJ-GUESS show colder subsurface temperatures at this site (Fig. 8). The simulated soil thermal regime at Samoylov reflects the colder climate at this site. All models show subzero temperatures below 1 m (Fig. 9). However, compared with other models, JULES and COUP show values much closer to 0°C. At the high-Arctic Bayelva site, all models simulate permafrost conditions (Fig. 10). The JULES and COUP models again show warmer temperature profiles than the other models.

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The soil thermal regime can also be investigated by studying the vertical temperature profiles regarding the annual means (Fig. 11), and minimum and maximum values (Fig. 12). In Figure 11, the distribution of mean values is similar to the analysis of topsoil conditions. The mean subsoil temperature is coldest at Samoylov followed by Bayelva, while Schilthorn is almost at the 0°C boundary (no deep soil temperature data available from Nuuk for this comparison). JSBACH, JULES, and COUP overestimate the temperatures at Schilthorn and Samoylov, but almost all models underestimate it at Bayelva. Figure 12 shows the temperature envelopes of observed and simulated values at each site. The minimum (maximum) temperature curve represents the coldest (warmest) possible conditions for the soil thermal regime at a certain depth. The models agree more on the maximum curve than the minimum curve (Fig. 12), indicating the differences in soil temperature simulation for colder periods. The HYBRID8 model almost always shows the coldest conditions, whereas the pattern of the other models changes depending on the site.

Figure 13 shows the yearly change of ALT for the three permafrost sites. Observations indicate a shallow ALT at Samoylov (Fig. 13b) and very deep ALT for Schilthorn (Fig 13a). All models overestimate the ALT at Samoylov (Fig. 13b), but there is disagreement among models in over- or underestimating the ALT at Schilthorn (Fig. 13a) and Bayelva (Fig. 13c).

4   Discussion

4.1   Topsoil temperature and surface insulation effects

Figure 2 has shown a large range among modeled temperature values, especially during winter and spring. As mentioned in the introduction, modeled mean soil temperatures are strongly related to the atmosphere-surface thermal connection, which is strongly influenced by snow cover and its properties.

Observations show warmer topsoil temperatures than air during autumn, winter, and spring (Fig. 3). This situation indicates that soil is insulated when compared to colder air temperatures. This can be attributed to the snow cover during these seasons (Fig. 4). The insulating property of snow keeps the soil warmer than air, while not having snow can result in colder topsoil temperatures than air (as for the HYBRID8 model, cf. Fig. 3). Even though the high albedo of snow provides a cooling effect for soil, the warming due to insulation dominates during most of the year. Depending on their snow depth bias, models show different relations between air and topsoil temperature. The amount of winter warm bias from snow depth overestimation in models depends on whether the site has a "sub- or supra-critical" snow height. With supra-critical conditions (e.g. at Schilthorn), the snow depth is so high that a small over- or underestimation in the model makes very little difference to the insulation. Only the timing of the snow arrival and melt-out is important. In sub-critical conditions (e.g. at Samoylov), the snow depth is so low that any overestimation leads to a strong

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warm bias in the simulation e.g. for JULES/COUP. This effect is also mentioned in Zhang T. (2005), where it is stated that snow depths of less than 50 cm have the greatest impact on soil temperatures. However, overestimated snow depth at Samoylov and Schilthorn does not always result in warmer soil temperatures in models as expected (Fig. 3b, 3c). At these sites, even though JSBACH, JULES and COUP show warmer soil temperatures in parallel to their snow depth overestimations, ORCHIDEE and LPJ-GUESS show the opposite. This behavior indicates different processes working in opposite ways. Nevertheless, most of the winter, autumn and spring topsoil temperature biases can be explained by snow conditions (Fig. 5a). Figure 5b shows that snow depth bias can explain the topsoil temperature bias even when the snow free season is considered, which is due to the long snow period at these sites (Table 2). This confirms the importance of snow representation in models for capturing topsoil temperatures at high latitudes and high altitudes.

On the other hand, considering dynamic heat transfer parameters (volumetric heat capacity and heat conductivity) in snow representation seems to be of lesser importance (JSBACH vs. other models, see Table 1). This is likely because a greater uncertainty comes from processes that are still missing in the models, such as wind drift, depth hoar formation and snow metamorphism. As an example, the landscape heterogeneity at Samoylov forms different soil thermal profiles for polygon center and rim. While the soil temperature comparisons were performed for the polygon rim, snow depth observations were taken from polygon center. Due to strong wind drift almost all snow is removed from the rim and also limited to ca. 50cm (average polygon height) at the center (Boike et al., 2008). This way, models inevitably overestimate snow depth and insulation, in particular on the rim where soil temperature measurements have been taken. Hence, a resulting winter warm bias is expected (Fig. 2a, models JSBACH, JULES, COUP).

During the snow free season, Samoylov has colder soil temperatures than air (Fig. 6c). Thicker moss cover and higher soil moisture content at Samoylov (Boike et al., 2008) are the reasons for cooler summer topsoil temperatures at this site. Increasing moss thickness changes the heat storage of the moss cover and it acts as a stronger insulator (Gornall et al., 2007), especially when dry (Soudzilovskaia et al., 2013). Additionally, high water content in the soil requires additional input of latent heat for thawing and there is less heat available to warm the soil.

Insulation strength during the snow free season is related to model vegetation/litter layer representations. 10 cm fixed moss cover in JSBACH and a 10 cm litter layer in LPJ-GUESS bring similar amounts of insulation. At Samoylov, where strong vegetation cover is observed in the field, these models perform better for the snow-free season (Fig. 6c). However, at Bayelva, where vegetation effects are not that strong, 10 cm insulating layer proves to be too much and creates colder topsoil temperatures than observations (Fig. 6d). And for the bare Schilthorn site, even a thin

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layer of surface cover (2.5 cm litter layer) creates colder topsoil temperatures in LPJ-GUESS (Fig. 6b).

At Bayelva, all models underestimate the observed topsoil temperatures all year long (Fig. 6d). With underestimated snow depth (Fig. 4d) and winter cold bias in topsoil temperature (Fig. 3d), models create a colder soil thermal profile that results in cooling of the surface from below even during the snow free season. Furthermore, using global reanalysis products instead of site observations (Table 3) might cause biases in incoming longwave radiation, which can also affect the soil temperature calculations. In order to assess model performance in capturing observed soil temperature dynamics, it is important to drive the models with a complete set of site observations. These analyses support the need for better vegetation insulation in models during the snow free season. The spatial heterogeneity of surface vegetation thickness remains an important source of uncertainty. More detailed moss representations were used in Porada et al. (2013) and Rinke et al. (2008), and such approaches can improve the snow free season insulation in models.

4.2   Soil thermal regime

Model differences in representing subsurface temperature dynamics are related to the surface conditions (especially snow) and soil heat transfer formulations. The ideal way to assess the soil internal processes would be to use the same snow forcing or under snow temperature for all models. However most of the land models used in this study are not that modular. Hence, intertwined effects of surface and soil internal processes must be discussed together here.

Figures 7-10 show the mismatch in modeled DZAA representations. Together with the soil water and ice contents, simulating DZAA is partly related to the model soil depth and some models are limited by their shallow depth representations (Fig. A1, Table 1). Apart from the different temperature values, models also simulate permafrost conditions very differently. As seen in Fig. 8, JSBACH and JULES do not simulate permafrost conditions at Schilthorn. In reality, there are almost isothermal conditions of about -0.7°C between 7 m and at least 100 m depth at this site (PERMOS, 2013), which are partly caused by the 3-dimensional thermal effects due to steep topography (Noetzli et al. 2008). Temperatures near the surface will not be strongly affected by 3-dimensional effects, as the monitoring station is situated on a small but flat plateau (Scherler et al. 2013), but larger depths get additional heat input from the opposite southern slope, causing slightly warmer temperatures at depth than for completely flat topography (Noetzli et al. 2008). The warm and isothermal conditions close to the freezing point at Schilthorn mean that a small temperature mismatch (on the order of 1°C) can result in non-permafrost conditions. This kind of temperature bias would not affect the permafrost condition at colder sites (e.g. Samoylov). In addition, having low water and ice content, and a comparatively low albedo, make the Schilthorn site very sensitive to interannual variations and make it more difficult for models to capture the soil thermal dynamics

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1 (Scherler et al., 2013). Compared to the other models with snow representation, ORCHIDEE and

2 LPJ-GUESS show colder subsurface temperatures at this site (Fig. 8). A thin surface litter layer

3 (2.5cm) in LPJ-GUESS contributes to the cooler Schilthorn soil temperatures in summer.

