

**Impacts of snow and organic soils parameterization on North-Eurasian soil  
temperature profiles simulated by the ISBA land surface model.**

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## ABSTRACT

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2        In this study we analysed how an improved representation of snowpack processes and soil  
3 properties in the multi-layer snow and soil schemes of the ISBA land surface model impacts the  
4 simulation of soil temperature profiles over North-Eurasian regions. For this purpose, we refine  
5 ISBA's snow layering algorithm and propose a parameterization of snow albedo and snow  
6 compaction/densification adapted from the detailed Crocus snowpack model. We also include a  
7 dependency on soil organic carbon content for ISBA's hydraulic and thermal soil properties. First,  
8 changes in the snowpack parameterization are evaluated against snow depth, snow water  
9 equivalent, surface albedo, and soil temperature at a 10cm depth observed at the Col de Porte field  
10 site in the French Alps. Next, the new model version including all of the changes is used over  
11 Northern-Eurasia to evaluate the model's ability to simulate the snow depth, the soil temperature  
12 profile and the permafrost characteristics. The results confirm that an adequate simulation of snow  
13 layering and snow compaction/densification significantly impacts the snowpack characteristics and  
14 the soil temperature profile during winter, while the impact of the more accurate snow albedo  
15 computation is dominant during the spring. In summer, the accounting for the effect of soil organic  
16 carbon on hydraulic and thermal soil properties improves the simulation of the soil temperature  
17 profile. Finally, the results confirm that this last process strongly influences the simulation of the  
18 permafrost active layer thickness and its spatial distribution.

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## 19 **1. Introduction**

20 Snowpack properties are known to be of primary importance for understanding the water  
21 and energy budgets of the land surface, especially in mountainous and boreal regions. From  
22 autumn to spring, solid precipitation is stored within the snowpack thereby modifying the  
23 terrestrial albedo and roughness length, and impacting the radiative and energy fluxes at the  
24 soil/atmosphere interface. During spring, the fresh water released by snowmelt contributes to soil  
25 infiltration, intense streamflow and large seasonal flood events, and it directly modulates the land  
26 surface evapotranspiration [Poutou et al. 2004; Niu and Yang 2006; Decharme and Douville  
27 2007]. Snowpack also acts as an insulating layer at the surface which prevents significant heat loss  
28 in the winter. Over North-Eurasian regions, as discussed by Paquin and Sushama [2015], this last  
29 process controls the temperature of the permafrost. It is defined as a soil that remains below 0°C  
30 for two or more consecutive years, and it has a significant influence on the summer permafrost  
31 active layer thickness, defined as the maximum annual thaw depth. In summary, snowpack  
32 properties drastically influence soil/atmosphere interactions during a large part of the year through  
33 their impacts on many land surface processes.

34 Beside the importance of snowpack properties for understanding the water and energy  
35 budgets of the land surface in northern regions, the physical properties of soil organic carbon (or  
36 peat soil) also play a significant role. North-Eurasian soils are very rich in organic carbon because  
37 the low soil temperatures in this region inhibit decomposition of dead plant material that  
38 accumulates over time, thereby forming peat deposits. Soil organic carbon exhibits very different  
39 hydraulic and thermal properties than mineral soil [Boelter 1969; Letts et al. 2000]. It is  
40 characterized by a very high porosity, a weak hydraulic suction, and a sharp vertical hydraulic  
41 conductivity profile from high values at the surface to very low values at the subsurface. This  
42 generally induces a relatively wet soil with a shallow water table [Letts et al. 2000]. Its low  
43 thermal conductivity and its relatively high heat capacity act as an insulator for soil temperature  
44 that prevents the soil from significant warming during the summer [Bonan and Shugart 1989;

45 Lawrence and Slater 2008]. Over permafrost regions, the hydraulic and thermal properties of soil  
46 organic carbon partly control the soil depth reached by the 0°C isotherm which, in turn, defines the  
47 thickness of the active layer during summer [Paquin and Sushama 2015]. Through its influence on  
48 soil temperature and wetness, it impacts the continental part of the carbon cycle and the land  
49 surface CO<sub>2</sub> and CH<sub>4</sub> emissions to the atmosphere [Walter et al. 2006; Zimov et al. 2006].

50 In atmospheric, climate, and hydrological models, the dynamics of the snowpack and the  
51 evolution of water and heat profiles within the soil are simulated using so-called Land Surface  
52 Models (LSM). These LSMs, like the simple bucket scheme of Manabe [1969], were initially  
53 developed over four decades ago in order to simulate realistic land surface water and energy  
54 budgets in atmospheric general circulation models. Now, LSMs are used in many applications  
55 such as hydrological and meteorological forecasts, global hydrological and biogeochemical  
56 studies, and climate evolution prediction. Many LSMs use multi-layer soil schemes in which the  
57 vertical transport of moisture and heat into the soil is explicitly solved for using diffusion  
58 equations [e.g. Decharme et al. 2011]. Because the total soil depth is discretized using multiple  
59 layers, these schemes allow the representation of the vertical root distribution [Zeng et al., 1998;  
60 Feddes et al., 2001; Braud et al., 2005], as well as the surface/groundwater capillary exchanges  
61 [e.g. Vergnes et al. 2014]. Finally, their coupling with a multi-layer snowpack scheme permits a  
62 representation of the interaction between cold physical processes, such as the effect of snow on  
63 soil temperature, hydrology, and freezing [Slater et al. 2001; Luo et al. 2003; Gouttevin et al.  
64 2012].

65 Three major classes of snowpack schemes exist in LSMs: single-layer schemes, multi-layer  
66 schemes of intermediate complexity, and detailed snowpack models. The first class was used  
67 preferentially in the past within forecast and climate models. The snowpack was represented with  
68 only one layer that evolves seasonally, which is characterized as having a high albedo, a low  
69 thermal conductivity, and a low thermal capacity [Manabe 1969; Verseghy 1991; Douville et al.  
70 1995]. More recently, these simple single-layer schemes have been replaced by intermediate

71 complexity models inspired by the pioneering work of Anderson [1979]. These schemes use a  
72 multi-layer approach with the minimum number of layers needed to simulate all of the  
73 macroscopic physical properties of the snowpack such as albedo, compaction, density, and water  
74 refreezing [Lynch-Stieglitz 1994; Loth and Graf 1998; Boone and Etchevers 2001; Brown et al.  
75 2006; Oleson et al. 2010; Dutra et al. 2010; Shrestha et al. 2010; Best et al. 2011; Kuipers  
76 Munneke et al. 2011]. Finally, more complex snowpack models have been developed primarily in  
77 support of avalanche forecasting, and more generally for all applications (including process  
78 studies) requiring a detailed representation of the vertical profile of the physical properties of  
79 snow. In addition to simulating macroscopic snowpack physical properties, they explicitly account  
80 for the time evolution of the snow microstructure driven by snow metamorphism, and the multiple  
81 feedback loops involving internal snow processes and the energy and mass balance at the air/snow  
82 and snow/ground interface [Brun et al. 1989, 1992; Jordan 1991; Bartelt and Lehning 2002]. In  
83 addition, these models can serve as a reference for the development and evaluation of intermediate  
84 complexity snowpack schemes.

85         The Interaction-Soil-Biosphere-Atmosphere (ISBA) LSM developed at Météo-France  
86 currently uses a multi-layer approach for the snowpack [Boone and Etchevers 2001] and the soil  
87 [Boone et al., 2000; Decharme et al. 2011]. ISBA is the land surface model embedded in the  
88 SURFEX (SURFace EXternalized) modeling platform [Masson et al. 2013], which is used in all of  
89 the atmospheric meso-scale, regional-scale and global-scale models of Météo-France, as well as in  
90 regional hydrological forecasting systems, global hydrological models and model chains in support  
91 of avalanche hazard warning [e.g. Lafaysse et al., 2013; Vernay et al., in press]. The ISBA multi-  
92 layer version was evaluated over many local or regional field datasets [Boone et al., 2000;  
93 Decharme et al. 2011, 2013; Canal et al. 2014; Parrens et al. 2014; Vergnes et al. 2014; Joetzjer et  
94 al. 2015], increasing our confidence in the model's capability to simulate realistic land surface  
95 processes under a variety of climate conditions. However, over cold regions, winter top soil  
96 temperatures tend to be underestimated [Wang et al. 2016] while during summer they are generally

97 too warm. The first biases are attributable to the ISBA multi-layer snowpack scheme of  
98 intermediate complexity developed by Boone et al. [2000] and based on Anderson [1979]. Indeed,  
99 when the ISBA multi-layer soil scheme is coupled with the detailed Crocus snowpack model, the  
100 winter soil temperature simulated at 20cm depth better matches observations over the Northern  
101 Eurasian regions [Brun et al. 2013]. Secondly, ISBA only accounts for mineral soil properties  
102 while many studies pointed out that the specific properties of soil organic carbon are required to  
103 simulate realistic soil thermal regime over cold regions [Nicolosky et al. 2007; Beringer et al. 2001;  
104 Lawrence and Slater 2008; Lawrence et al. 2008; Dankers et al. 2011].

105         The present study focuses on the impact of improving the representation of snowpack and  
106 soil properties in the ISBA LSM to reproduce snow characteristics and soil temperature profiles  
107 over cold regions. We replaced the original Boone and Etchevers [2001] representation of snow  
108 layering, albedo and snow compaction by adapting some parameterizations used in the Crocus  
109 snowpack model [e.g. Vionnet et al. 2012]. In addition, we added a parameterization of the organic  
110 carbon effect on hydraulic and thermal soil properties based on the pedotransfer function of  
111 Boelter [1969] and inspired by works of Letts et al. [2000] and Lawrence and Slater [2008]. The  
112 changes in the snowpack parameterizations are first evaluated at the Col de Porte field site located  
113 in the French Alps [Morin et al. 2012]. This dataset includes many observations at a daily time  
114 step such as snow depth, snow water equivalent, surface albedo and soil temperature at 10 cm  
115 from 1993 to 2011. In addition the meteorological observations required to drive the model are  
116 given at a 3-hourly time step over the same period. The new parameterizations were evaluated next  
117 over the North-Eurasian region using the same experimental design as Brun et al. [2013] using in-  
118 situ evaluation datasets of snow depth and soil temperature profile measurements and  
119 meteorological driving data from a global reanalysis. To quantify the model's ability to simulate  
120 the permafrost characteristics, two additional datasets were used that estimate the location of  
121 permafrost boundaries and the active layer thickness over the Yakutia region. A brief review of the  
122 ISBA multi-layer model is given in section 2, all of the snowpack and soil parameterization

123 improvements and updates are presented in section 3, sections 4 and 5 describe the model  
124 evaluation over the Col de Porte field site and the North-Eurasian region, respectively. Finally, a  
125 discussion and the main conclusions are given in section 6.

## 126 **2. Review of the ISBA land surface model**

### 127 *2.1. Soil processes*

128 The ISBA multi-layer model solves the one-dimensional Fourier law and the mixed-form  
129 of the Richards equation explicitly to calculate the time evolution of the soil energy and water  
130 budgets [Boone et al., 2000; Decharme et al. 2011]. In each layer  $i$ , the closed-form equations  
131 between the soil liquid water content,  $w$  ( $\text{m}^3 \cdot \text{m}^{-3}$ ), and the soil hydrodynamic parameters, such as  
132 the soil matric potential,  $\psi$  (m), and the hydraulic conductivity,  $k$  ( $\text{m} \cdot \text{s}^{-1}$ ), are determined according  
133 to the Brooks and Corey [1966] model adapted by Campbell [1974] as follows:

$$134 \quad \psi(i) = \psi_{sat}(i) \left( \frac{w(i)}{w_{sat}(i)} \right)^{-b(i)} \quad \text{and} \quad k(i) = k_{sat}(i) \left( \frac{\psi(i)}{\psi_{sat}(i)} \right)^{\frac{2b(i)+3}{b(i)}} \quad (1)$$

135 where,  $b$  represents the dimensionless shape parameter of the soil-water retention curve,  $w_{sat}$   
136 ( $\text{m}^3 \cdot \text{m}^{-3}$ ) the soil porosity, and  $\psi_{sat}$  (m) and  $k_{sat}$  ( $\text{m} \cdot \text{s}^{-1}$ ) the soil matric potential and hydraulic  
137 conductivity at saturation, respectively. In this study, the heat and soil moisture transfers within the  
138 soil are computed using 14 layers up to a 12 m depth. The depth of the 14 layers (0.01m, 0.04m,  
139 0.1m, 0.2m, 0.4m, 0.6m, 0.8m, 1.0m, 1.5m, 2.0m, 3.0m, 5.0m, 8.0m, 12.0m) have been chosen to  
140 minimize numerical errors in solving the finite-differenced diffusive equations, especially in the  
141 uppermost meter of the soil [Decharme et al. 2011]. Saturated hydraulic conductivity, matric  
142 potential at saturation, and porosity of the mineral soil are related to the soil texture [Noilhan and  
143 Lacarrère 1995]. The total heat capacity of the mineral soil in each layer is computed as the sum of  
144 the soil matrix, water and ice heat capacities weighted by the volumetric water and ice content  
145 [Peters-Lidard et al. 1998]. The thermal conductivity of the mineral soil is computed via a more

146 complex combination of water, ice and soil conductivities as proposed by Peters-Lidard et al.  
147 [1998].