4 Differences at Samoylov are more related to the snow depth biases. As previously mentioned,

5 subcritical snow conditions at this site amplify the soil temperature overestimation coming from

6 snow depth bias (Fig. 5). Considering their better match during snow free season (Fig. 6c), the

7 warmer temperatures in deeper layers of JULES and COUP can be attributed to overestimated snow

8 depths for this site by these two models (Fig. 9). Additionally, JULES and COUP models simulate

9 generally warmer soils conditions than the other models, because these models include heat transfer

10 via advection in addition to heat conduction. Heat transfer by advection of water is an additional

11 heat source for the subsurface in JULES and COUP, which can also be seen in the results for

12 Bayelva (Fig. 10). In combination with that, COUP has a greater snow depth at Samoylov (Fig. 5),

13 resulting in even warmer subsurface conditions than JULES. Such conditions demonstrate the

14 importance of the combined effects of surface processes together with internal soil physics.

15 Due to different heat transfer rates among models, internal soil processes can impede the heat

16 transfer and result in delayed warming or cooling of the deeper layers. JSBACH, ORCHIDEE,

17 JULES and COUP show a more pronounced time lag of the heat/cold penetration into the soil,

18 while HYBRID8 and LPJ-GUESS show either a very small lag or no lag at all (Figs. 7-10). This

19 time lag is affected by the method of heat transfer (e.g. advection and conduction, see above), soil

20 heat transfer parameters (soil heat capacity/conductivity), the amount of simulated phase change,

21 vertical soil model resolution and internal model timestep. Given that all models use some sort of

22 heat transfer method including phase change (Table 1) and similar soil parameters (Table 3), the

23 reason for the rapid warming/cooling at deeper layers of some models can be missing latent heat of

24 phase change, vertical resolution or model timestep. Even though the mineral (dry) heat transfer

25 parameters are shared among models, they are modified afterwards due to the coupling of

26 hydrology and thermal schemes. This leads to changes in the model heat conductivities depending

27 on how much water and ice they simulate in that particular layer. Unfortunately, not all models

28 output soil water and ice contents in a layered structure similar to soil temperature. This makes it

29 difficult to assess the differences in modeled phase change, and the consequent changes to soil heat

30 transfer parameters. A better quantification of heat transfer rates would require a comparison of

31 simulated water contents and soil heat conductivities among models, which is beyond the scope of

32 this paper.

33 The model biases in matching the vertical temperature curves (minimum, maximum, mean) are

34 related to the topsoil temperature bias in each model for each site, but also the above-mentioned soil

35 heat transfer mechanisms and bottom boundary conditions. Obviously, models without snow

15

1 representation (e.g. HYBRID8) cannot match the minimum curve in Fig. 12. However, snow depth

2 bias (Fig. 5) cannot explain the minimum curve mismatch for ORCHIDEE, COUP, and LPJ-

3 GUESS at Schilthorn (Fig. 12b). This highlights the effects of soil heat transfer schemes once

4 again.

5 In general, permafrost specific model experiments require deeper soil representation than 5-10

6 meters. As discussed in Alexeev et al. (2007), more than 30 m soil depth is needed for capturing

7 decadal temperature variations in permafrost soils. The improvements from having such extended

8 soil depth are shown in Lawrence et al. (2012), when compared to their older model version with

9 shallow soil depth (Lawrence and Slater, 2005). Additionally, soil layer discretization plays an

10 important role for the accuracy of heat and water transfer within the soil, and hence can effect the

11 ALT estimations. Most of the model setups in our intercomparison have less than 10 m depths, so

12 they lack some effects of processes within deeper soil layers. However, most of the models used in

13 global climate simulations have similar soil depth representations and the scope here is to compare

14 models that are not only aimed to simulate site-specific permafrost conditions at high resolution but

15 to show general guidelines for future model developments.

16 ## 4.3 Active layer thickness

17 As seen above, surface conditions (e.g. insulation) alone are not enough to explain the soil thermal

18 regime, as subsoil temperatures and soil water and ice contents affect the ALT as well. For

19 Schilthorn, LPJ-GUESS generally shows shallower ALT values than other models (Fig. 13a); it also

20 shows the largest snow depth bias (Fig. 5), excluding snow as a possible cause for this shallow ALT

21 result. However, if snow depth bias alone could explain the ALT difference, ORCHIDEE would

22 show different values than HYBRID8, which completely lacks any snow representation. At

23 Schilthorn, COUP has a high snow depth bias (Fig. 5) but still shows a very good match with the

24 observed ALT (Fig. 13a), mainly because snow cover values at Schilthorn are very high so ALT

25 estimations are insensitive to snow depth biases as long as modeled snow cover is still sufficiently

26 thick to have the full insulation effect (Scherler et al. 2013).

27 All models overestimate the snow depth at Samoylov (Fig. 5) and most of them lack a proper moss

28 insulation (Fig. 6c), which seems to bring deeper ALT estimates in Samoylov (Fig. 13b). However,

29 HYBRID8 does not have snow representation, yet it shows the deepest ALT values, which means

30 lack of snow insulation is not the reason for deeper ALT values in this model. As well as lacking

31 any vegetation insulation, soil heat transfer is also much faster in HYBRID8 (see section 3.2),

32 which allows deeper penetration of summer warming into the soil column.

33 Surface conditions alone cannot describe the ALT bias in Bayelva either. LPJ-GUESS shows the

34 lowest snow depth (Fig. 5) together with deepest ALT (Fig. 13c), while JULES shows similar snow

35 depth bias as LPJ-GUESS but the shallowest ALT values. As seen from Fig. 10, LPJ-GUESS

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Overestimating snow depth (Fig. 5) and lacking the moss insulation (Fig. 6c) seems to match with deeper

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allows deeper heat penetration at this site. So, not only the snow conditions, but also the model's heat transfer rate is critical for correctly simulating the ALT.

5 Conclusions

We have evaluated different land models' soil thermal dynamics against observations using a site-level approach. The analysis of the simulated soil thermal regime clearly reveals the importance of reliable surface insulation for topsoil temperature dynamics and of reliable soil heat transfer formulations for subsoil temperature and permafrost conditions. Our findings include the following conclusions.

1. At high latitudes and altitudes, model snow depth bias explains most of the topsoil temperature biases.
2. The sensitivity of soil temperature to snow insulation depends on site snow conditions (sub-/supra-critical).
3. Surface vegetation cover and litter/organic layer insulation is important for topsoil temperatures in the snow-free season, therefore models need more detailed representation of moss and top organic layers.
4. Model heat transfer rates differ due to coupled heat transfer and hydrological processes. This leads to discrepancies in subsoil thermal dynamics.
5. Surface processes alone cannot explain the whole soil thermal regime; subsoil conditions and model formulations affect the soil thermal dynamics.

For permafrost and cold-region related soil experiments, it is important for models to simulate the soil temperatures accurately, because permafrost extent, active layer thickness and permafrost soil carbon processes are strongly related to soil temperatures. There is major concern about how the soil thermal state of these areas affects the ecosystem functions, and about the mechanisms (physical/biogeochemical) relating atmosphere, oceans and soils in cold regions. With the currently changing climate, the strength of these couplings will be altered, bringing additional uncertainty into future projections.

In this paper, we have shown the current state of a selection of land models with regard to capturing surface and subsurface temperatures in different cold-region landscapes. It is evident that there is much uncertainty, both in model formulations of soil internal physics and especially in surface processes. To achieve better confidence in future simulations, model developments should include better insulation processes (for snow: compaction, metamorphism, depth hoar, wind drift; for moss: dynamic thickness and wetness). Models should also perform more detailed evaluation of their soil heat transfer rates with observed data, for example comparing simulated soil moisture and soil heat conductivities.

Appendix A: Model layering schemes and depths of soil temperature observations

Table A1: Selected depths of observed and modeled soil temperatures referred as "topsoil temperature" in Figures 1, 2, 4, 5 and 6.

	Nuuk	**Schilthorn**	**Samoylov**	**Bayelva**
OBSERVATION	5 cm	20 cm	6 cm	6 cm
JSBACH	3.25 cm	18.5 cm	3.25 cm	3.25 cm
ORCHIDEE	6.5 cm	18.5 cm	6.5 cm	6.5 cm
JULES	5 cm	22.5 cm	5 cm	5 cm
COUP	5.5 cm	20 cm	2.5 cm	5.5 cm
HYBRID8	3.5 cm	22 cm	3.5 cm	3.5 cm
LPJ-GUESS	5 cm	25 cm	5 cm	5 cm

Exact depths of each soil layer used in model formulations:

JSBACH: 0.065, 0.254, 0.913, 2.902, 5.7 m

ORCHIDEE: 0.04, 0.05, 0.06, 0.07, 0.08, 0.1, 0.11, 0.14, 0.16, 0.19, 0.22, 0.27, 0.31, 0.37, 0.43, 0.52, 0.61, 0.72, 0.84, 1.00, 1.17, 1.39, 1.64, 1.93, 2.28, 2.69, 3.17, 3.75, 4.42, 5.22, 6.16, 7.27 m

JULES: 0.1, 0.25, 0.65, 2.0 m

COUP: different for each site

Nuuk: 0.01 m intervals until 0.36 m, then 0.1 m intervals until 2 m and then 0.5 m intervals until 6 m

Schilthorn: 0.05 m then 0.1 m intervals until 7 m, and then 0.5 m intervals until 13 m

Samoylov: 0.05 m then 0.1 m intervals until 5 m, and then 0.5 m intervals until 8 m

Bayelva: 0.01 m intervals until 0.3 m, then 0.1 m intervals until 1 m and then 0.5 m intervals until 6 m

HYBRID8: different for each site

Nuuk: 0.07, 0.29, 1.50, 5.00 m

Schilthorn: 0.07, 0.30, 1.50, 5.23 m

Samoylov: 0.07, 0.30, 1.50, 6.13 m

Bayelva: 0.07, 0.23, 1.50, 5.00 m

LPJ-GUESS: 0.1 m intervals until 2 m (additional padding layer of 48 m depth)

Depths of soil temperature observations for each site:

NUUK: 0.01,0.05,0.10,0.30 m

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1 **SCHILTHORN:**   0.20,0.40,0.80,1.20,1.60,2.00,2.50,3.00,3.50,4.00,5.00,7.00,9.00,10.00 m

2 **SAMOYLOV:**   0.02,0.06,0.11,0.16,0.21,0.27,0.33,0.38,0.51,0.61,0.71 m

3 **BAYELVA:**   0.06,0.24,0.40,0.62,0.76,0.99,1.12 m

depth [m]

JSBACH
ORCHIDEE
JULES
COUP
HYBRID8
LPJ-GUESS
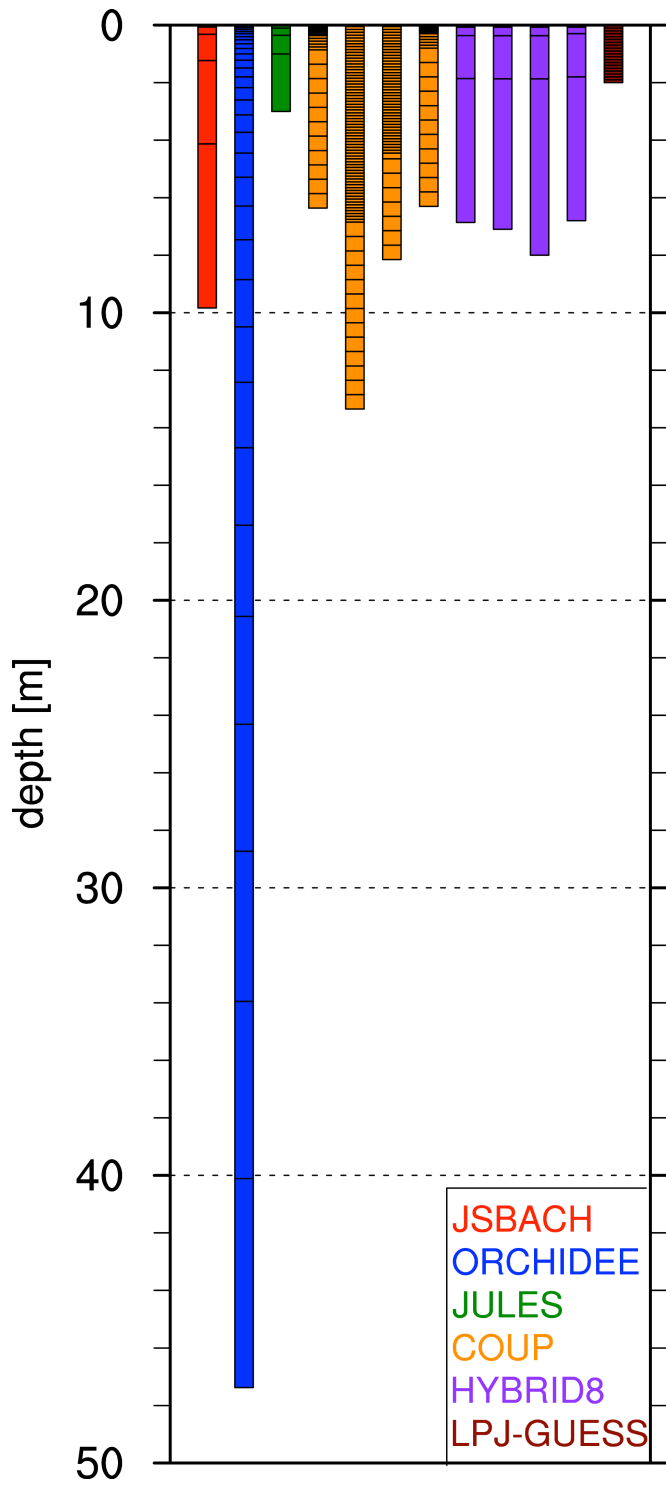
Figure A1: Soil layering schemes of each model. COUP and HYBRID8 models use different layering schemes for each study site, which are represented with different bars (from left to right: Nuuk, Schilthorn, Samoylov and Bayelva).

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1   Table 1: Model details related to soil heat transfer

	JSBACH	ORCHIDEE	JULES	COUP	HYBDRID8	LPJ-GUESS
Soil freezing	Yes	Yes	Yes	Yes	Yes	Yes
Soil heat transfer method	Conduction	Conduction	Conduction Advection	Conduction Advection	Conduction Advection	Conduction
Dynamic soil heat transfer parameters	Yes	Yes	Yes	Yes	Yes	Yes
Soil depth	10m	43m	3m	Variable (>5m)	Variable (>5m)	2m
Bottom boundary condition	Zero heat flux	Geothermal heat flux (0.057 W/m^2)	Zero heat flux	Geothermal heat flux (0.011 W/m^2)	Zero heat flux	Zero heat flux
Snow layering	5 layers	3 layers	3 layers	1 layer	No snow representation	1 layer
Dynamic snow heat transfer parameters	No	Yes	Yes	Yes	-	Yes (only heat capacity)
Insulating vegetation cover	10cm moss layer	-	-	-	-	Site-specific litter layer
Model timestep	30min	30min	30min	30min	30min	1day

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1   Table 2: Site details

	NUUK	SCHILTHORN	SAMOYLOV	BAYELVA
Latitude	64.13° N	46.56° N	72.4° N	78.91° N
Longitude	51.37° W	7.08° E	126.5° E	11.95° E
Mean annual air temperature	-1.3 °C	-2.7 °C	-13 °C	-4.4 °C
Mean annual ground temperature	3.2 °C	-0.45 °C	-10 °C (?)	-2/-3 °C
Annual precipitation	900 mm	1963 mm	200 mm	400 mm
Avg. length of snow cover	7 months	9.5 months	9 months	9 months
Vegetation cover	Tundra	Barren	Tundra	Tundra

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Table 3: Details of driving data preparation for site simulations

		NUUK	SCHILTHORN	SAMOYLOV	BAYELVA
ATMOSPHERIC FORCING VARIABLES	**Air temperature**	*In-situ*	*In-situ*	*In-situ*	*In-situ*
	Precipitation	*In-situ*	*In-situ*	*In-situ* (snow season from WATCH)	*In-situ*
	Air pressure	*In-situ*	WATCH	WATCH	*In-situ*
	Atm. humidity	*In-situ*	*In-situ*	*In-situ*	*In-situ*
	Incoming longwave radiation	*In-situ*	*In-situ*	*In-situ*	WATCH
	Incoming shortwave radiation	*In-situ*	*In-situ*	WATCH	*In-situ*
	Net radiation	*In-situ*	-	*In-situ*	-
	Wind speed	*In-situ*	*In-situ*	*In-situ*	*In-situ*
	Wind direction	*In-situ*	-	*In-situ*	-
	Time period	26.06.2008-31.12.2011	01.10.1999-30.09.2008	14.07.2003-11.10.2005	01.01.1998-31.12.2009
STATIC SOIL PARAMETERS	**Soil porosity**	46%	50%	60%	41%
	Soil field capacity	36%	44%	31%	22%
	Mineral soil depth	36cm	710cm	800cm	30cm
	Dry soil heat capacity	2.213×10^6 ($Jm^{-3}K^{-1}$)	2.203×10^6 ($Jm^{-3}K^{-1}$)	2.1×10^6 ($Jm^{-3}K^{-1}$)	2.165×10^6 ($Jm^{-3}K^{-1}$)
	Dry soil heat conductivity	6.84 ($Wm^{-1}K^{-1}$)	7.06 ($Wm^{-1}K^{-1}$)	5.77 ($Wm^{-1}K^{-1}$)	7.93 ($Wm^{-1}K^{-1}$)
	Sat. hydraulic conductivity	2.42×10^{-6} (ms^{-1})	4.19×10^{-6} (ms^{-1})	2.84×10^{-6} (ms^{-1})	7.11×10^{-6} (ms^{-1})
	Saturated moisture potential	0.00519 (m)	0.2703 (m)	0.28 (m)	0.1318 (m)