148 The soil ice content tendency (partial time derivative) is solved explicitly in each layer of  
149 the soil and accounts for ice sublimation and vegetation insulation effect at the surface [e.g. Boone  
150 et al., 2000]. The liquid water content that can freeze is limited by a maximum value,  $w_{lmax}$  ( $m^3 \cdot m^{-3}$ ),  
151 computed as a function of temperature based on the Gibbs free-energy method [Fuchs et al.  
152 1978]:

$$153 \quad w_{lmax}(i) = w_{sat}(i) \times \min \left[ 1.0, \left( \frac{L_f}{g \psi_{sat}(i)} \frac{T_g(i) - T_f}{T_g(i)} \right)^{-1/b(i)} \right] \quad (2)$$

154 where  $w_{sat}$  ( $m^3 \cdot m^{-3}$ ) is the soil porosity in each layer  $i$ ,  $T_g$  (K) the soil temperature,  $g$  ( $m \cdot s^{-2}$ ) the  
155 terrestrial gravity constant,  $T_f$  (273.16 K) is the triple-point temperature for water, and  $L_f$  ( $3.337$   
156  $\times 10^5$  J.kg $^{-1}$ ) the latent heat of fusion. The total water content in each soil layer is conserved during  
157 phase changes. When the soil freezes, the liquid water content will decrease owing to a  
158 corresponding increase in soil ice content. Finally, the maximum temperature,  $T_{max}$  (K), used for  
159 phase changes can be determined via the same Gibbs free-energy method :

$$160 \quad T_{max}(i) = \frac{L_f T_f}{L_f - g \psi(i)} \quad (3)$$

161 where the soil matric potential  $\psi$  is defined using Equation 1. Thus, this scheme induces  
162 dependencies of water phase changes to soil textures and to the degree of soil humidity. The  
163 coarser the soil texture, the larger the quantity of water that will freeze at a given temperature. As  
164 the soil becomes dry, the temperature that allows freezing drops. More details can be found in the  
165 supplementary material of Masson et al. [2013] ([http://www.geosci-model-](http://www.geosci-model-dev.net/6/929/2013/gmd-6-929-2013-supplement.pdf)  
166 [dev.net/6/929/2013/gmd-6-929-2013-supplement.pdf](http://www.geosci-model-dev.net/6/929/2013/gmd-6-929-2013-supplement.pdf))

## 167 2.2. *Snowpack internal processes*

168 The original ISBA explicit multi-layer snow scheme developed by Boone and Etchevers  
169 [2001] is a snowpack scheme of intermediate complexity made in order to take into account for



170 some processes such as snow mass and heat vertical redistribution, snow compaction, water  
 171 percolation and refreezing, and explicit heat conduction at the snow/soil interface. Many of these  
 172 processes, such as snow compaction or absorption of solar energy, are based on works of  
 173 Anderson [1976] and Loth et al. [1993]. The thermal conductivity of snow (Appendix A) is  
 174 computed via the snow density [Yen 1981]. An additional term depends on the snow temperature  
 175 to account for vapor transfer through the snowpack [Sun et al. 1999]. The time evolution of the  
 176 snow mass is linked to snowmelt, water freezing, evaporation, and liquid flow. The liquid water  
 177 content into the snowpack is simulated as a succession of bucket-type reservoirs. A maximum  
 178 liquid water holding capacity ( $W_{lmax}$ ) is computed in each layer. It varies from 3% to 10% of the  
 179 snow mass according to a decrease in snow density after Anderson [1976]. A liquid water flux is  
 180 generated when the liquid water content exceeds this threshold. More details can be found in  
 181 Boone and Etchevers [2001] and only internal physical processes of the snowpack discussed in  
 182 this study are described below.

### 183 2.2.1. Snow layering

184 In the original ISBA explicit snow scheme, three-layers are used for snow layering because  
 185 it is considered to be the minimum number required to resolve adequately the snow thermal profile  
 186 within the snowpack [Lynch-Stieglitz 1994; Loth and Graf, 1998; Boone and Etchevers 2001]. The  
 187 algorithm that computes the snow grid thicknesses  $\Delta z$  of each layer,  $i$ , is described as follows:

$$\begin{cases} \Delta z(1) = \delta 0.25 h_{sn} + (1 - \delta) 0.05 \\ \Delta z(2) = \delta 0.5 h_{sn} + (1 - \delta) \times \min[0.5, 0.05 + 0.34(h_{sn} - \Delta z(1))] \\ \Delta z(3) = \delta 0.25 h_{sn} + (1 - \delta)(h_{sn} - \Delta z(1) - \Delta z(2)) \end{cases} \quad \text{with} \quad \begin{cases} \delta = 1 & \forall (h_{sn} \leq 0.2) \\ \delta = 0 & \forall (h_{sn} > 0.2) \end{cases} \quad (4)$$

189 where  $h_{sn}$  (m) is the total snow depth. As long as the snow remains below 0.2m, the fraction of the  
 190 total depth that defines the thickness of each layer remains with a fine resolution at the top and the  
 191 base of the snowpack. When the snow depth exceeds 0.2m, the thickness of the first layer remains  
 192 equal to 0.05m, in order to adequately solve the diurnal cycle of the surface energy balance. In  
 193 addition, for large snow depth values, the second layer thickness cannot exceed 0.5m because  
 194 density and heat vertical gradients are generally the largest near the top of the snowpack. The

195 vertical grid is updated at the beginning of each time step before the computation of the other  
 196 snowpack internal processes.

### 197 2.2.2. Snow compaction

198 The evolution of snow density,  $\rho_{sn}$  ( $\text{kg.m}^{-3}$ ) in each layer,  $i$ , is the sum of snow compaction  
 199 due to change in snow viscosity,  $\eta$  (Pa s), and settling due to freshly fallen snow,  $\xi$  ( $\text{s}^{-1}$ ), following  
 200 Anderson [1976] and Loth et al. [1993]:

$$201 \frac{1}{\rho_{sn}(i)} \frac{\partial \rho_{sn}(i)}{\partial t} = \frac{\sigma(i)}{\eta(i)} + \xi(i) \quad \text{with} \quad \sigma(i) = g \sum_{j=1}^i [\Delta z(j) \rho_{sn}(j)] \quad (5)$$

202 where  $\sigma$  (Pa) is the snow vertical stress. The snow viscosity and settling of new snow are solved  
 203 using two empirical exponential functions of snow density and temperature,  $T_{sn}$  (K), :

$$204 \begin{cases} \eta(i) = v_0 \exp(v_1(T_f - T_{sn}(i)) + v_2 \rho_{sn}(i)) \\ \xi(i) = s_0 \exp(-s_1(T_f - T_{sn}(i)) - s_2 \times \max(0, \rho_{sn}(i) - \rho_d)) \end{cases} \quad (6)$$

205 where  $v_0 = 3.7 \cdot 10^7$  Pa s,  $v_1 = 0.081 \text{ K}^{-1}$ ,  $v_2 = 0.018 \text{ m}^3 \cdot \text{kg}^{-1}$ ,  $s_0 = 2.8 \cdot 10^{-6} \text{ s}^{-1}$ ,  $s_1 = 0.04 \text{ K}^{-1}$ ,  $s_2 = 0.046$   
 206  $\text{m}^3 \cdot \text{kg}^{-1}$ , and  $\rho_d = 150 \text{ kg.m}^{-3}$  are empirical parameters calibrated by Anderson [1976]. The  
 207 minimum density of snow is constrained to be  $50 \text{ kg.m}^{-3}$ . The snowfall density,  $\rho_{snow}$  ( $\text{kg.m}^{-3}$ ), is  
 208 expressed as a function of wind speed,  $V_a$  ( $\text{m.s}^{-1}$ ), and air temperature,  $T_a$  (K), following an  
 209 experimental study of Pahaut [1976] :

$$210 \rho_{snow} = a_\rho + b_\rho (T_a - T_f) + c_\rho V_a^{1/2} \quad (7)$$

211 where the coefficients  $a_\rho = 109 \text{ kg.m}^{-3}$ ,  $b_\rho = 6 \text{ kg.m}^{-3} \cdot \text{K}^{-1}$ , and  $c_\rho = 26 \text{ kg.s}^{1/2} \cdot \text{m}^{-7/2}$ .

### 212 2.2.3. Transmission of solar radiation and Snow albedo

213 The absorption of incident shortwave solar radiation,  $R_{sw}$  ( $\text{W.m}^{-2}$ ), within the snowpack is  
 214 solved over a single spectral band. It uses an exponential decrease of incoming radiation with  
 215 snow depth [Anderson 1976; Loth et al. 1993]. So, the net shortwave radiation  $Q_{sn}$  ( $\text{W.m}^{-2}$ )  
 216 absorbed by the snow level,  $i$ , is given by:

$$217 Q_{sn}(i) = (1 - \alpha_{sn}) R_{sw} \exp\left(-\sum_{j=1}^i [\beta_{sn}(j) \Delta z(j)]\right) \quad (8)$$

218 where  $\alpha_{sn}$  is the dimensionless snow albedo, and  $\beta_{sn}$  ( $\text{m}^{-1}$ ) the extinction coefficient of snow which  
219 is given by :

$$220 \quad \beta_{sn}(i) = C_v \rho_{sn}(i) / \sqrt{d_{opt}(i)} \quad (9)$$

221 As shown by Bohren and Barkstrom [1974], this extinction of snow is directly related to its  
222 density, the optical diameter  $d_{opt}$  (m), and a constant  $C_v = 3.8 \cdot 10^{-3} \text{m}^{5/2} \cdot \text{kg}^{-1}$ . The optical diameter is  
223 empirically linked to the snow density following a simple polynomial regression established by  
224 Anderson [1976]:

$$225 \quad d_{opt}(i) = \min(d_{max}, g_1 + g_2 \times \rho_{sn}(i)^4) \quad (10)$$

226 where  $d_{max}$  (m) is the maximum value equal to  $2.796 \cdot 10^{-3} \text{m}$ , and the coefficients  $g_1 = 1.6 \cdot 10^{-4}$ ,  
227 and  $g_2 = 1.1 \cdot 10^{-13} \text{m}^{13} \cdot \text{kg}^{-4}$  were calibrated by Anderson [1976]. The time evolution of snow albedo  
228 is modelled in a simple way using time constants after Douville et al. [1995]. A linear decrease  
229 rate is used for dry snow and an exponential decrease is used for wet snow while the snow albedo  
230 increases linearly with snowfall intensity [Boone and Etchevers 2001]. The snow albedo is  
231 constrained to be between its minimum value,  $\alpha_{min} = 0.5$ , and its maximum,  $\alpha_{max} = 0.85$ .

### 232 **3. Changes in explicit snow and soil schemes**

#### 233 *3.1. Changes in snowpack internal processes*

##### 234 *3.1.1. Snow layering*

235 Detailed snowpack models use more than a dozen layers to simulate well the snow thermal  
236 profile and the snowpack stratigraphy [Armstrong and Brun 2008; Vionnet et al 2012]. This  
237 configuration allows a good computation of the diurnal cycle through the use of fine top layers,  
238 while bottom layers are also sufficiently thin to ensure a good computation of the heat conduction  
239 at the snow/soil interface. However, these models were rarely used in global atmospheric, climate,  
240 and/or hydrological models due to their high computational costs partly due to the use of a large  
241 number of layers. For this reason, the multi-layer snow scheme in ISBA was developed using only  
242 three layers representing a good compromise between a reasonable simulation of the snow thermal

243 profile [Boone and Etchevers 2001] and a low computing time. Today, such computational  
 244 limitations are less of a constraint and a larger number of layers can be used in this scheme. The  
 245 number of snow layers in ISBA was increased to 12 with two fine layers at the top and the bottom  
 246 of the snowpack using the following simple algorithm:

$$\begin{cases}
 \Delta z(i) = \min\left(\delta_i, \frac{h_{sn}}{12}\right) & \forall i \leq 5 \quad \text{or} \quad \forall i \geq 9 \\
 \Delta z(6) = 0.3d_r - \min[0, 0.3d_r - \Delta z(5)] \\
 \Delta z(7) = 0.4d_r + \min[0, 0.3d_r - \Delta z(5)] + \min[0, 0.3d_r - \Delta z(9)] \\
 \Delta z(8) = 0.3d_r - \min[0, 0.3d_r - \Delta z(9)] \\
 d_r = h_{sn} - \sum_{i=1}^5 \Delta z(i) - \sum_{i=9}^{12} \Delta z(i)
 \end{cases} \quad (11)$$

248 where the constants are defined as:  $\delta_1 = 0.01\text{m}$ ,  $\delta_2 = 0.05\text{m}$ ,  $\delta_3 = 0.15\text{m}$ ,  $\delta_4 = 0.5\text{m}$ ,  $\delta_5 = 1\text{m}$ ,  $\delta_9 =$   
 249  $1\text{m}$ ,  $\delta_{10} = 0.5\text{m}$ ,  $\delta_{11} = 0.1\text{m}$ , and  $\delta_{12} = 0.02\text{m}$ . For a snow depth below 0.1m, each layer has the  
 250 same thickness of 0.00833m. When the snow depth is above 0.2m, the thicknesses of the first and  
 251 the last layers reach their constant values of 0.01m and 0.02m respectively to reasonably resolve  
 252 the diurnal cycle and the snow/soil heat exchanges. However, to keep as much as possible the  
 253 information of an historical snowfall event, the grid thicknesses are updated only if the two first  
 254 layers or the last layer become too small or too large. This condition can be summed-up as  
 255 follows:

$$\Delta z(i) < \frac{1}{2} \min\left(\delta_i, \frac{h_{sn}}{12}\right) \quad \text{or} \quad \Delta z(i) > \frac{3}{2} \min\left(\delta_i, \frac{h_{sn}}{12}\right) \quad \forall i = \{1, 2, 12\} \quad (12)$$

257 For example, for a total snow depth of 1m, if the thickness of the top layer becomes lower than  
 258 0.005m or greater than 0.015m at the beginning of a time step, the layer thicknesses of the entire  
 259 snowpack are recalculated with Equation 11 and the snow mass and heat are redistributed  
 260 accordingly. A similar algorithm was also developed for the 6 and 9 layer cases, but these results  
 261 are not reported here. In terms of snowpack layering, the main difference with the Crocus scheme  
 262 is the fact that the total number of layers is constant, while in Crocus only the maximum number of

263 layers is specified (typically 20 or 50) and the model dynamically uses a number of layers which  
 264 varies in time within this pre-defined constraint [Vionnet et al 2012].

### 265 3.1.2. Snow compaction

266 In the new version of the snow scheme, the evolution of snow density in each layer is due  
 267 to snow compaction resulting from changes in snow viscosity [Brun et al 1989] and wind-induced  
 268 densification of near surface snow layers [Brun et al. 1997]. This wind-driven compaction process  
 269 is assumed to occur when wind velocity exceeds a threshold value that depends on snow surface  
 270 characteristics. This process is especially important for simulating the evolution of the snow  
 271 density over polar regions. Brun et al. [1997] pointed out that this process is also critical for  
 272 reproducing the snow thermal conductivity and the snow temperature profile over these regions.  
 273 Therefore, the time tendency of snow density in each layer is computed as follows:

$$274 \frac{\partial \rho_{sn}(i)}{\partial t} = \rho_{sn}(i) \frac{\sigma(i)}{\eta(i)} + \max\left(0, \frac{\rho_{wmax} - \rho_{sn}(i)}{\tau_w(i)}\right) \quad (13)$$

275 where  $\rho_{wmax}$  ( $\text{kg.m}^{-3}$ ) is the maximum density equal to  $350 \text{ kg.m}^{-3}$  below which the snow  
 276 densification occurs during wind-driven compaction,  $\tau_w$  (s) the compaction rate of this process  
 277 (Appendix B), and  $\sigma$  (Pa) the vertical stress in each layer. This stress is computed as the weight of  
 278 the overlaying layers. At the top of the snow pack, half the mass of the uppermost layer is used.  
 279 The vertical stress in each layer is then given by:

$$280 \begin{cases} \sigma(1) = \frac{g \Delta z(1) \rho_{sn}(1)}{2} \\ \sigma(i) = g \sum_{j=1}^{i-1} [\Delta z(j) \rho_{sn}(j)] \quad \forall i > 1 \end{cases} \quad (14)$$

281 The snow viscosity is a function of snow density, temperature, and liquid water content,  $W_l$   
 282 ( $\text{kg.m}^{-2}$ ), and it is given as follows:

$$283 \begin{cases} \eta(i) = \frac{\eta_0}{f_w(i)} \frac{\rho_{sn}(i)}{\rho_0} \exp(a_\eta \times \min(\Delta T_\eta, T_f - T_{sn}(i)) + b_\eta \rho_{sn}(i)) \\ f_w(i) = 1 + 10 \times \min\left(1.0, \frac{W_l(i)}{W_{lmax}(i)}\right) \end{cases} \quad (15)$$

284 where  $\eta_0$  (Pa s) is a reference viscosity equal to 7622370 Pa s,  $\rho_0$  ( $\text{kg.m}^{-3}$ ) is a reference density  
 285 equal to  $250\text{kg.m}^{-3}$ ,  $W_{lmax}$  ( $\text{kg.m}^{-2}$ ) represents the maximum liquid water holding capacity (e.g.  
 286 section 2.2) and the constants  $a_\eta = 0.1\text{K}^{-1}$ ,  $b_\eta = 0.023 \text{ m}^3.\text{kg}^{-1}$ , and  $\Delta T_\eta = 5\text{K}$ . The viscosity  
 287 dependence on snow temperature is limited according to Schleef et al. [2014] who pointed out that  
 288 the impact of snow temperature on snow densification becomes negligible at low temperatures.  
 289 The last dimensionless function,  $f_w$ , describes the decrease of viscosity in presence of liquid water.  
 290 Finally, the snowfall density is computed as previously (Equation 7).

### 291 3.1.3. Transmission of solar radiation and Snow albedo

292 The absorption of incident shortwave solar radiation,  $R_{sw}$  ( $\text{W.m}^{-2}$ ), within the pack is now  
 293 solved over three spectral bands according to Brun et al. [1992]. The first band ( $[0.3-0.8] \mu\text{m}$ )  
 294 represents the ultra-violet and visible range, while the two others ( $[0.8-1.5] \mu\text{m}$  and  $[1.5-2.8] \mu\text{m}$ )  
 295 represent two near-infrared ranges. The total net shortwave radiation,  $Q_{sn}$ , absorbed by the snow  
 296 level  $i$ , is the sum of the absorption in each spectral bands,  $k$ , and is given by:

$$297 \quad Q_s(i) = R_{sw} \sum_{k=1}^3 \left[ \omega(k)(1 - \alpha_{sn}(k)) \exp\left(-\sum_{j=1}^i [\beta_{sn}(k, j)\Delta z(j)]\right) \right] \quad (16)$$

298 where  $\omega$  is the empirical weight of each spectral bands equal to 0.71, 0.21, and 0.08 for  $[0.3-0.8]$ ,  
 299  $[0.8-1.5]$  and  $[1.5-2.8] \mu\text{m}$ , respectively. As previously, the extinction coefficient of snow,  $\beta_{sn}$ ,  
 300 depends on density and optical diameter of snow. The snow albedo,  $\alpha_{sn}$ , is a function of the snow  
 301 optical diameter and of the age of the first layer of the snowpack. The age dependency is limited to  
 302 the first band (visible range) and aims to represent the decrease of the snow albedo by impurities  
 303 from deposition in a very simple way. Indeed, trace amount of light-absorbing impurities can  
 304 significantly reduce snow albedo in the visible range but have no effect on the near-infrared range  
 305 [Warren 1984]. In each band, both the albedo and the extinction coefficient of snow are computed  
 306 according to Brun et al. [1992] as follows:

$$\begin{aligned}
\alpha_{sn}(1) &= \max \left[ 0.6, \min(0.92, 0.96 - 1.58\sqrt{d_{opt}(I)}) - \min \left( 1, \max \left( \frac{1}{2}, \frac{P_a}{P_{ref}} \right) \right) \times 0.2 \frac{A_{sn}(I)}{A_{ref}} \right] \\
\beta_{sn}(1, i) &= \max [40, 0.00192 \rho_s(i) / \sqrt{d_{opt}(i)}] \\
\alpha_{sn}(2) &= \max [0.3, 0.9 - 15.4\sqrt{d_{opt}(I)}] \\
\beta_{sn}(2, i) &= \max [100, 0.01098 \rho_s(i) / \sqrt{d_{opt}(i)}] \\
\alpha_{sn}(3) &= 0.88 + 346.2d' - 32.31\sqrt{d'} \quad \text{with } d' = \min[0.0023, d_{opt}(i)] \\
\beta_{sn}(3, i) &= +\infty
\end{aligned} \tag{17}$$

307 where  $A_{sn}$  is the age of the first snow layer expressed in days,  $A_{ref}$  a reference age set to 60 days  
308 that modulates the snow albedo decrease due to impurities,  $P_a$  (Pa) is the near surface atmospheric  
309 pressure, and  $P_{ref}$  (Pa) a reference pressure equal to 870hPa. The optical diameter of snow is  
310 simply given by Equation (10) but is now also dependent on snow age:  
311

$$312 \quad d_{opt}(i) = \min [d_{max}, g_1 + g_2 \times \rho_{sn}(i)^4 + g_3 \times \min(15, A_{sn}(i))] \tag{18}$$

313 where  $g_3$  is the rate of increase of the optical diameter of snow with snow age. It is set to  $0.5 \cdot 10^{-4}$   
314  $\text{m}\cdot\text{day}^{-1}$  through calibration. The motivation to add this snow age dependency on snow optical  
315 diameter is discussed in section 6.

316 The snow age for each layer is the time, in days, since the snow has fallen. When a  
317 snowfall event occurs, the fresh snow characteristics including its age (0 at time of snowfall) are  
318 averaged out with the snow already present in the first layer according to their respective masses.  
319 Finally, when the layer thicknesses of the entire snowpack are recalculated with Equation 11 and  
320 12, the snow age is redistributed accordingly. Typically, the age of snow in the first shallower  
321 layers remains between 0 and a dozen days during winter and increases during spring, while the  
322 last deeper layers age continuously.

### 323 3.2. *Effects of soil organic carbon on soil hydraulic and thermal properties*

324 North-Eurasian soils are rich in organic carbon as shown in Figure 1. This figure represents  
325 the soil organic carbon content of two soil horizons (0-30cm and 30-70cm) aggregated at a  $0.5^\circ$  by  
326  $0.5^\circ$  horizontal resolution and estimated from the Harmonized World Soil Database (HWSD;  
327 <http://webarchive.iiasa.ac.at/Research/LUC/External-World-soil-database/HTML/>) at a 1 km

328 resolution from the Food and Agricultural Organization [FAO 2012]. The parameterization of the  
 329 impact of soil organic carbon on hydraulic and thermal properties in ISBA is based on  
 330 pedotransfer functions of Boelter [1969], and on the work by Letts et al. [2000] and Lawrence and  
 331 Slater [2008]. The pedotransfer functions of Boelter [1969] link the soil water retention at different  
 332 pressure levels to the fiber content of a peat soil. Letts et al. [2000] describe the vertical profile of  
 333 hydraulic properties such as soil matric potential and hydraulic conductivity at saturation for a  
 334 typical organic soil. The hydraulic properties change sharply from the near surface where peat is  
 335 weakly decomposed (fibric soil) to the sub-surface with moderately and well decomposed peat  
 336 (hemic and sapric soils respectively). Lawrence and Slater [2008] proposed a linear combination  
 337 of such soil organic properties with the standard mineral soil properties.

338 In ISBA, before averaging soil organic with mineral properties, a typical peat soil profile is  
 339 computed for the model soil grid using a power function for each hydraulic property,  $\alpha_{peat}$ , found  
 340 in Table 1. For each soil layer  $i$ , this function is described as:

$$341 \quad \alpha_{peat}(i) = \alpha_{fibric} z(i)^\beta \quad \text{with} \quad \beta = \frac{\ln(\alpha_{sapric} / \alpha_{fibric})}{\ln(d_{sapric} / d_{fibric})} \quad (19)$$

342 where  $z$  (m) is the depth of the considered soil grid node,  $\alpha_{fibric}$  and  $\alpha_{sapric}$  the fibric and sapric  
 343 parameter values (Table 1),  $d_{fibric}$  (m) the depth arbitrarily set to 0.01m where the profile starts to  
 344 depart from fibric values, and  $d_{sapric}$  (m) the depth of 1m where the soil properties reach the sapric  
 345 values according to Letts et al. [2000].

346 To determine the organic fraction of soil, the density profile of the soil carbon must be  
 347 known for the entire soil grid. Using the HWSD database, the soil carbon densities in the first  
 348 0.3m,  $\rho_{top}$  ( $\text{kg}\cdot\text{m}^{-3}$ ), and the remaining 0.7m below,  $\rho_{sub}$  ( $\text{kg}\cdot\text{m}^{-3}$ ), are known:

$$349 \quad \rho_{top} = \frac{S_{top}}{\Delta d_{top}} \quad \text{and} \quad \rho_{sub} = \frac{S_{sub}}{\Delta d_{sub}} \quad (20)$$

350 where  $S_{top}$  and  $S_{sub}$  ( $\text{kg}\cdot\text{m}^{-2}$ ) are the topsoil and subsoil organic carbon contents respectively,  $\Delta d_{top}$   
 351 and  $\Delta d_{sub}$  (m) the thicknesses of each observed soil horizon (0.3 and 0.7m respectively). We  
 352 extrapolate the density present below 1m from this observed near-surface profile (Equation 20).



353 The extrapolation assumes that the carbon profile decreases sharply with soil depth according to a  
 354 power function. The shape of this function is given by the observed profile if the topsoil organic  
 355 carbon density is superior to the subsoil density. Otherwise, the density of soil carbon below a 1m  
 356 depth,  $\rho_{deep}$  (kg.m<sup>-3</sup>), is taken equal to the subsoil density:

$$\begin{aligned}
 & \rho_{deep} = (1 - \delta)\rho_{sub} + \delta \frac{s_{top} + s_{sub}}{\Delta d_{deep} - \Delta d_{top} - \Delta d_{sub}} \left[ \left( \frac{\Delta d_{deep}}{\Delta d_{top} + \Delta d_{sub}} \right)^\beta - 1 \right] \\
 & \delta = \begin{cases} 0 & \forall \rho_{top} \leq \rho_{sub} \\ 1 & \forall \rho_{top} > \rho_{sub} \end{cases} \quad \text{and} \quad \beta = \frac{\ln[s_{top}/(s_{top} + s_{sub})]}{\ln[\Delta d_{top}/(\Delta d_{top} + \Delta d_{sub})]}
 \end{aligned}
 \tag{21}$$

358 where  $\Delta d_{deep}$  (m) is an infinite soil thickness taken arbitrarily equal to 1000m.

359 Finally, the soil carbon density profile,  $\rho_{soc}$  (kg.m<sup>-3</sup>), over the entire soil grid is computed  
 360 using these three soil horizons and a simple linear interpolation at each grid node that conserves  
 361 the total soil carbon mass (Figure 2). The fraction of the soil that is organic,  $f_{soc}$ , in each layer is  
 362 determined assuming this simple relationship:

$$f_{soc}(i) = \frac{\rho_{soc}(i)}{(1 - w_{sat,peat}(i))\rho_{om}}
 \tag{22}$$

364 where  $\rho_{om}$  (kg.m<sup>-3</sup>) is the pure organic matter density equal to 1300 kg.m<sup>-3</sup> [Farouki 1986] and  
 365  $w_{sat,peat}$  the porosity of the peat soil profile computed using Equation 19 and Table 1. As in  
 366 Lawrence and Slater [2008], this fraction is used to combine the standard mineral soil properties  
 367 with soil organic properties using weighted arithmetic or geometric averages, depending on the  
 368 parameter (Table 1). An example of this method is shown in Figure 2 for soil porosity, soil  
 369 saturated hydraulic conductivity and soil heat capacity.

## 370 4. Local scale evaluation of snow processes at the Col de Porte site (France)

### 371 4.1. Experimental data set

372 The Col de Porte field site (45°17'N, 05°45'E) is located at an elevation of 1325m in the  
 373 French Alps near Grenoble [Morin et al. 2012]. It consists in a 50m by 50m square covered by  
 374 grass, mowed approximately once a month in summer depending on its growth rate. Soil textures

375 (30% clay, 60% sand) are characteristic of a sandy-clay-loam soil that is very poor in organic  
376 carbon. For this reason, this site is only used to evaluate the effect of changes in snow  
377 parameterizations while changes in soil physics can be not tested. The atmospheric forcing  
378 variables (air temperature, rain and snow rates, air humidity, atmospheric pressure, wind speed,  
379 long-wave and short-wave incident radiation) are available at a one hour time step from August 1<sup>st</sup>,  
380 1993 to July 31, 2011. It consists of a combination of in-situ measurements, roughly from  
381 September to June each year, and the regional reanalysis SAFRAN from June to September each  
382 year (see Morin et al. [2012] for details).