Table 4: Details of model spin up procedures

|	JSBACH	ORCHIDEE	JULES	COUP	HYBRID8	LPJ-GUESS
Spin-up data	Observed climate	Observed climate	Observed climate	Observed climate	Observed climate	WATCH* data
Spin-up duration	50 years	10,000 years	50 years	10 years	50 years	500 years

*500 years forced with monthly WATCH reanalysis data from the 1901-1930 period, followed by daily WATCH forcing from 1901-until YYYY-MM-DD, then daily site-data.
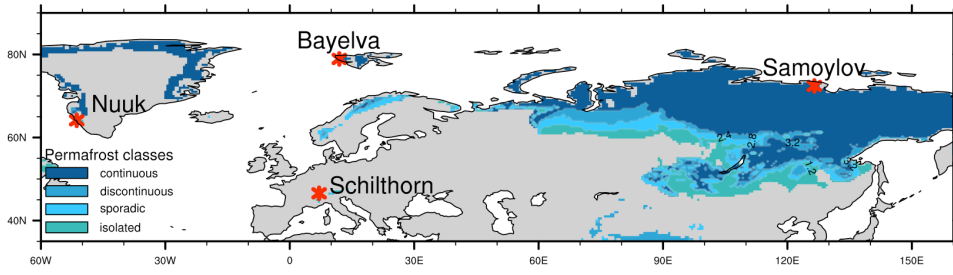
Figure 1: Location map of the sites used in this study. The background map is color coded with the IPA permafrost classes from Brown et al. (2002).
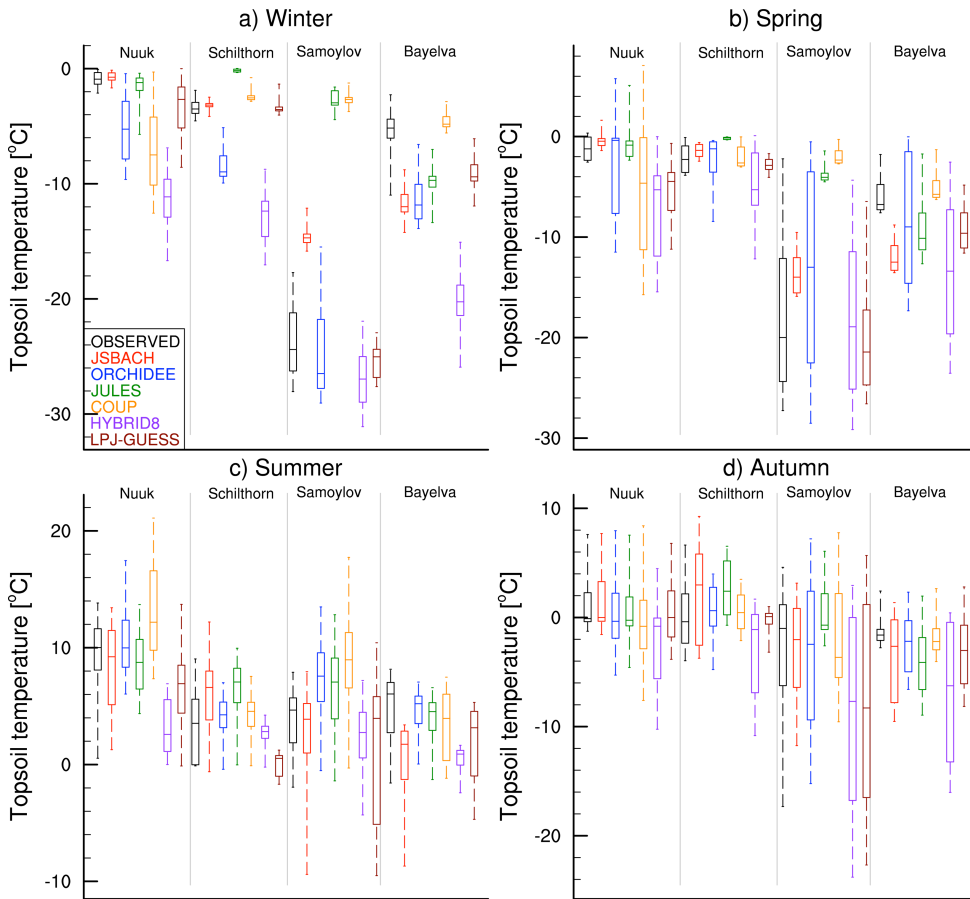
Figure 2: Box plots showing the topsoil temperature for observation and models for different seasons. Boxes are drawn with 25th percentile, mean and 75th percentiles while the whiskers show the min and max values. Seasonal averages of soil temperatures are used for calculating seasonal values. Each plot includes 4 study sites divided by the gray lines. Black boxes show observed values and colored boxes distinguish models. See Table A1 in Appendix-A for exact soil depths used in this plot.
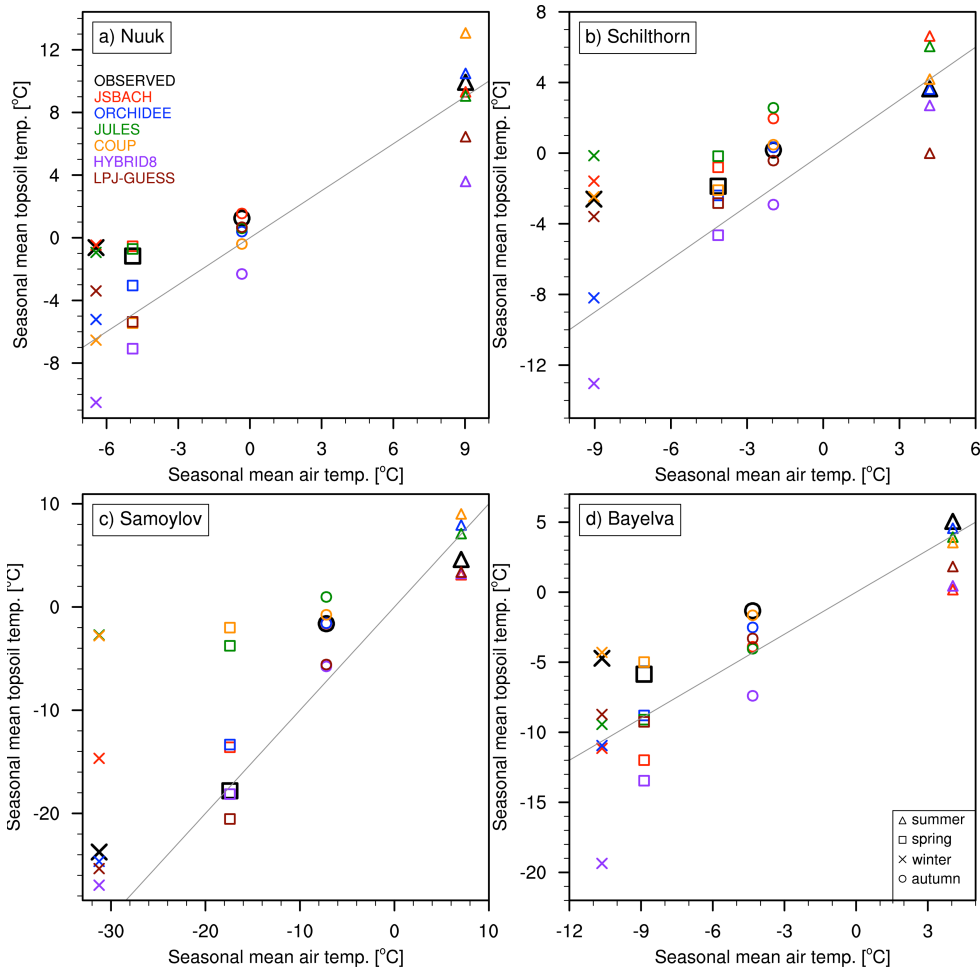
Figure 3: Scatter plots showing air/topsoil temperature relation from observations and models at each site for different seasons. Seasonal mean observed air temperature is plotted against the seasonal mean modeled topsoil temperature separately for each site. Black markers are observed values, colors distinguish models and markers distinguish seasons. Gray lines represent the 1:1 line. See Table A1 in Appendix-A for exact soil depths used in this plot.
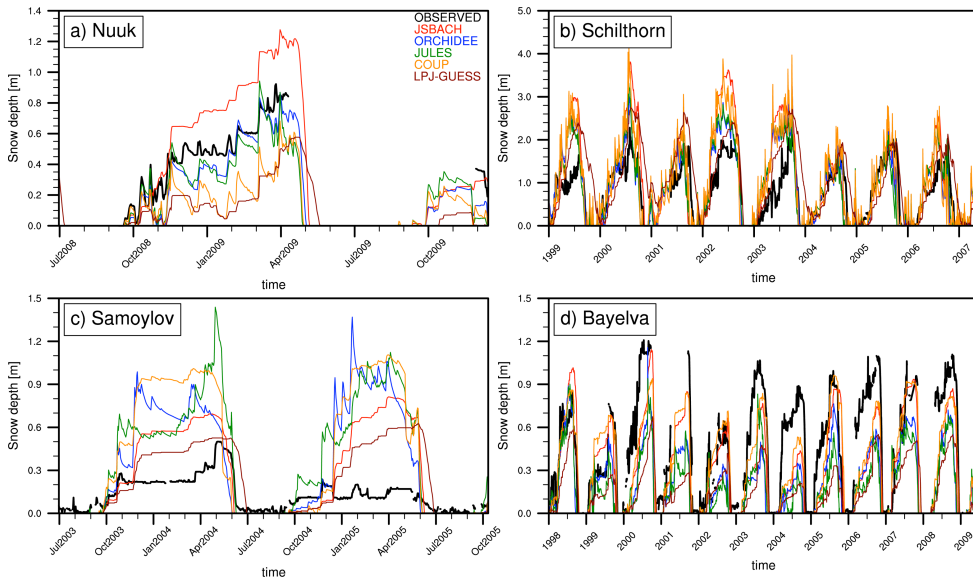
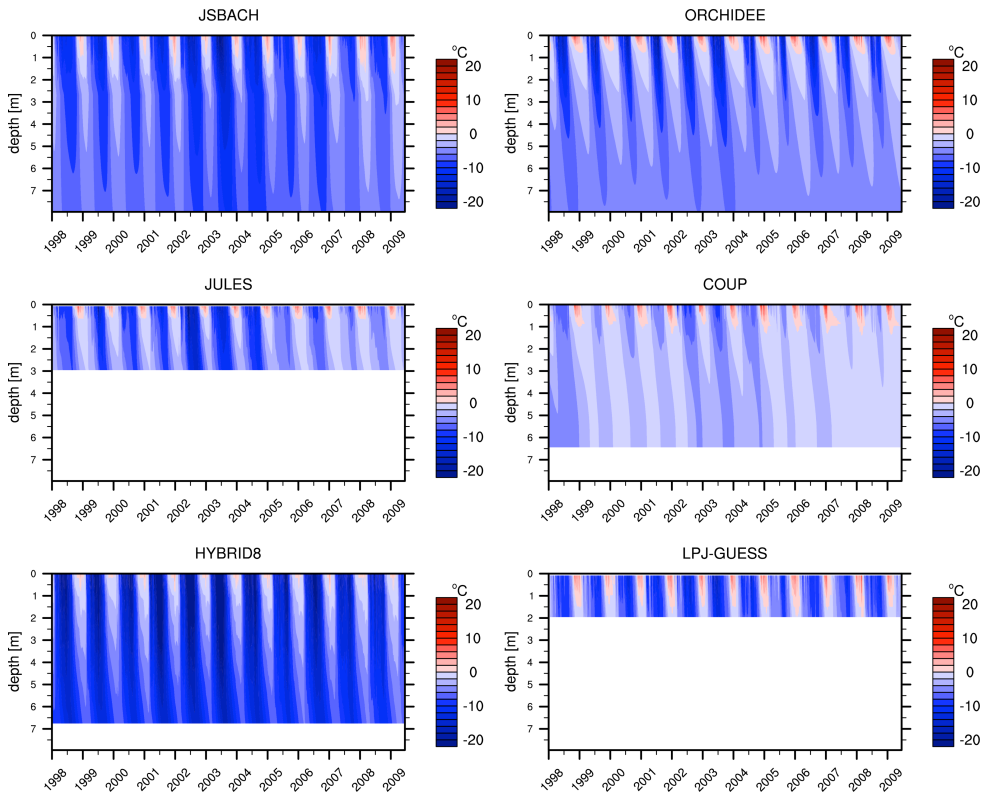Figure 4: Time series plots of observed and simulated snow depths for each site. Thick black lines are observed values and colored lines distinguish simulated snow depths from models.

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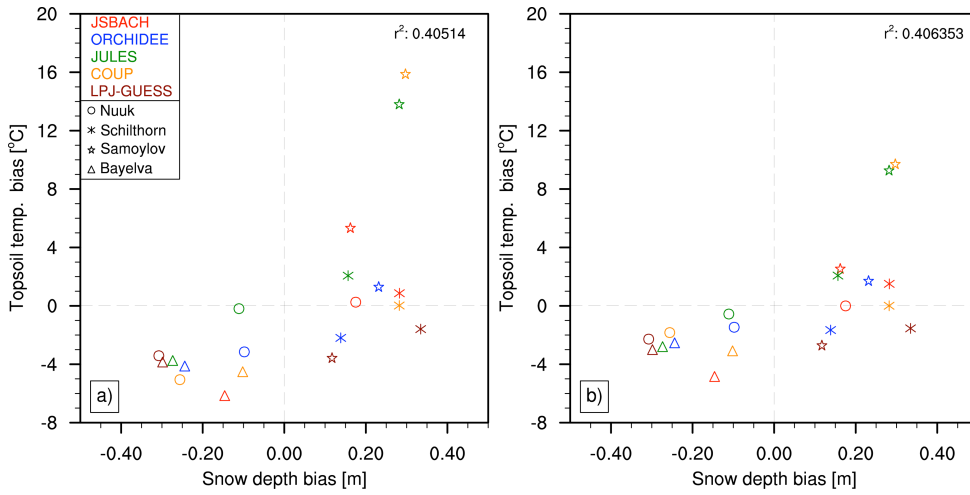
Figure 5: Scatter plots showing the relation between snow depth bias and topsoil temperature bias during snow season (a) and the whole year (b). Snow season is defined separately for each model, by taking snow depth values over 5 cm to represent the snow-covered period. The average temperature bias of all snow-covered days is used in (a), and the temperature bias in all days (snow covered and snow free seasons) is used in plot (b). Markers distinguish sites and colors distinguish models. See Table A1 in Appendix-A for exact soil depths used in this plot.
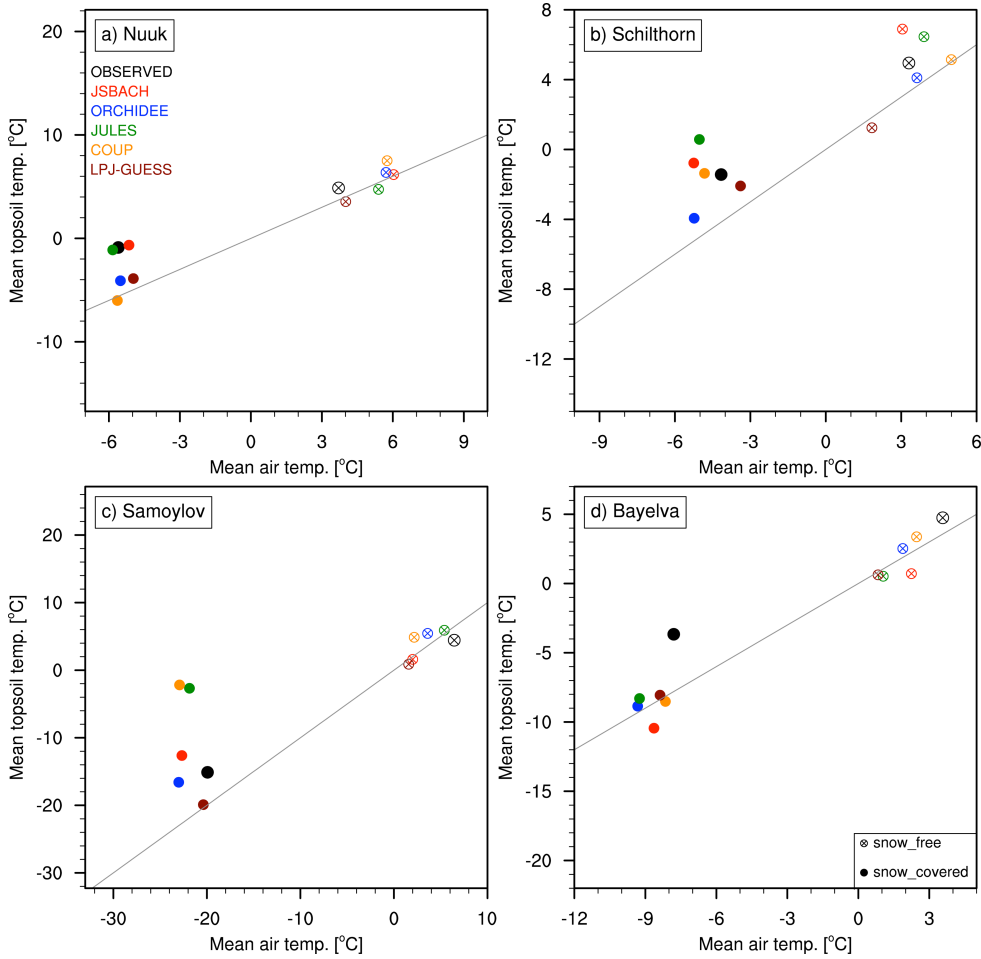
Figure 6: Scatter plots showing air/topsoil temperature relation from observations and models at each site for snow and snow-free seasons. Snow season is defined separately for observations and each model, by taking snow depth values over 5 cm to represent the snow-covered period. The average temperature of all snow covered (or snow free) days of the simulation period is used in the plots. Markers distinguish snow and snow free seasons and colors distinguish models. Gray lines represent the 1:1 line. See Table A1 in Appendix-A for exact soil depths used in this plot.