383         The Col de Porte dataset includes many observations at a daily time step for evaluating  
384 land surface models. In this study, the observed snow depth, surface albedo and soil temperature at  
385 10 cm are used to evaluate model simulations over the entire period. The snow water equivalent  
386 (SWE) is also used for this model evaluation but daily values are only available from 2001 to  
387 2011. Snow depth is measured using ultra-sound depth gauges with an accuracy of 1cm. Surface  
388 albedo is computed as the total daily reflected solar flux divided by the total daily incoming solar  
389 flux. We estimate the uncertainty in surface albedo to be about 10% based on the 10% uncertainty  
390 in observed radiative fluxes reported by Morin et al [2012]. Soil temperature is measured using  
391 automatic probes with an accuracy of 0.1K. SWE is measured using a cosmic ray sensor placed on  
392 the ground and exhibits an uncertainty of 10%. Three skill scores are used to compare model  
393 results to the observations. The mean annual bias measures the capability of the model to  
394 represent the observed mean. To evaluate the model ability to represent the observed day to day  
395 variability, two statistical quantities are used; the square correlation ( $r^2$ ), and the centered root  
396 mean square error (c-rmse). It is computed by subtracting the simulated and observed annual  
397 means from their respective time series before computing a standard root mean square error.

#### 398 4.2. *Model configuration*

399         Four simulations were done to evaluate the effect of the different changes in the snow  
400 parameterization detailed in section 3:

- 401 • *CTL* uses Boone and Etchevers [2001] formulation for snow layering (3 layers), snow  
402 compaction, and snow albedo as described in section 2.2
- 403 • *SNL* is similar to *CTL* in terms of snow compaction and albedo but uses the new snow  
404 layering with 12 snow layers described in section 3.1.1.
- 405 • *CPT* uses 12 snow layers as in *SNL* but the compaction and the wind-induced densification  
406 of near surface snow layers are computed using formulations of Brun et al [1989 and 1997],  
407 both described in section 3.1.2.
- 408 • *NEW* uses all the package of snow equations described in section 3.1: 12 snow layers, the  
409 new snow compaction/densification, but also the spectral representation of the snow albedo  
410 (section 3.1.3).

411 For all of the simulations, the snow is assumed to cover the entire grid-cell (the snow  
412 fraction set to 1) as long as the snow remains present. The effective roughness length of snow is  
413 set to its usual value of 0.001m. The grid-cell is assumed to be entirely covered by grass with a  
414 root depth of 1m, the leaf area index varies from 0.1 in winter to 1 in summer, and the snow-free  
415 surface albedo is prescribed as 0.2. The model calculates soil temperature, moisture and ice  
416 content in each of the 14 soil layers corresponding to a soil depth of 12m. The model was run with  
417 a 15-minute time step from August 1<sup>st</sup>, 1993 to July 31, 2011. The model was spun-up by  
418 performing fifty iterations of the first two years (August, 1993 to July, 1995). This spin-up  
419 represents a total of one hundred years, and this was determined to guarantee that the water and  
420 heat profiles were equilibrated over the 12m soil depth of ISBA. Results are then evaluated over  
421 the entire period.

#### 422 4.3. Results

423 Figure 3 and 4 show an overview of the four simulations performed at the Col de Porte in  
424 terms of snow depth, SWE, surface albedo and soil temperature at 10cm. A quick look at the time  
425 series indicates that all of the model versions match the observations relatively well. However,  
426 annual statistics show a clear hierarchy between the four experiments. The snow depth statistics

427 shows that the new snow compaction/densification algorithm has a positive impact on the  
428 snowpack simulation. Indeed, both the *CPT* and *NEW* experiments exhibit the lowest bias and c-  
429 rmse for twelve of the eighteen years. However, the comparison to SWE data does not allow a  
430 discrimination between the four simulations, even if the c-rmse of the *NEW* experiment is the best  
431 for seven of the ten years. The surface albedo from the *NEW* simulation is clearly better than the  
432 albedo from the other experiments: bias and c-rmse are the best for all years (Figure 4). The soil  
433 temperature bias and c-rmse are also reduced by the *NEW* experiment (for ten of seventeen years)  
434 compared to the other simulations. Thus, accounting for different spectral bands within the snow  
435 albedo calculation has a significant positive impact on the energy balance of the snow-soil system.

436         The average seasonal cycle of snow depth, SWE, surface albedo and soil temperature at  
437 10cm represented in Figure 5 highlights the qualities and weaknesses of the different  
438 parameterizations by focusing on the snow season (October to May). The corresponding statistics  
439 for the winter (DJF), spring (MAM) and the entire period are given in Table 2. The comparison of  
440 *SNL* to *CTL* indicates that the increase in number of snow layers from 3 to 12 improves the snow  
441 depth, SWE and winter soil temperature simulation. Change in snow compaction (from *SNL* to  
442 *CPT*) improves the seasonal cycle of snow depth and SWE and especially the maximum value.  
443 The seasonal and total biases in Table 2 verify this result and show the same behavior for winter  
444 soil temperature, although it is difficult to see visually from Figure 5. For these three variables, the  
445 simulated time variability is also improved from *CTL* to *SNL* to *CPT* as shown by the other  
446 seasonal and total scores (c-rmse and  $r^2$ ) in Table 2. Finally, the new spectral albedo scheme (from  
447 *CPT* to *NEW*) has a drastic impact on the snowpack simulation in spring. As shown by Figure 5  
448 and Table 2, the new spectral albedos clearly improve the simulation of other variables during this  
449 period. They induce a sharp springtime snowmelt with a strong decrease in snow depth and SWE.  
450 The snow insulation during spring is thus less important and allows the soil surface to warm up  
451 faster. As a result, the model is capable of reproducing the strong soil warming observed in April

452 (Figure 5). Not surprisingly, the soil temperature skill scores for spring and the whole period are  
453 drastically improved although there is a slight degradation in winter.

454 Figure 6 shows daily mean time series of the snow density and temperature profiles  
455 averaged over the snow season for each experiment. With only 3 snow layers (*CTL*), the density  
456 distribution is more uniform than using the new snow layering scheme with 12 layers (*SNL*). The  
457 significant densification of the bottom layers in *SNL* is the main process responsible for the snow  
458 depth and SWE improvements observed in Figure 5 and Table 2. In addition, the better  
459 representation of the vertical density profile, that results in less dense and thus more insulating  
460 surface snow layers from November to February, leads to a better insulation of the bottom snow  
461 layer from the atmosphere and thus to higher temperatures of the bottom snow and top soil layers.  
462 This explains the skill scores improvement found in winter soil temperature in Table 2. The new  
463 snow compaction scheme (*CPT*) tends to increase the density contrast between the top and the  
464 bottom snow layers. The snowpack is also denser than with *SNL* leading to the strong decrease in  
465 snow depth observed in Figure 5 and to the better skill scores in snow depth over each period  
466 (Table 2).

467 *CPT* also results in a small warming at the bottom of the snowpack which slightly heats the  
468 soil temperature compared to *SNL*. Finally, the spectral albedo scheme (*NEW*) has a limited effect  
469 on the snow density profile but results in a slightly colder snowpack than in *CPT* and even *SNL*  
470 (not shown) due to the large daily winter albedos seen in Figure 5. This is the main reason for the  
471 lower winter soil temperatures with *NEW* than *CPT* and *SNL* (Table 2).

## 472 5. Simulations over North-Eurasia

### 473 5.1. Numerical experiment design and observational dataset

474 The experimental design used here is close to that proposed by Brun et al. [2013]. The  
475 region considered (35°N to 85°N, 25°E to 180°E) covers Eastern-Europe, Russia and Siberia  
476 (Figure 7). The ISBA land surface model is run at a 0.5° by 0.5° spatial resolution using the  
477 Interim Re-Analysis (ERA-I; <http://www.ecmwf.int/en/research/climate-reanalysis/era-interim>)

478 [Dee et al. 2011]. ERA-I meteorological variables are extracted with a 3-hourly frequency in order  
479 to represent the diurnal cycle. This reanalysis covers the time period from 1979 to the present.  
480 Many details about ERA-I can be found in Dee et al [2011] and an evaluation of its performance is  
481 provided in Berrisford et al. [2011]. For precipitation, the monthly ERA-I precipitation are  
482 rescaled to match the observed Global Precipitation Climatology Center (GPCC) Full Data  
483 Product V5 (<http://gpcc.dwd.de>) as proposed by Decharme and Douville [2006a]. This method  
484 conserves the 3-hourly chronology of the ERA-I precipitation but ensures a reasonable monthly  
485 amount [Szczypta et al. 2012]. Brun et al. [2013] pointed out the significantly better performance  
486 of this ERA-I scaled GPCC forcing product in simulating North-Eurasian snowpack variables  
487 compared to the ERA-I precipitation or other “state of the art” global scale atmospheric forcings.

488 To evaluate snow and soil temperature simulations, several in-situ dataset are used. As in  
489 Brun et al. [2013], the Historical Soviet Daily Snow Depth (HSDSD;  
490 [http://nsidc.org/data/docs/noaa/g01092\\_hsd/d/index.html](http://nsidc.org/data/docs/noaa/g01092_hsd/d/index.html)) compiled by Amstrong [2001] was used  
491 in the current study. It consists in daily snow depth measurements taken at synoptic stations  
492 following the World Meteorological Organization (WMO) standards. WMO requires the  
493 measurements to be taken in bare ground open areas or clearings with regular grass cutting. These  
494 snow depth data are therefore representative of open areas of bare ground or those covered with  
495 very short grass. This dataset starts in 1881 with a few stations and ends in 1995. Considering that  
496 ERA-I starts in 1979, the model simulations are done from 1979 according to Brun et al. [2013].  
497 263 HSDSD stations are available over this period with approximately half of them without any  
498 missing data. We chose to use only the stations where the difference between the local and the  
499 ERA-I elevation is less than 100m to avoid temperature biases for instance that would be directly  
500 due to the low resolution of ERA-I. We also only kept the stations where the number of days with  
501 a non zero snow depth measurement over the entire period is superior to 100 days, and that have at  
502 least 8 days with snow measurement per year. With this filter, the number of available stations

503 decreases to 158, which remains acceptable. Most stations are located in Russia and Western-  
504 Siberia with only a few in Eastern-Siberia (Figure 7).

505 The second source of observations is the Russian Historical Soil Temperature (RHST)  
506 dataset compiled by Zhang et al. [2001] over Siberia  
507 (<http://data.eol.ucar.edu/codiac/dss/id=106.ARCSS078>). Data coverage extends from the 1800s  
508 through 1990, but is not continuous. We compared our model results over the 1979-1990 period.  
509 Similar to snow depth, soil temperature stations are subject to WMO standards and are located in  
510 open area sites. We used the same criteria as for snow depth. Only stations with local elevations  
511 close to the ERA-I altitude (less than 100m difference) are used. In addition, only stations with at  
512 least 36 months of observations (at least 3 years out of 12) are kept. Most soil temperature sites are  
513 collocated with snow depth sites (Figure 7). Measurements were taken at 20cm, 80cm, 160cm and  
514 320cm depth. For each depth, 95, 48, 48, and 82 stations, respectively, were available for model  
515 evaluation. The spatial distribution of these stations is shown in Figure 7 for 20cm and 160cm  
516 depths.

517 To quantify the capability of the model to simulate the permafrost characteristics, three  
518 datasets are used. The first dataset is the Circum-Artic Map of Permafrost and Ground Ice  
519 Conditions (<http://nsidc.org/data/ggd318>) edited by Brown et al. [2002]. This dataset is available at  
520 a 0.5° by 0.5° resolution and shows the continuous, discontinuous, isolated and sporadic  
521 permafrost boundaries. The second dataset gives access to in-situ observations on active layer  
522 thickness collected by the Circumpolar Active Layer Monitoring (CALM;  
523 <http://www.gwu.edu/~calm/>) since the 1990s to 2015 [Brown et al. 2000]. Over the studied  
524 domain, 233 monitoring sites are available. To compare with simulations performed at a 0.5° by  
525 0.5° resolution, 89 virtual stations have been computed from the 233 original sites by averaging all  
526 stations in each 0.5° by 0.5° grid-cells. The last dataset is an estimate of the active layer thickness  
527 over North-West-Siberia before the 1990s. This data set is based on the map of landscapes and  
528 permafrost conditions in Yakutia (<http://doi.pangaea.de/10.1594/PANGAEA.808240>). It gives

529 access to the mean and standard deviation of the most probable active layer thickness in each grid  
530 box at  $0.5^\circ$  by  $0.5^\circ$  resolution. All details can be found in Beer et al. [2013].

## 531 5.2. *Model configuration*

532 Three experiments using the ISBA land surface model forced by the ERA-I scaled GPCP  
533 atmospheric dataset are performed using the same configuration. In addition to the *CTL* (old snow  
534 scheme) and *NEW* (new snow scheme) experiments already described in section 4, we performed  
535 one simulation using the parameterization of the impact of the soil organic carbon on the  
536 hydrologic and thermal soil properties. This last experiment, called *NEW-SOC*, uses the new snow  
537 and soil-property schemes described in section 3.1 and 3.2, respectively. As previously, the model  
538 determines the temperature, liquid water and ice content evolution in each of the 14 soil layers  
539 corresponding to a total soil depth of 12m. The model is run with a 15-minute time step from  
540 January 1<sup>st</sup>, 1979 to Decembre 31, 2013. The model's spin-up uses twenty iterations of the first  
541 five years (1979 to 1983) of the atmospheric forcing, representing a total of one hundred years.

542 In ISBA, we use a series of twelve sub-grid independent patches per grid cell in order to  
543 account for land cover heterogeneity. Land cover parameters such as Leaf Area Index (LAI),  
544 vegetation height, vegetation/soil albedos, and rooting depth are prescribed for each sub-grid  
545 patch. The dominant patches present in the model over the Northern-Eurasian region are bare soil,  
546 grassland/tundra, deciduous forest, coniferous boreal forest, and C3 crops in the South. The  
547 fraction of each surface type within each grid box is used to compute the grid box average of the  
548 water and energy budgets. Some other processes, such as surface runoff, dripping from the canopy  
549 reservoir, and soil infiltration account for sub-grid parameterizations. More details can be found in  
550 Decharme and Douville [2006b] and Decharme et al. [2013].

551 For all of the simulations, the grid-cell fraction covered by snow evolves according to the  
552 simulated snow depth and is different for bare soil and vegetated areas (Appendix C) in each land  
553 cover patch. As was the case for the Col de Porte experiment, the effective roughness length of  
554 snow retains its usual value of 0.001m. The land surface parameters used by ISBA are specified



555 according to the 1-km resolution ECOCLIMAP-II database [Faroux et al., 2013]. LAI, vegetation  
556 height, and vegetation/soil albedos are prescribed for the twelve vegetation sub-grid patches based  
557 on a mean annual cycle at a 10-day time step. The rooting depth is specified for each vegetation  
558 type according to Canadell et al. [1996]. It ranges from 0.5m to 1.5m for tundra and temperate  
559 grassland, and from 2m to 3m for forest. The soil textural properties are given by the HWSO  
560 database at 1 km resolution while the topographic information is specified according to the 30-  
561 arcsecond resolution GTOPO30 data set.

### 562 5.3. Results

563 Figure 7 presents a quantitative comparison between the observed and simulated snow  
564 depth and soil temperature over Northern-Eurasia. Because in-situ observations were collected in  
565 bare ground open areas and/or clearings with regular grass cutting following the WMO standards  
566 as mentioned previously, they are compared to snow depths and soil temperature profiles  
567 simulated by the ISBA bare soil sub-grid patch alone. This patch exhibits conditions which are  
568 closest to those at the corresponding field sites, as is generally the case for ISBA in this kind of  
569 comparison [Decharme et al. 2013]. The simulation represented here is the *NEW-SOC* experiment  
570 that seems to capture well the snow depth and soil temperature spatial distributions. For snow  
571 depth, the latitudinal gradient is well respected. The lower soil temperature along a southwest-  
572 northeast transect is also well simulated.

573 The seasonal cycles of daily snow depths and monthly soil temperatures (Figure 8) clearly  
574 show the biases of the *CTL* simulation and the improvements due to the new snow and soil  
575 representations. The seasonal cycles and the total skill scores are computed using the  
576 measurements and simulations for all stations over the entire observed periods. ISBA globally  
577 underestimates the snow depth from December through February with no clear difference between  
578 *CTL* and *NEW* (or *NEW-SOC*). However, the springtime snow melting is drastically improved by  
579 the new snow scheme inducing a better simulated seasonality. This fact is confirmed by some  
580 other quantitative comparisons. The average number of days per year with observed snow on the

581 ground for all in-situ stations is 150.7 days. *CTL* simulates 158.7 days against 151.5 days for  
582 *NEW*. On average, the last day of the snow season is day number 281.6 when starting on July first.  
583 *CTL* goes beyond this date by more than 9 days while for *NEW* it is only 2 days (day number 283).  
584 These results are consistent with the model evaluation at the Col de Porte field site (section 4). As  
585 could be expected also, the new physical soil properties (*NEW-SOC*) play a minimal role in the  
586 snow depth simulation. The seasonal cycle of the soil temperature profile confirms that the new  
587 snow scheme induces a warmer soil in winter compared to *CTL*, and it strongly reduces the cold  
588 bias of *CTL*. The effect of soil organic carbon is especially observable during spring and summer.  
589 *NEW* exhibits a warm bias for each soil horizon while *NEW-SOC*, with more insulating soils,  
590 reduces this weakness.

591         These improvement in snow depth and soil temperature are confirmed by the spatial  
592 distributions of their seasonal skill scores (bias and c-rmse). Figure 9 shows the spatial  
593 distributions of snow depth seasonal skill scores (bias and c-rmse) during winter and spring. No  
594 clear differences among these simulations appear in winter while the bias and c-rmse of many  
595 stations are improved in spring by the new snow scheme. The springtime snow depth is simulated  
596 in an acceptable manner by *NEW*, while *CTL* exhibits a significant overestimation. This fact is  
597 confirmed by total scores given in each of the panels. In winter, regardless of the experiments,  
598 *ISBA* underestimates snow depth measurements at many stations, especially in the Northern and  
599 Western parts of the domain (Figure 9).

600         The spatial distribution of soil temperature seasonal skill scores simulated at 20 cm and 160  
601 cm depth during winter is given in Figure 10. Regardless of the region, the generalized cold bias  
602 found over all stations with *CTL* is drastically reduced with the new snow scheme and the  
603 interannual variability (c-rmse) is largely improved. In summer (Figure 11), as was already shown  
604 in Figure 8, *NEW-SOC* is in better agreement with observations compared to *NEW* regardless of  
605 the soil horizon (lower c-rmse) even if a slight cold bias appears at the subsurface as shown by the  
606 negative total bias found at 320cm depth. The *NEW* experiment overestimates the temperature

607 profile measurements at many stations near the surface, but less-so at a 320 cm depth. So, it seems  
608 that the subsurface cooling in the *NEW-SOC* experiment is too intensive. But in fact at 320 cm  
609 depth, the simulated soil temperature in the western part of the domain remains quasi unchanged  
610 between *NEW-SOC* and *NEW*. The best total scores found on Figures 8 and 11 without soil  
611 organic carbon by the *NEW* experiment are in fact due to error compensation between the cold and  
612 warm biases simulated in the western and eastern part of the domain, respectively.

613         The effect of soil organic carbon content on soil temperature profile is also especially  
614 observable in terms of the simulated permafrost characteristics. The observed and simulated  
615 locations of permafrost boundaries are compared in Figure 12. Regardless of the experiment, ISBA  
616 generally simulates acceptable boundaries even if the permafrost limit extends slightly too far  
617 south in the western part of the domain. This figure also shows the spatial distribution of active  
618 layer thicknesses simulated by the *NEW* and the *NEW-SOC* experiments. The active layer  
619 thickness in the model is computed as the maximum depth reached each year by the 0°C isotherm  
620 in the soil approximated via a linear interpolation between the last positive temperature node going  
621 down from the surface and the first negative temperature node. As expected from the lower  
622 summer soil temperatures with *NEW-SOC* (Figure 9 and 11), the active layer is shallower.  
623 However, this comparison with the limits of different permafrost types does not allow to determine  
624 which simulation leads to the most accurate active layer thicknesses. The comparison with the  
625 CALM data given in Figure 12 seems to show that *NEW-SOC* simulates a more accurate spatial  
626 distribution of the active layer thickness. This result is confirmed by Figure 13 that shows the  
627 estimated and simulated active layer thicknesses over the Yakutia region. Estimations from Beer et  
628 al. [2013] present a strong latitudinal gradient with an increase in active layer thickness from the  
629 north to the south. Both experiments exhibit such profiles. However, the active layer thickness  
630 simulated by *NEW-SOC* is in better agreement with these estimations than those by *NEW*. The  
631 latitudinal zonal average confirms this result.

## 632 **6. Discussion and Conclusion**

633 In this study, the impact of improved representation of snowpack and soil properties in the  
634 ISBA LSM to simulate snow characteristics and soil temperature profiles over cold regions was  
635 analysed. ISBA's representations of snow layering, albedo, and compaction were updated by  
636 incorporating some parameterizations of the detailed Crocus snowpack model. In addition, a  
637 simple parameterization of the soil organic carbon effect on hydraulic and thermal soil properties  
638 was introduced based on previous work [Boelter 1969; Letts et al. 2000; Lawrence and Slater  
639 2008]. The model is evaluated first over the Col de Porte field site in the French Alps [Morin et al.  
640 2012] in order to isolate the changes in the snowpack parameterization, and second over the North-  
641 Eurasian region to analyze the model's ability to simulate snow depth, soil temperature profile and  
642 permafrost characteristics.

643 Changes in the snowpack parameterizations induce noticeable improvements in the  
644 simulated snow depth, SWE, surface albedo and soil temperature at the Col de Porte (field) site.  
645 The new snow layering algorithm with 12 layers permits a refinement of the vertical distribution of  
646 density and temperature in the snowpack leading to slight improvements in simulated snow depth,  
647 SWE, and soil temperature during winter. The densification of the snowpack with the new  
648 compaction scheme, which increases the density contrast between the top and the bottom snow  
649 layers, has a significant positive impact on snow depth and winter soil temperature. Finally, the  
650 new spectral albedo scheme clearly improves the simulation of the springtime surface albedo that  
651 allows a better simulation of the snowpack characteristics and soil temperature during melting at  
652 the end of the snow season.

653 It must be noted that the large improvement in snow albedo in spring is mainly due to the  
654 use of snow age in the diagnostics of the optical diameter of snow (Equation 18). Without this  
655 parameterization, the surface albedo is strongly overestimated in winter and, to a lesser extent in  
656 spring at the Col de Porte field site, with a larger bias and c-rmse for all variables compared to the  
657 new version of ISBA (not shown). The optical diameter of snow strongly controls the near-infrared  
658 albedo, while impurities mostly affect the albedo in the visible spectrum [Wiscombe and Warren

659 1981]. This increase of snow optical diameter with time is necessary to represent well the decrease  
660 in spectrally integrated albedo with age. However, the increase of snow optical diameter is not  
661 only a function of snow density as parameterized by Anderson [1976] in Equation (10), but it is  
662 also due to snow metamorphism, which is macroscopically driven by snow temperature and snow  
663 thermal gradients. Several complex parameterizations exist to explicitly represent the evolution of  
664 snow optical diameter according to these processes [e.g. Carmagnola et al. 2014]. Nevertheless,  
665 for the sake of simplicity, we just use a snow age dependency in the diagnostic of snow optical  
666 diameter with a limitation at fifteen days (Equation 18). This simple diagnostic allows the model  
667 to reasonably match the explicit computation of the optical diameter of snow simulated in the  
668 Crocus model (not shown). The good results of the ISBA model at the Col de Porte field site  
669 reinforce this choice.

670         The positive impacts of the new ISBA snow scheme are confirmed when tested over the  
671 North-Eurasian region with an important number of open field in-situ snow depth and soil  
672 temperature stations. Winter snow depths are slightly better simulated with the new version and  
673 the winter soil temperature cold bias obtained with the old version of ISBA is clearly reduced. This  
674 fact confirms that the physics used in snow schemes is of primary importance for adequately  
675 simulating the snow insulating effect that prevents soil from getting too cold in winter [Slater et al.  
676 2001; Luo et al. 2003; Gouttevin et al. 2012; Paquin and Sushama 2015]. Another important  
677 impact of changes in the ISBA snow scheme over the North-Eurasian region is seen in spring  
678 when the snowmelt is well reproduced. As shown over the Col de Porte (field) site, this is mainly  
679 due to the new parameterization of spectral snow albedo.

680         Nevertheless, regardless of the model version used, simulated winter snow depths are  
681 generally underestimated compared to in situ observations. The cause of this underestimation is  
682 not trivial. The first source of uncertainty can be attributable to the GPCC precipitation  
683 measurements that do not account for wind undercatch leading to a possible underestimation of  
684 solid precipitation during winter [Adam and Lettenmaier 2003, Brun et al., 2013]. Besides

685 uncertainties related to the atmospheric forcing, the snow depth underestimation can be due to the  
686 non-explicit representation of snow metamorphism. Indeed, in similar experimental conditions  
687 over the Northern Eurasian region, the winter snow depth simulated by the detailed Crocus  
688 snowpack model did not exhibit the same problem [Brun et al. 2013] and the main remaining  
689 difference between Crocus and ISBA is now restricted almost entirely to the explicit simulation of  
690 snow metamorphism. In Crocus, the viscosity of layers composed of faceted crystals and depth  
691 hoar snow types is increased [Vionnet et al., 2012], which leads to reducing the overall  
692 compaction rate of snowpack undergoing temperature conditions conducive to such snow types,  
693 and this is consistent with the situation described above.

694         Taking into account soil organic carbon in soil physical properties logically plays a  
695 minimal role in the simulated snowpack behaviour. However, this process has drastic impacts on  
696 the summer soil temperature profile because it allows the soil to remain cool during spring and  
697 summer as shown in previous studies [Bonan and Shugart 1989; Lawrence and Slater 2008;  
698 Dankers et al. 2011]. Consequently, the spatial distribution of the permafrost active layer thickness  
699 simulated by the new version of ISBA is in better agreement with estimations from Beer et al.  
700 [2013] over the Yakutia region. This result is in agreement with Paquin and Sushama [2015] who  
701 showed that the hydraulic and thermal properties of soil organic carbon partly control the thickness  
702 of the active layer during summer. However, spatial observations of permafrost characteristics on  
703 the global scale are still very scarce, and if available, they are static and don't allow the study of  
704 long term trends and inter-annual variability.

705         This model validation should ideally be extended over all cold regions (e.g. North America,  
706 Greenland, etc...) but considering that North-Eurasia is representative of such regions, some  
707 important conclusions are confirmed by this study:

- 708         • An adequate simulation of snow layering and snow compaction/densification is important  
709         in order to represent well winter snowpack characteristics and the soil temperature profile.

710 • Snow albedo strongly controls the simulation of the springtime snow characteristics and the  
711 melting timing.

712 • To account for soil organic carbon in terms of the soil physical properties drastically  
713 impacts the simulation of summer soil temperature profile and hence the permafrost active layer  
714 thickness and its spatial distribution.

715 Finally, these conclusions underscore the fact that the representation of snowpack characteristics  
716 and soil thermal processes are of primary importance for studying permafrost vulnerability under  
717 climate change conditions, especially if the continental carbon cycle is considered due to the  
718 strong interaction between soil thermal processes and soil organic carbon decomposition with  
719 release of greenhouse gases.

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## APPENDIX A

### *Snow thermal conductivity*

729  
730  
731 The snow thermal conductivity is computed as a function of snow density  
732 following Yen [1981]. It also accounts for vapor transfer in the snow using a simple  
733 parameterization from Sun et al. [1999]. This process is especially important at low snow densities  
734 and at high altitude. So the snow thermal conductivity,  $\lambda_{sn}$  ( $\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ ), in each layer is given by:

$$735 \quad \lambda_{sn}(i) = \lambda_{ice} \left( \frac{\rho_{sn}(i)}{\rho_w} \right)^{1.88} + \frac{P_0}{P_a} \times \max \left( 0, k_1 - \frac{k_2}{T_{sn}(i) - k_3} \right) \quad (\text{A1})$$

736 where  $\lambda_{ice}$  ( $\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ ) is the thermal conductivity of ice equal to  $2.2 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ ,  $\rho_w$  ( $\text{kg}\cdot\text{m}^{-3}$ ) the  
737 water density,  $P_a$  (Pa) the air pressure,  $P_0$  (Pa) a reference pressure equal to  $1000\text{hPa}$ , and the  
738 coefficients  $k_1 = -0.06023 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$ ,  $k_2 = 2.5425 \text{ W}\cdot\text{m}^{-1}$  and  $k_3 = 289.99 \text{ K}$ .