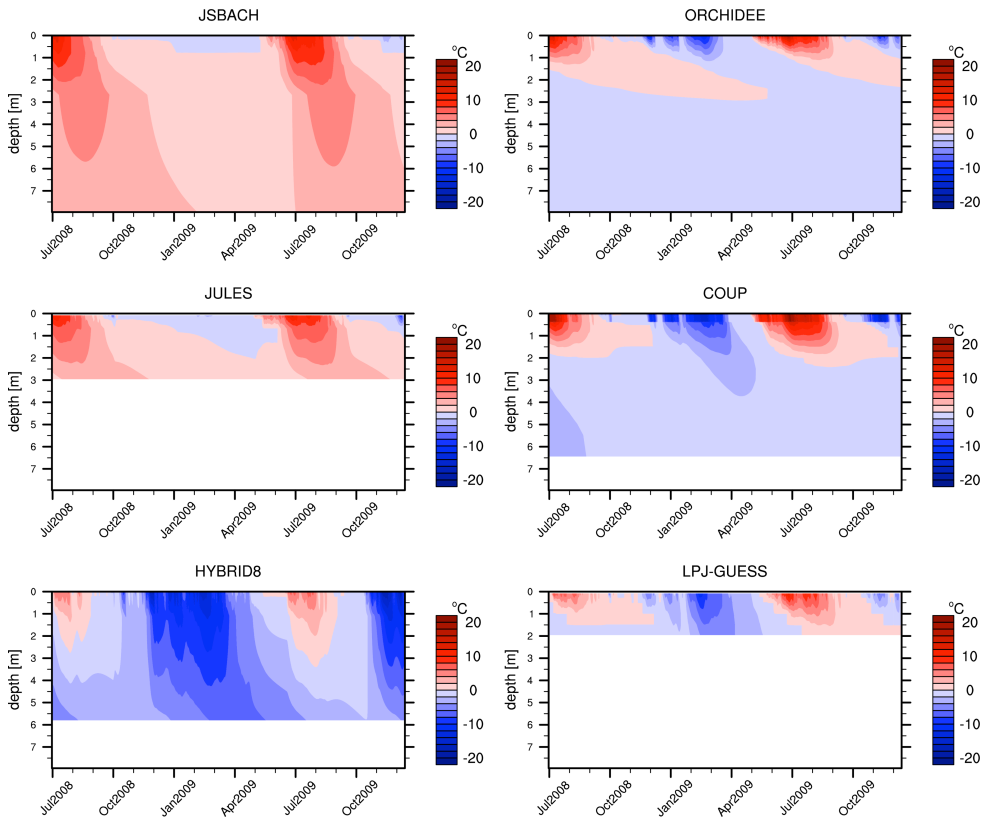
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3   Figure 7: Time-depth plot of soil temperature evolution at the Nuuk site for each model. Simulated soil temperatures are
4   interpolated into 200 evenly spaced nodes to represent a continuous vertical temperature profile. The deepest soil
5   temperature calculation is taken as the bottom limit for each model (no extrapolation applied).
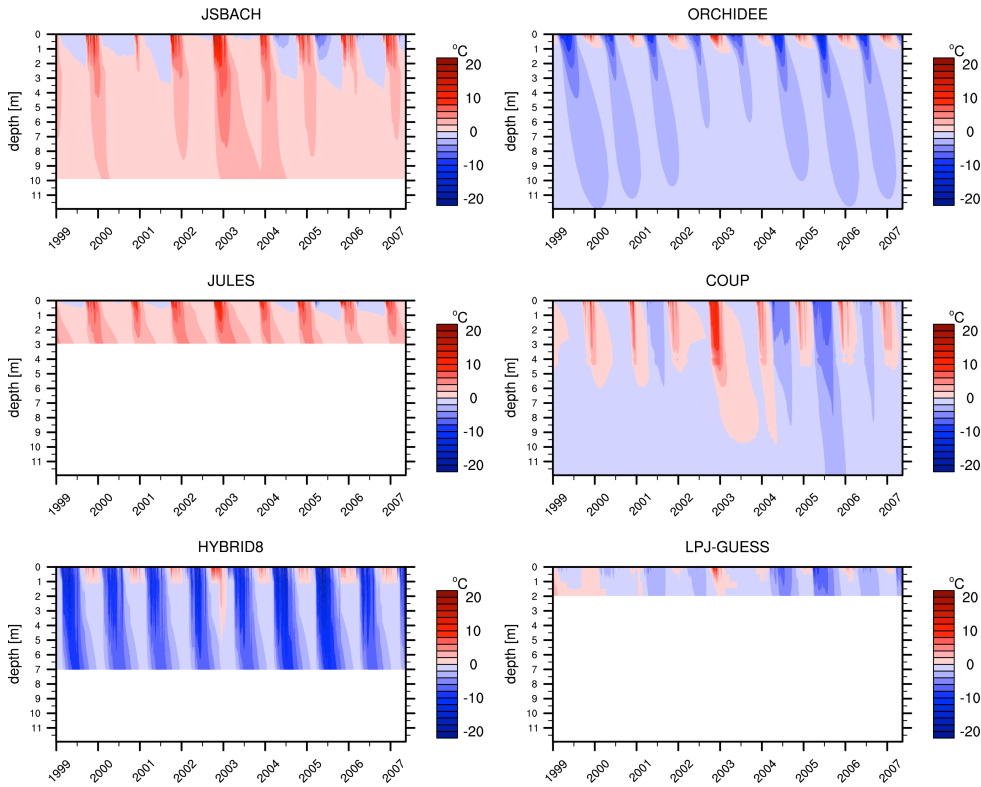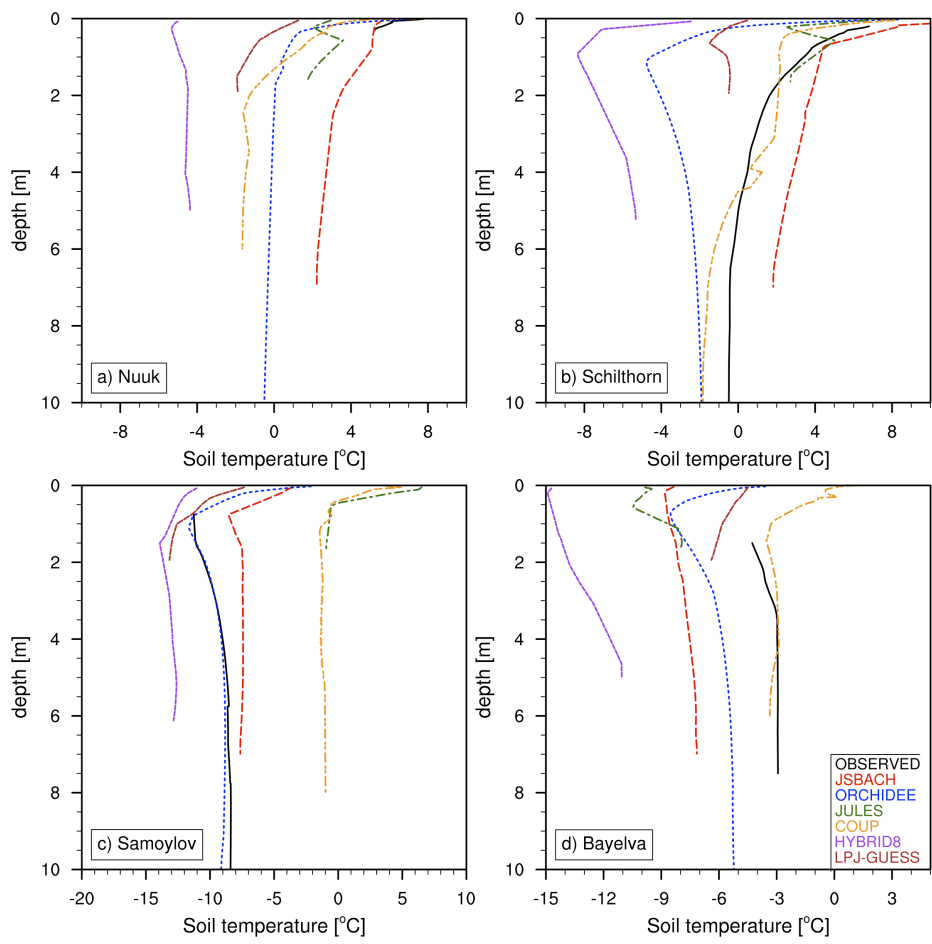
Figure 8: Time-depth plot of soil temperature evolution at Schilthorn site for each model. Simulated soil temperatures are interpolated into 200 evenly spaced nodes to represent a continuous vertical temperature profile. The deepest soil temperature calculation is taken as the bottom limit for each model (no extrapolation applied).
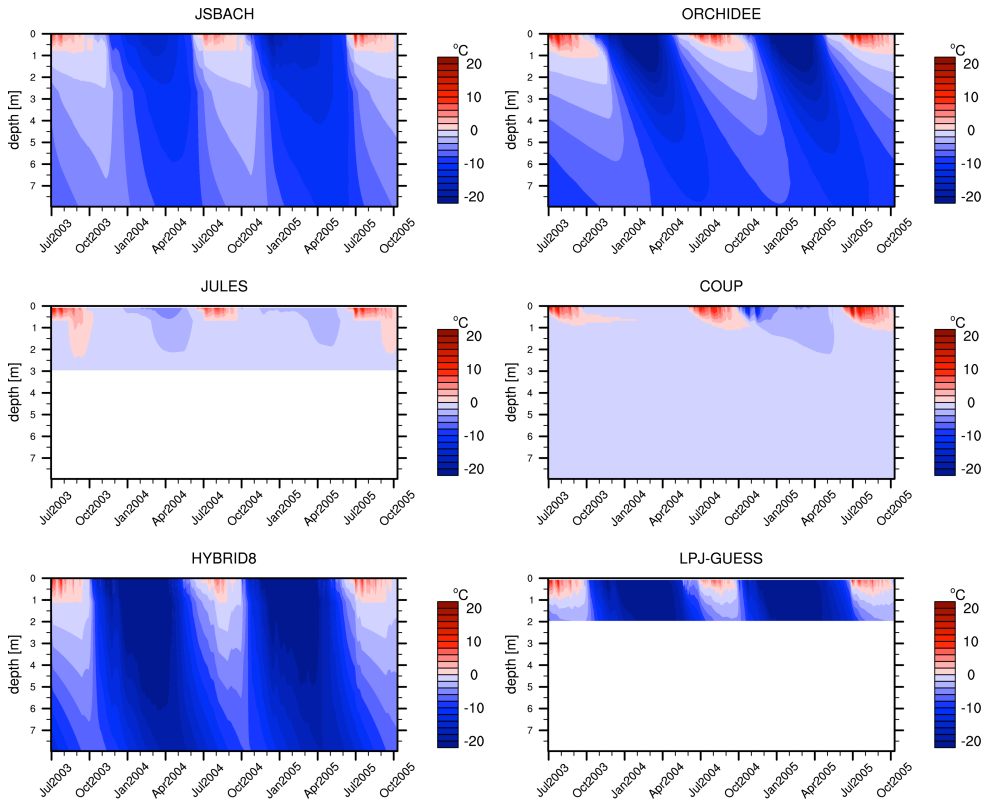
Figure 9: Time-depth plot of soil temperature evolution at Samoylov site for each model. Simulated soil temperatures are interpolated into 200 evenly spaced nodes to represent a continuous vertical temperature profile. The deepest soil temperature calculation is taken as the bottom limit for each model (no extrapolation applied).