## APPENDIX B

739 Wind-induced densification of near surface snow layers Following Brun et al. [1997], the  
740 compaction rate,  $\tau_w$ , of wind-induced densification of near surface snow layers is computed using  
741 several steps. First, a mobility index,  $\Gamma_{mob}$ , that describes the potential for snow erosion for each  
742 snow layer is computed as a function of snow density:  
743

$$744 \quad \Gamma_{mob}(i) = a_{mob} \left[ 1.0 - \max \left( 0, \frac{\rho_s(i) - \rho_{snmin}}{\rho_{mob}} \right) \right] \quad (\text{B1})$$

745 where  $\rho_{snmin} = 50\text{kg}\cdot\text{m}^{-3}$  is the minimum density of snow,  $\rho_{mob}$  a reference density of  $295\text{kg}\cdot\text{m}^{-3}$ ,  
746 and the dimensionless constant  $a_{mob} = 1.25$ . Secondly, a wind-driven compaction index,  $\Gamma_w$ ,  
747 combining the mobility index and the near surface atmospheric wind speed:

$$748 \quad \Gamma_w(i) = 1 - a_\Gamma \exp(-b_\Gamma \kappa_v V_a) + \Gamma_{mob}(i) \quad (\text{B2})$$

749 where  $\kappa_v = 1.25$  is a dimensionless coefficient for gust diagnosis from average wind speed, and the  
750 constants  $a_\Gamma = 2.868$  and  $b_\Gamma = 0.085 \text{ s}\cdot\text{m}^{-1}$ . A positive value of  $\Gamma_w$  indicates that wind-driven  
751 compaction can occur. Compaction rate from the surface is then propagated to the layers beneath,

752 following an exponential decrease, until it meets a snow layer having a negative wind-driven  
 753 compaction index. For each layer, this compaction rate is computed as follows:

$$754 \quad \tau_w(i) = \frac{2\kappa_v \pi_\tau}{f_\tau(i)} \quad \text{with} \quad f_\tau(i) = \max(0, \Gamma_w(i)) \times \exp\left(-a_\tau \sum_{j=1}^i (\Delta z(j)(b_\tau - \Gamma_w(j)))\right) \quad (\text{B3})$$

755 where  $\pi_\tau$  (s) is a time constant of one day, and the constants  $a_\tau = 10$  and  $b_\tau = 3.25$ .

## 756 APPENDIX C

### 757 Grid-cell snow fraction

758 At regional and/or global scale the snow fraction,  $p_{sn}$ , for each patch of the ISBA land  
 759 surface model is computed as the sum between the bare ground snow covered fraction,  $p_{sng}$ , and  
 760 the fraction of vegetation covered by snow,  $p_{snv}$ , weighted by the vegetation fraction of the patches  
 761 covered by vegetation,  $f_{veg}$ . The snow fraction is thus computed as follows:

$$762 \quad p_{sn} = (1 - f_{veg})p_{sng} + f_{veg}p_{snv} \quad \text{with} \quad \begin{cases} p_{sng} = \min(1, h_{sn}/h_{sng}) \\ p_{snv} = h_{sn}/(h_{sn} + w_{snv}z_{0veg}) \end{cases} \quad (\text{C1})$$

763 where  $h_{sn}$  (m) is the total snow depth,  $h_{sng}$  (m) a ground snow depth threshold sets to 0.01m,  $z_{0veg}$   
 764 (m) the vegetation roughness length, and  $w_{snv}$  a coefficient set to 2.  $f_{veg}$  is specified for each  
 765 vegetation patch. It is equal to 0.0 for bare soil, 0.95 for grassland/tundra as well as for temperate  
 766 and boreal forest, and varies exponentially according to the leaf area index (LAI) for crop types.  
 767  $z_{0veg}$  varies for each vegetation type and is computed from typical vegetation height,  $h_{veg}$ , as  
 768 follows:

$$769 \quad z_{0veg} = \max(0.001, 0.13 \times h_{veg}) \quad (\text{C2})$$

770 For woody vegetation,  $h_{veg}$  is assumed constant over time. It ranges from 30m for tropical forests  
 771 and 20m for coniferous boreal forests to 15m, 10m or 5m for temperate forests and 2m for bushes.  
 772 For herbaceous plants,  $h_{veg} = LAI / 6$ , with LAI the leaf area index given by the ECOCLIMAP  
 773 database. It ranges approximately from 0.01m to 0.8m for grassland/tundra. Finally, the height of  
 774 crop types is related to an exponential function of LAI and has a height of 1m before maturity

775 defined as a LAI of  $3.5 \text{ m}^2 \cdot \text{m}^{-2}$ . More details on these physiographic parameters can be found in  
776 Masson et al. [2003].

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## TABLE CAPTIONS

1011

1012 **Table 1** – The peat soil hydraulic and thermal parameter values used in ISBA for fibric and sapric  
 1013 soil.  $w_{sat}$  ( $m^3.m^{-3}$ ) is the porosity,  $w_{fc}$  ( $m^3.m^{-3}$ ) the water content at field capacity specified as  
 1014 matric potential at -0.1 bar for peat soil,  $w_{wilt}$  ( $m^3.m^{-3}$ ) the water content at wilting point (matric  
 1015 potential of -15 bar),  $b$  the dimensionless shape parameter of the soil-water retention curve,  $\psi_{sat}$   
 1016 (m) the soil matric potential,  $k_{sat}$  ( $m.s^{-1}$ ) the soil hydraulic conductivity at saturation,  $c$  ( $J.m^{-3}.K^{-1}$ )  
 1017 the soil heat capacity of organic matter,  $\lambda_s$  ( $W.m^{-1}.K^{-1}$ ) the thermal conductivity of soil matrix, and  
 1018  $\lambda_{dry}$  ( $W.m^{-1}.K^{-1}$ ) the dry soil thermal conductivity. For pedotransfer functions of Boelter [1969], the  
 1019 fiber content in fibric soil is assumed to be equal to 76.8 % against 21.8 % in sapric soil in order to  
 1020 reach soil porosity values close to Letts et al. [2000]. The method for averaging mineral soil  
 1021 properties with peat soil values using the fraction of soil that is organic is also given for each  
 1022 parameter.

$\alpha_{peat}$	<i>Fibric soil</i>	<i>Sapric soil</i>	<i>Sources</i>	<i>Mineral/Peat average</i>
$w_{sat}$	0.930	0.845	Letts et al. [2000] and Boelter [1969]	Arithmetic
$w_{fc}$	0.369	0.719	PTF from Boelter [1969]	Arithmetic
$w_{wilt}$	0.073	0.222	PTF from Boelter [1969]	Arithmetic
$b$	2.7	12	Letts et al. [2000]	Arithmetic
$\psi_{sat}$	-0.0103	-0.0101	Letts et al. [2000]	Arithmetic
$k_{sat}$	$2.8 \cdot 10^{-4}$	$1.0 \cdot 10^{-7}$	Letts et al. [2000]	Geometric
$c$	$2.5 \cdot 10^{-6}$	$2.5 \cdot 10^{-6}$	Farouki [1986]	Arithmetic
$\lambda_s$	0.25	0.25	Farouki [1986]	Geometric
$\lambda_{dry}$	0.05	0.05	Farouki [1986]	Geometric

1023

1023 **Table 2** – Daily skill scores simulated by each experiment at Col de Porte for snow depth, SWE,  
1024 albedo and soil temperature at 10cm over the number of point measurement,  $n$ . The bias, centred  
1025 root mean square errors (c-rmse) and square correlation ( $r^2$ ) described in section 4.1 are shown.  
1026 The best scores are given in bold.

	<i>Period</i>	<i>Criterion</i>	<i>Experiments</i>			
			<i>CTL</i>	<i>SNL</i>	<i>CPT</i>	<i>NEW</i>
<i>Snow depth (m)</i>	<i>DJF</i> ( $n=1624$ )	<i>bias</i>	0.126	0.108	<b>0.074</b>	0.089
		<i>c-rmse</i>	0.159	0.157	<b>0.126</b>	0.130
		$r^2$	0.863	0.870	<b>0.907</b>	0.900
	<i>MAM</i> ( $n=1656$ )	<i>bias</i>	0.165	0.127	0.077	<b>0.027</b>
		<i>c-rmse</i>	0.223	0.192	0.169	<b>0.155</b>
		$r^2$	0.845	0.878	0.884	<b>0.900</b>
	<i>All</i> ( $n=4737$ )	<i>bias</i>	0.102	0.082	0.053	<b>0.041</b>
		<i>c-rmse</i>	0.176	0.157	0.130	<b>0.126</b>
		$r^2$	0.889	0.908	0.923	<b>0.927</b>
<i>SWE (kg.m<sup>-2</sup>)</i>	<i>DJF</i> ( $n=835$ )	<i>bias</i>	12.329	6.196	<b>4.934</b>	8.887
		<i>c-rmse</i>	38.331	35.004	<b>34.476</b>	36.079
		$r^2$	0.901	0.913	<b>0.915</b>	0.911
	<i>MAM</i> ( $n=887$ )	<i>bias</i>	25.022	19.064	16.352	<b>0.334</b>
		<i>c-rmse</i>	61.138	57.204	55.699	<b>49.583</b>
		$r^2$	0.861	0.872	0.876	<b>0.900</b>
	<i>All</i> ( $n=2310$ )	<i>bias</i>	13.851	9.169	7.648	<b>2.981</b>
		<i>c-rmse</i>	45.641	42.267	41.134	<b>38.100</b>
		$r^2$	0.902	0.910	0.913	<b>0.924</b>
<i>Albedo (-)</i>	<i>DJF</i> ( $n=1456$ )	<i>bias</i>	0.047	0.047	0.047	<b>0.045</b>
		<i>c-rmse</i>	0.076	0.076	0.076	<b>0.074</b>
		$r^2$	0.528	<b>0.535</b>	0.533	0.506
	<i>MAM</i> ( $n=1516$ )	<i>bias</i>	0.077	0.077	0.076	<b>0.023</b>
		<i>c-rmse</i>	0.119	0.117	0.115	<b>0.080</b>
		$r^2$	0.768	0.785	0.792	<b>0.889</b>
	<i>All</i> ( $n=4101$ )	<i>bias</i>	0.048	0.046	0.045	<b>0.026</b>
		<i>c-rmse</i>	0.101	0.098	0.098	<b>0.082</b>
		$r^2$	0.858	0.869	0.871	<b>0.905</b>
<i>Soil temperature 10cm (K)</i>	<i>DJF</i> ( $n=1323$ )	<i>bias</i>	-1.082	-1.009	<b>-0.962</b>	-1.032
		<i>c-rmse</i>	0.892	0.837	<b>0.797</b>	0.811
		$r^2$	0.234	0.234	0.272	<b>0.279</b>
	<i>MAM</i> ( $n=838$ )	<i>bias</i>	-0.646	-0.624	-0.606	<b>-0.199</b>
		<i>c-rmse</i>	2.109	1.995	1.967	<b>1.701</b>
		$r^2$	0.827	0.848	0.852	<b>0.896</b>
	<i>All</i> ( $n=2237$ )	<i>bias</i>	-1.121	-1.079	-1.049	<b>-0.936</b>
		<i>c-rmse</i>	1.650	1.591	1.569	<b>1.519</b>
		$r^2$	0.871	0.880	0.883	<b>0.894</b>

1027

1027

## FIGURE CAPTIONS

1028 **Figure 1** – Spatial distribution of the observed soil organic carbon content over two soil horizon  
1029 (0-30cm and 30-70cm) at 0.5° by 0.5° resolution. Observations come from the Harmonized World  
1030 Soil Database at 1 km resolution of the Food and Agricultural Organization.

1031 **Figure 2** – Parameterization of the effect of soil organic carbon (SOC) on soil hydraulic and  
1032 thermal properties. The soil organic carbon density profile,  $\rho_{soc}$ , is given by Equation 21 using a  
1033 top soil organic carbon content of 10 kg.m<sup>-2</sup>, a sub soil content of 15 kg.m<sup>-2</sup>, and via a simple  
1034 linear interpolation at each soil grid nodes that conserves the total soil carbon mass. The fraction of  
1035 the soil that is organic,  $f_{soc}$ , in each layer is determined assuming a simple relationship between this  
1036 last soil organic carbon density profile and an idealized peat soil density profile (Equation 22).  
1037 Examples for the soil porosity,  $w_{sat}$ , the soil saturated hydraulic conductivity,  $k_{sat}$ , and the soil heat  
1038 capacity,  $c$ , are given. Dotted lines represent vertical homogeneous mineral soil properties, dashed  
1039 lines the idealized peat soil properties, and plain lines the resulting combined soil properties using  
1040 averaging method sums-up in Table 1.

1041 **Figure 3** – Overview of the four experiments performed at the Col de Porte field site. Daily  
1042 simulated and observed data for snow depth (top) and SWE (bottom) are provided for 18 and 10  
1043 years respectively. In-situ observations are in black, the *CTL* simulation in blue, *SNL* in green,  
1044 *CPT* in orange, and *NEW* in red. The corresponding statistics are given in terms of annual bias and  
1045 c-rmse for each year by measurements periods.

1046 **Figure 4** – As Figure 3 but for surface albedo (top) and soil temperature at 10 cm depth (bottom).

1047 **Figure 5** – Daily mean annual cycles of snow depth, SWE, surface albedo, and soil temperature at  
1048 10 cm depth simulated (colours) and observed (black) at the Col de Porte field site. The  
1049 corresponding skill scores are given in Table 2. Over all panels, the grey shadow corresponds to  
1050 the uncertainty in in-situ measurements as discussed in section 4.1. The observed snow depth  
1051 exhibits an accuracy of  $\pm 1$ cm, the soil temperature is measured with a precision of  $\pm 1$ K, while  
1052 uncertainties in SWE and surface albedo is near  $\pm 10\%$ .



1053 **Figure 6** – Daily mean annual cycles of snow density ( $\text{kg.m}^{-3}$ ) and snowpack internal temperature  
1054 ( $^{\circ}\text{C}$ ) simulated by the four experiments over 18 years at the Col de Porte field site.

1055 **Figure 7** – Quantitative comparison between observed (plain circles) and simulated (plain fields)  
1056 daily snow depth and monthly soil temperature at 20cm and 160cm depths over the Northern-  
1057 Eurasia. Results from the bare soil sub-grid patch alone of the *NEW-SOC* simulation are presented  
1058 because in-situ measurements have been collected in open areas following the WMO standards as  
1059 mentioned in section 5.1.

1060 **Figure 8** – Mean annual cycles of observed and simulated daily snow depth and monthly soil  
1061 temperature profiles. The mean cycles are computed by averaging all simulated or observed mean  
1062 annual cycles at each station. However, total skill scores (bias and c-rmse) found in each panel are  
1063 computed merging together all simulated or observed time series of all stations over the entire  
1064 observed periods.

1065 **Figure 9** – Daily snow depth skill scores (bias and c-rmse) simulated by the *CTL* and the *NEW*  
1066 experiments during winter (DJF) and spring (MAM) over the Northern-Eurasia and expressed in  
1067 meters. Total scores given between parentheses are computed by merging together all simulated or  
1068 observed daily time series of all stations for each season.

1069 **Figure 10** – Monthly soil temperature skill scores at 20cm and 160cm depths simulated by the  
1070 *CTL* and the *NEW* experiments during winter and expressed in degrees Celsius. Total scores (bias  
1071 and c-rmse) are given for each panel.

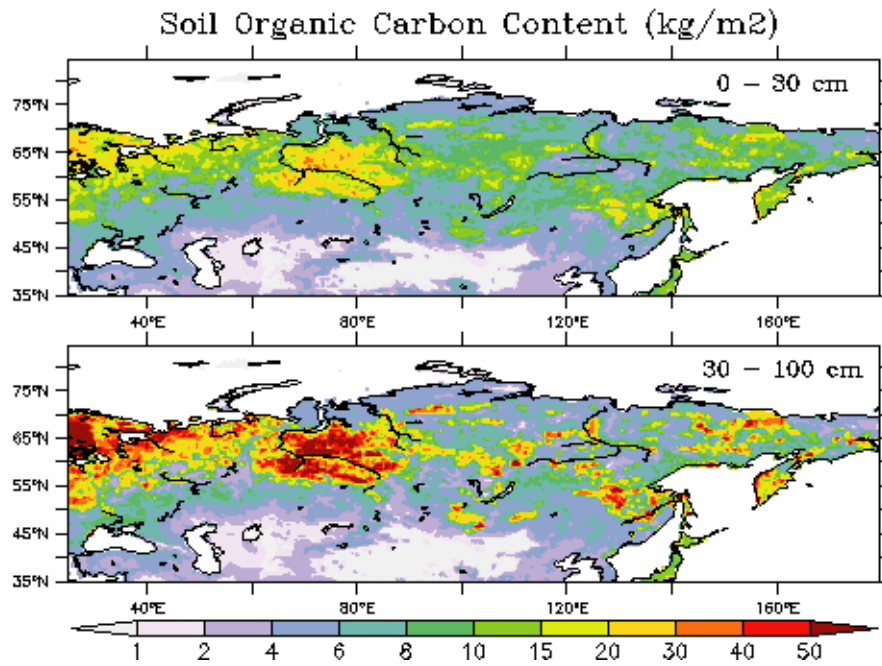
1072 **Figure 11** – Monthly soil temperature profile bias simulated by the *NEW* (left) and the *NEW-SOC*  
1073 (right) experiments during summer and expressed in degrees Celsius. Total skill scores (bias; c-  
1074 rmse) are given in the top-panel for each soil horizon.

1075 **Figure 12** – Distribution of permafrost characteristics. The NSIDC estimated limits of continuous,  
1076 discontinuous, sporadic and isolated permafrost regions are shown in the top panel. In each panel  
1077 the red lines correspond to the observed boundary of the entire permafrost region. In the middle  
1078 and the bottom panels, the mean active layer thicknesses simulated over the 1990-2013 period by

1079 the *NEW* and the *NEW-SOC* experiments are shown and compared to observations from the  
1080 CALM network (circles). Total skill scores are given for each experiment.

1081 **Figure 13** – Estimated and simulated active layer thicknesses over the Yakutia region. Estimations  
1082 before the 1990s are given by Beer et al. [2013] while the *NEW* and the *NEW-SOC* experiments  
1083 are averaged over the 1979-1990 period. The estimated and simulated latitudinal zonal averages  
1084 are shown over the last panel where Beer et al. [2013] estimations are in black, *NEW* in blue and  
1085 *NEW-SOC* in red. Dashed lines correspond to uncertainties in active layer thicknesses estimations  
1086 computed using standard deviations provided with the dataset.

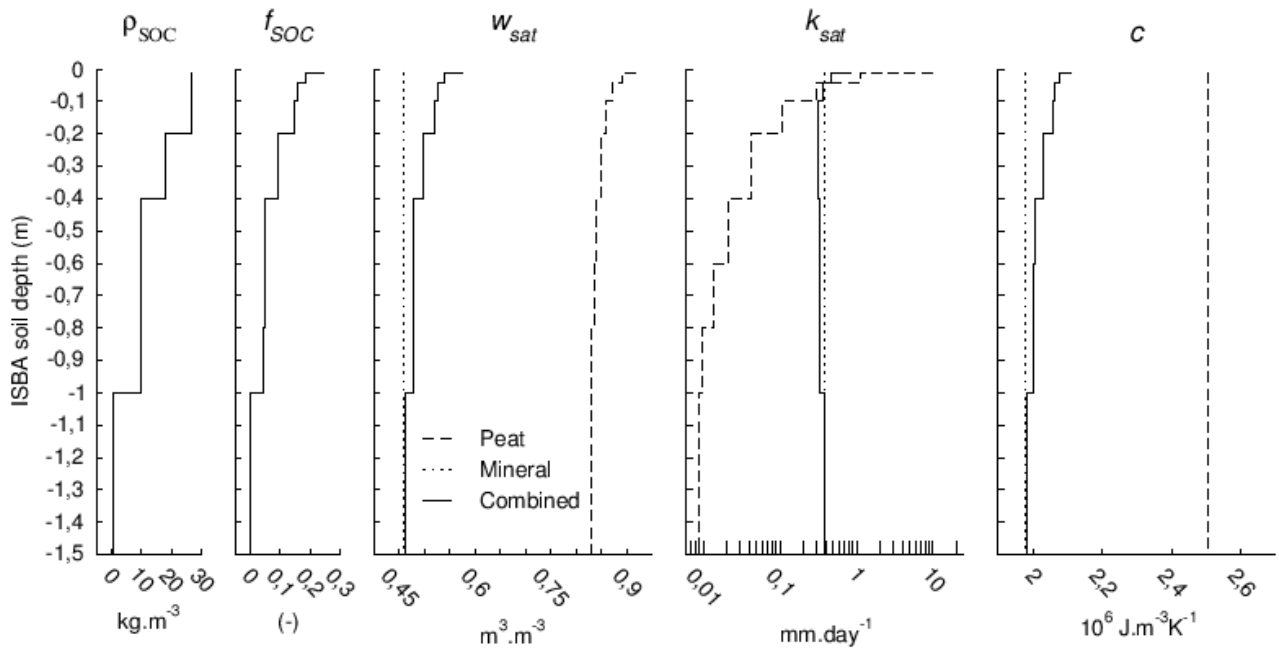
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1088 **Figure 1** – Spatial distribution of the observed soil organic carbon content over two soil horizon  
 1089 (0-30cm and 30-70cm) at 0.5° by 0.5° resolution. Observations come from the Harmonized World  
 1090 Soil Database at 1 km resolution of the Food and Agricultural Organization.

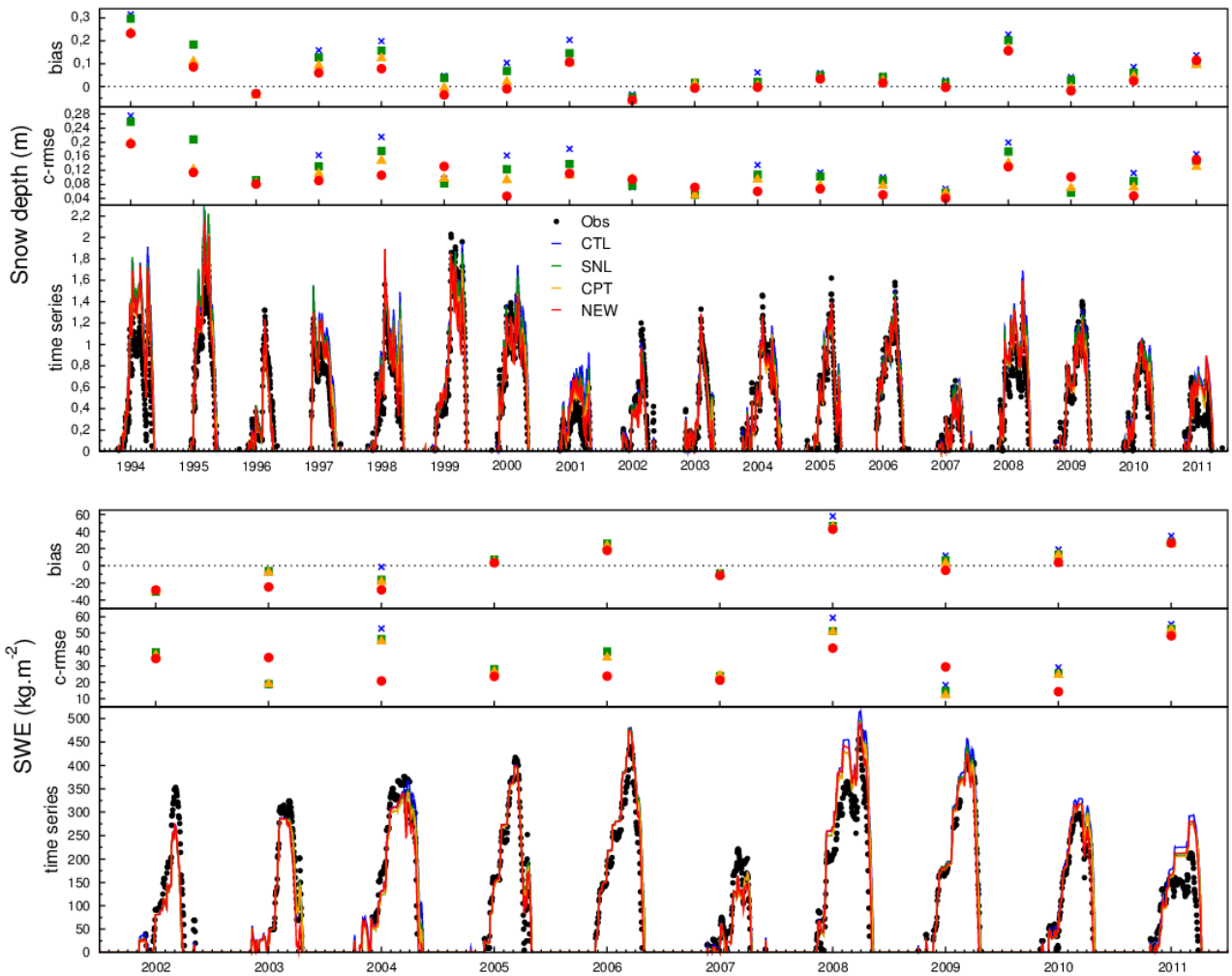
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1092 **Figure 2** – Parameterization of the effect of soil organic carbon (SOC) on soil hydraulic and  
 1093 thermal properties. The soil organic carbon density profile,  $\rho_{soc}$ , is given by Equation 21 using a  
 1094 top soil organic carbon content of  $10 \text{ kg.m}^{-2}$ , a sub soil content of  $15 \text{ kg.m}^{-2}$ , and via a simple  
 1095 linear interpolation at each soil grid nodes that conserves the total soil carbon mass. The fraction of  
 1096 the soil that is organic,  $f_{soc}$ , in each layer is determined assuming a simple relationship between this  
 1097 last soil organic carbon density profile and an idealized peat soil density profile (Equation 22).  
 1098 Examples for the soil porosity,  $w_{sat}$ , the soil saturated hydraulic conductivity,  $k_{sat}$ , and the soil heat  
 1099 capacity,  $c$ , are given. Dotted lines represent vertical homogeneous mineral soil properties, dashed  
 1100 lines the idealized peat soil properties, and plain lines the resulting combined soil properties using  
 1101 averaging method sums-up in Table 1.