Figure 10: Time-depth plot of soil temperature evolution at Bayelva site for each model. Simulated soil temperatures are interpolated into 200 evenly spaced nodes to represent a continuous vertical temperature profile. The deepest soil temperature calculation is taken as the bottom limit for each model (no extrapolation applied).

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2 Figure 11: Vertical profiles of annual soil temperature means of observed and modeled values at each site. Black thick

3 lines are the observed values while colored dashed lines distinguish models. (Samoylov and Bayelva observations are

4 from borehole data).
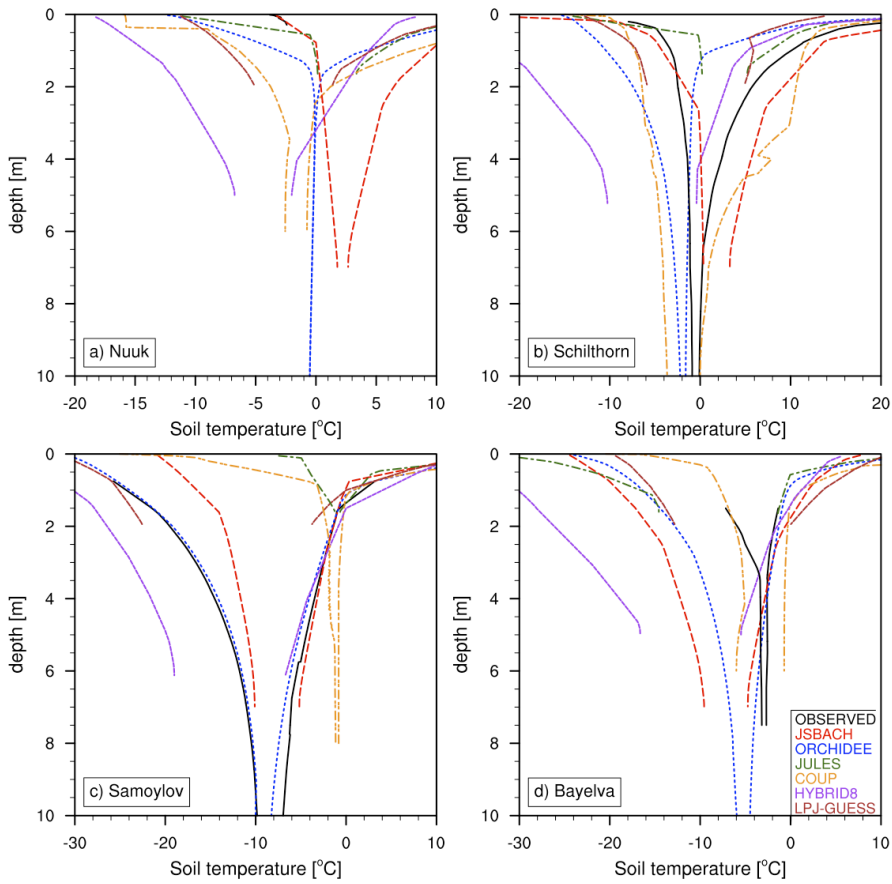
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2 Figure 12: Soil temperature envelopes showing the vertical profiles of soil temperature amplitudes of each model at
3 each site. Soil temperature values of observations (except Nuuk) and each model are interpolated to finer vertical
4 resolution and max and min values are calculated for each depth to construct max and min curves. For each color, the
5 right line is the maximum and the left line is the minimum temperature curve. Black thick lines are the observed values
6 while colored dashed lines distinguish models.

a) Schilthorn

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b) Samoylov

c) Bayelva
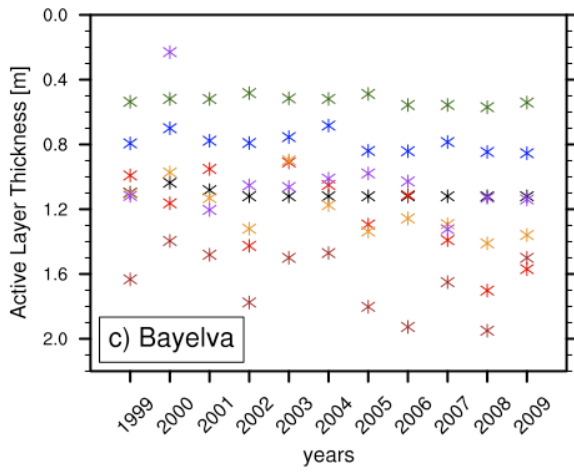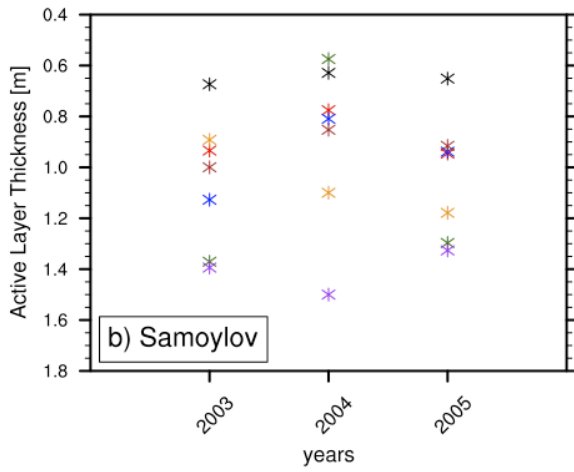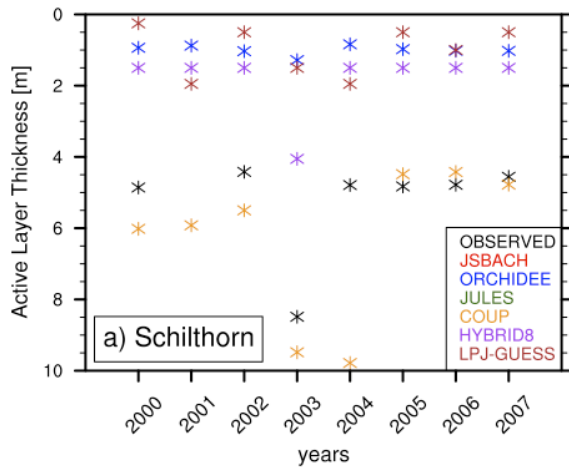
Figure 13: Active layer thickness (ALT) values for each model and observation at the three permafrost sites. ALT calculation is performed separately for models and observations by interpolating the soil temperature profile into finer resolution and estimating the maximum depth of 0°C for each year. Plots a, b and c show the temporal change of ALT at Schilthorn (2001 is omitted because observations have major gaps, also JSBACH and JULES are excluded as they simulate no permafrost at this site), Samoylov and Bayelva respectively. Colors distinguish models and observations

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comparisons are performed for

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Validation of topsoil temperature from observations and models gives an important estimate of the accuracy of several model processes such as atmosphere-soil coupling, surface insulation, subsoil thermal dynamics and hydrology.

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requires analyzing

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separately and inspecting key seasonal processes.

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distributions

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as well as the

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Winter and spring comparisons have larger biases than summer and autumn (Fig. 2). Observed

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temperatures are

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. Also there is a bigger range of temperature distribution at this site.

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, although the non-permafrost Nuuk site has warmer conditions than the others.

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As has been shown in a number of studies (e.g. Koven et al., 2013; Scherler et al., 2013; Gubler et al., 2013; Fiddes et al., 2013) modeled mean soil temperatures are strongly related to the atmosphere-surface thermal connection, which is strongly influenced by snow cover and its properties. Snow cover can increase the mean annual ground temperature and reduce the seasonal freezing depth (Zhang T., 2005).

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In this figure, soil

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vary for each model with

air temperature value, as

When snow insulation protects the topsoil from cooling, values lay above the 1:1 line. During the snow free season, when only vegetation cover, litter layers or organic layer insulation protects the topsoil from warming, values stay below the 1:1 line. However, insulation strength can change dramatically with critical levels of snow depth or dryness of the vegetation cover. Aside from the insulation effects, the relation between air and topsoil temperature depends also on atmosphere/topsoil temperature gradient, soil type/wetness and subsoil temperatures.

The site observations show warmer topsoil temperatures than air during autumn, winter and spring (Fig. 3). This situation indicates that soil is insulated when compared to colder air temperatures. This can be attributed to the snow cover during these seasons (Fig. 4). High insulating property of snow keeps the soil warmer than air, while not having snow usually results in colder topsoil temperatures than air (as for the HYBRID8 model, cf. Fig. 3). Even though the high albedo of snow provides a cooling effect for soil, the warming due to insulation dominates during most of the year. Depending on their snow depth bias, models show different relations between air and topsoil temperature. Figure 4 shows the changes in snow depth from observed and modeled values. Compared to observations, the models that underestimate the snow depth at Nuuk and Bayelva (Fig.

However, considering dynamic heat transfer parameters (volumetric heat capacity and heat conductivity) in snow representation seems to be of lesser importance (JSBACH vs. other models, see Table 1). This is related to the fact that most global models generally lack other important site-specific snow processes such as strong wind drifts (creating patchy snow cover), depth hoar formation and snow metamorphism (changing snow pack properties), snowmelt water infiltration into soil (additional heat transfer mechanism) and

snow albedo changes with these processes. As an example, the landscape heterogeneity at Samoylov forms different soil thermal profiles for polygon center and rim. While the soil temperature comparisons were performed for the polygon rim, snow depth observations were taken from polygon center. Due to strong wind drift almost all snow is removed from the rim and also limited to ca. 50cm (average polygon height) at the center (Boike et al., 2008). This way, models are forced to overestimate snow depth and insulation, in particular on the rim where soil temperature measurements have been taken. Hence, a resulting winter warm bias is no surprise (Fig. 2a, models JSBACH, JULES, COUP).

During summer, observed values show warmer topsoil temperatures than air at Nuuk and Bayelva, while the opposite is seen at Schilthorn and Samoylov (Fig. 3). Thicker moss cover and higher moisture content at Samoylov (Boike et al., 2008) is the reason for better insulation (hence cooler summer topsoil temperatures) at this site. Increasing thickness changes the heat storage of the moss cover and acts as a stronger insulator (Gornall et al., 2007). Additionally, water content of the moss layer affects the heat transfer parameters (Soudzilovskaia et al., 2013). However, without any plant cover, the cooler topsoil temperatures at Schilthorn point to non-vegetation induced insulation in summer. As snow can be persistent over spring season at high latitudes/altitudes and does not completely disappear in summer months (Fig. 4), it is worthwhile to separate snow and snow free seasons for these comparisons. Figure 6 shows the same atmosphere/topsoil temperature comparison as in Figure 3 but using snow and snow free seasons instead of conventional seasons. Evidently, without snow cover, the Schilthorn site indeed has warmer topsoil temperatures than air as expected (Fig. 6b).

Insulation strength is lower during the snow-free season but the model results are inconsistent with each other at all sites except Bayelva (Fig. 6). 10cm fixed moss cover in JSBACH and 10cm litter layer in LPJ-GUESS brings similar amounts of insulation. At Samoylov, where strong vegetation cover is observed in the field, these models perform better for snow free season (Fig.

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), hence hindering heat penetration from the surface.

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The simulated soil thermal regime at Samoylov reflects the colder climate at this site. All models show subzero temperatures below 1m (Fig. 9). However, compared to other models, JULES and COUP show values much closer to 0°C.

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latter. Together, these effects create warmer soil profiles.
At

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site, all models create permafrost conditions

The JULES and COUP models again show warmer temperature profiles than the other models. These models include soil heat transfer by advection that is lacked by other models.

In general, permafrost specific model experiments require deeper soil representation than 5-10 meters. As discussed in Alexeev et al. (2007), at least 30 m soil depth is needed for capturing decadal temperature variations in permafrost soils. The improvements from having such extended soil depth are shown in Lawrence et al. (2012) when compared

their older model version with shallow soil depth (Lawrence and Slater, 2005). Additionally, soil layer discretization plays an important role for the accuracy of

/water

within the soil, and hence can effect the ALT estimations. Most of the models in our intercomparison have less than 10 m depths, so they lack some effects of deep soil factors. However, most of the models used in global climate simulations have similar soil depth representations and the scope here is to compare models that are not directly aimed to simulate permafrost but to show general guidelines for future model developments. Apart from the permafrost conditions, the heat transfer rate also differs

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The soil thermal regime can also be investigated by studying the vertical temperature profiles regarding the annual means (Fig. 11), and minimum and maximum values (Fig. 12). In Fig. 11, the distribution of mean values is similar to the analysis of topsoil conditions. JSBACH and JULES and COUP overestimate the temperatures at Schilthorn and Samoylov, but almost all models underestimate it at Bayelva. Figure 12 shows the temperature envelopes of observed and simulated values at each site. The min. (max.) temperature curve represents the coldest (warmest) possible conditions for the soil thermal regime at a certain depth.

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Models

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due to its higher heat transfer rate.

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predicting the ALT.

These results clearly show the risks of estimating ALT by topsoil temperatures alone. Many previous studies (Lunardini, 1981; Kudryatsevet al., 1974; Romanovsky and Osterkamp, 1997; Shiklomanov and Nelson, 1999; Stendel et al., 2007, Anisimov et al., 1997) used simple relationships connecting topsoil temperatures and

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By doing so, they missed the effects of soil internal factors on ALT. A good review of widely used analytical approximations and differences to numerical approaches is given by Riseborough et al. (2008). More recently, Koven et al. (2013) and Slater and Lawrence (2013) have highlighted large model uncertainties in estimating permafrost extent and ALT values with similar approaches relating topsoil temperatures and permafrost conditions in ESMs. However, other processes like the heat transfer scheme in the subsurface layers and resulting water and ice contents are also important with regard to their impact on the soil temperature profile, and hence the ALT. The equilibrium models are not capable to estimate ALT in long-term simulations. So, using transient models and considering internal soil physical factors are critical to properly assess ALT within climate change context.

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Snow depth underestimation in models always leads to a cold bias in topsoil temperature, whereas snow depth overestimation does not always lead to a warm bias in topsoil temperatures.

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Active layer thickness is related to both surface conditions and the soil thermal regime. ALT estimation by topsoil temperatures can bring large errors.

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as well as soil internal physics of model formulations.

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