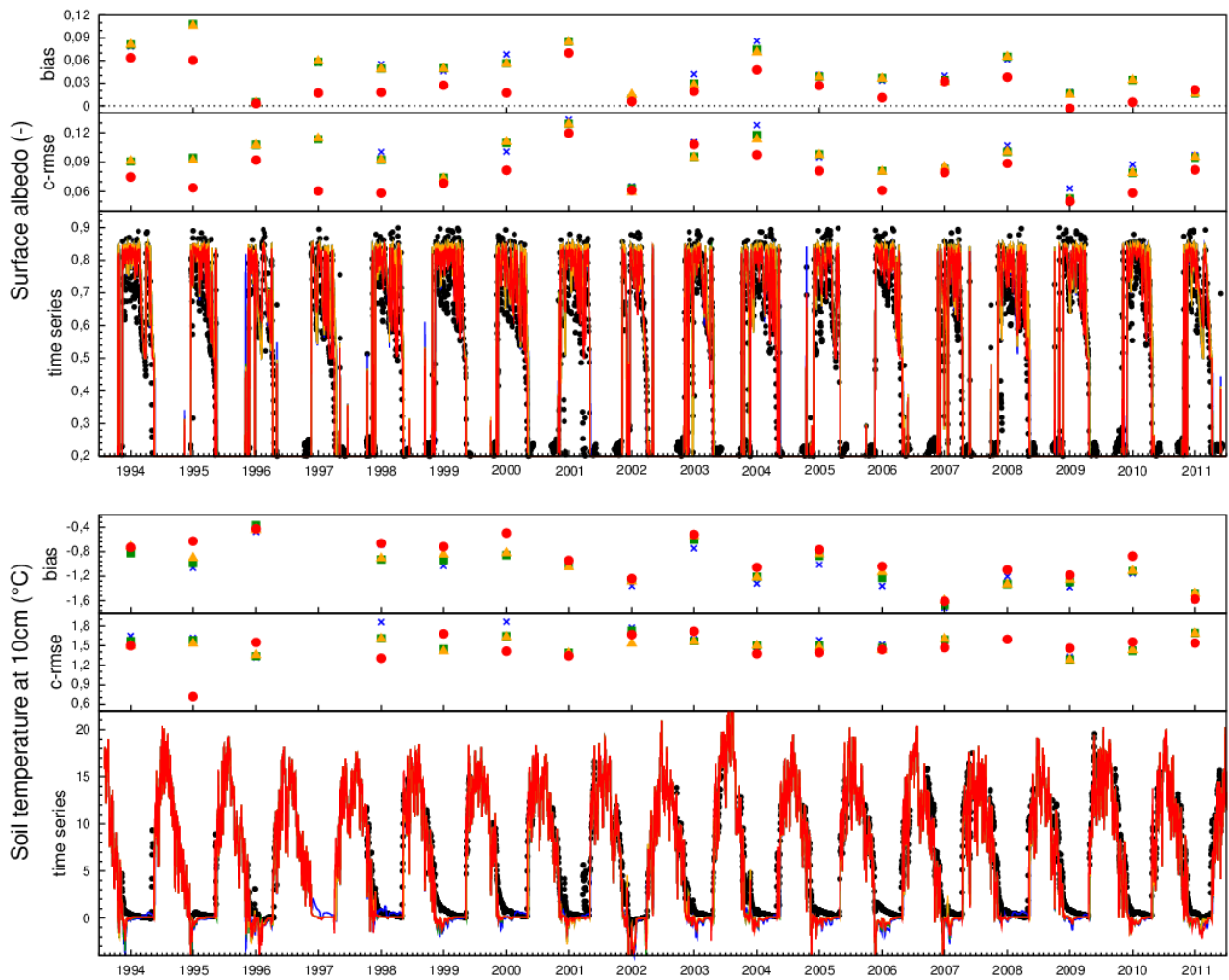
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1103 **Figure 3** – Overview of the four experiments performed at the Col de Porte field site. Daily  
 1104 simulated and observed data for snow depth (top) and SWE (bottom) are provided for 18 and 10  
 1105 years respectively. In-situ observations are in black, the *CTL* simulation in blue, *SNL* in green,  
 1106 *CPT* in orange, and *NEW* in red. The corresponding statistics are given in terms of annual bias and  
 1107 c-rmse for each year by measurements periods.

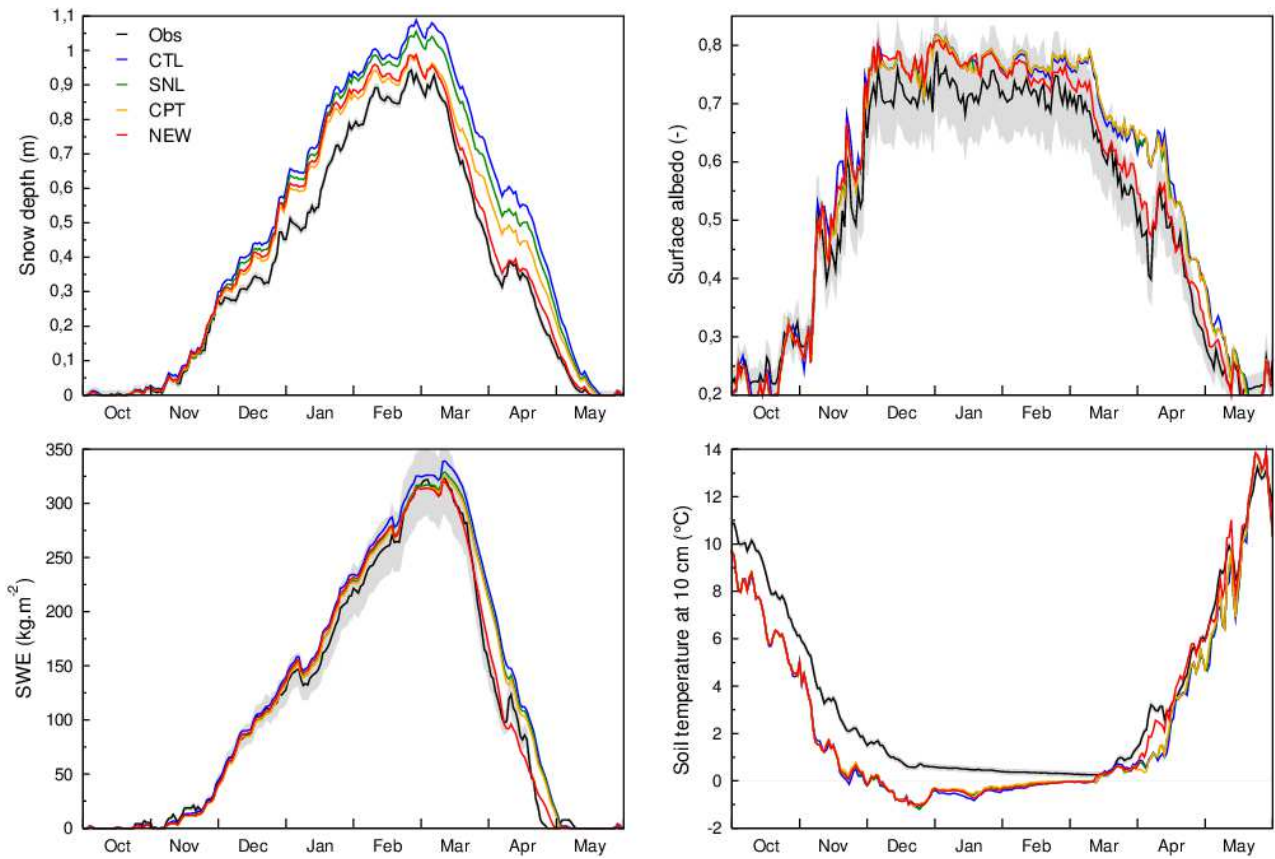
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1109 **Figure 4** – As Figure 3 but for surface albedo (top) and soil temperature at 10 cm depth (bottom).

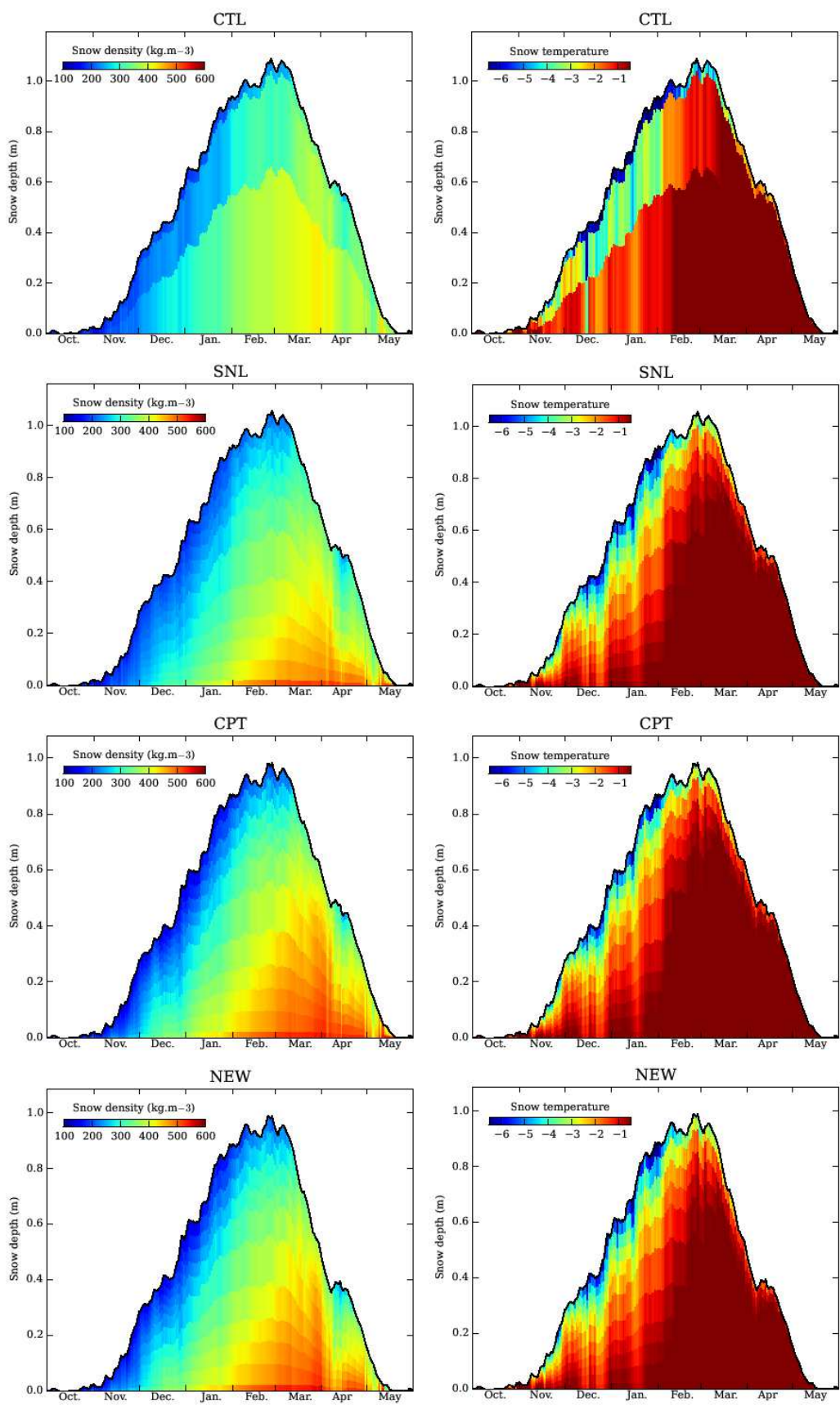
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1111 **Figure 5** – Daily mean annual cycles of snow depth, SWE, surface albedo, and soil temperature at  
 1112 10 cm depth simulated (colours) and observed (black) at the Col de Porte field site. The  
 1113 corresponding skill scores are given in Table 2. Over all panels, the grey shadow corresponds to  
 1114 the uncertainty in in-situ measurements as discussed in section 4.1. The observed snow depth  
 1115 exhibits an accuracy of  $\pm 1\text{cm}$ , the soil temperature is measured with a precision of  $\pm 1\text{K}$ , while  
 1116 uncertainties in SWE and surface albedo is near  $\pm 10\%$ .

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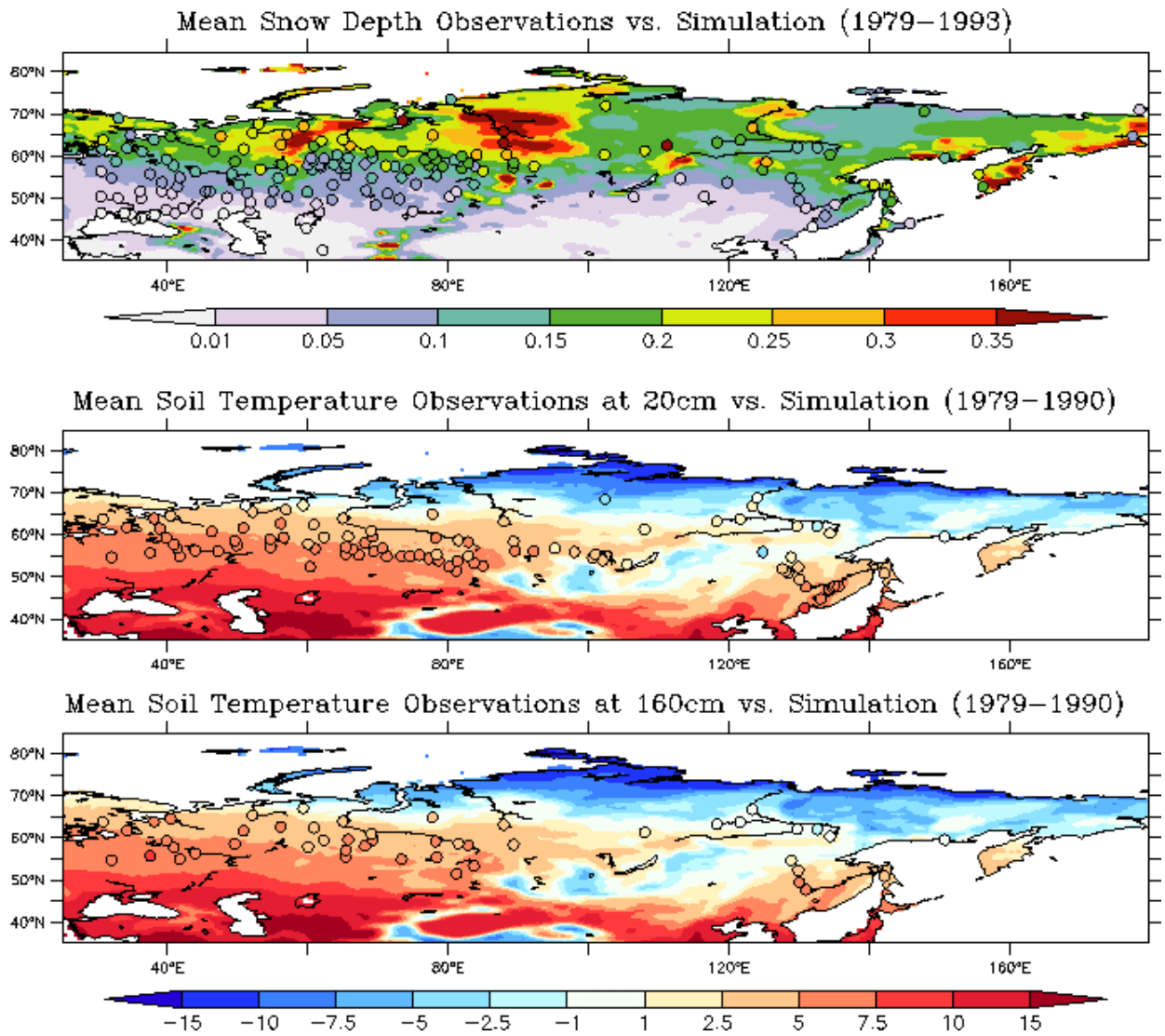


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1118 **Figure 6** – Daily mean annual cycles of snow density ( $\text{kg.m}^{-3}$ ) and snowpack internal temperature

1119 ( $^{\circ}\text{C}$ ) simulated by the four experiments over 18 years at the Col de Porte field site.

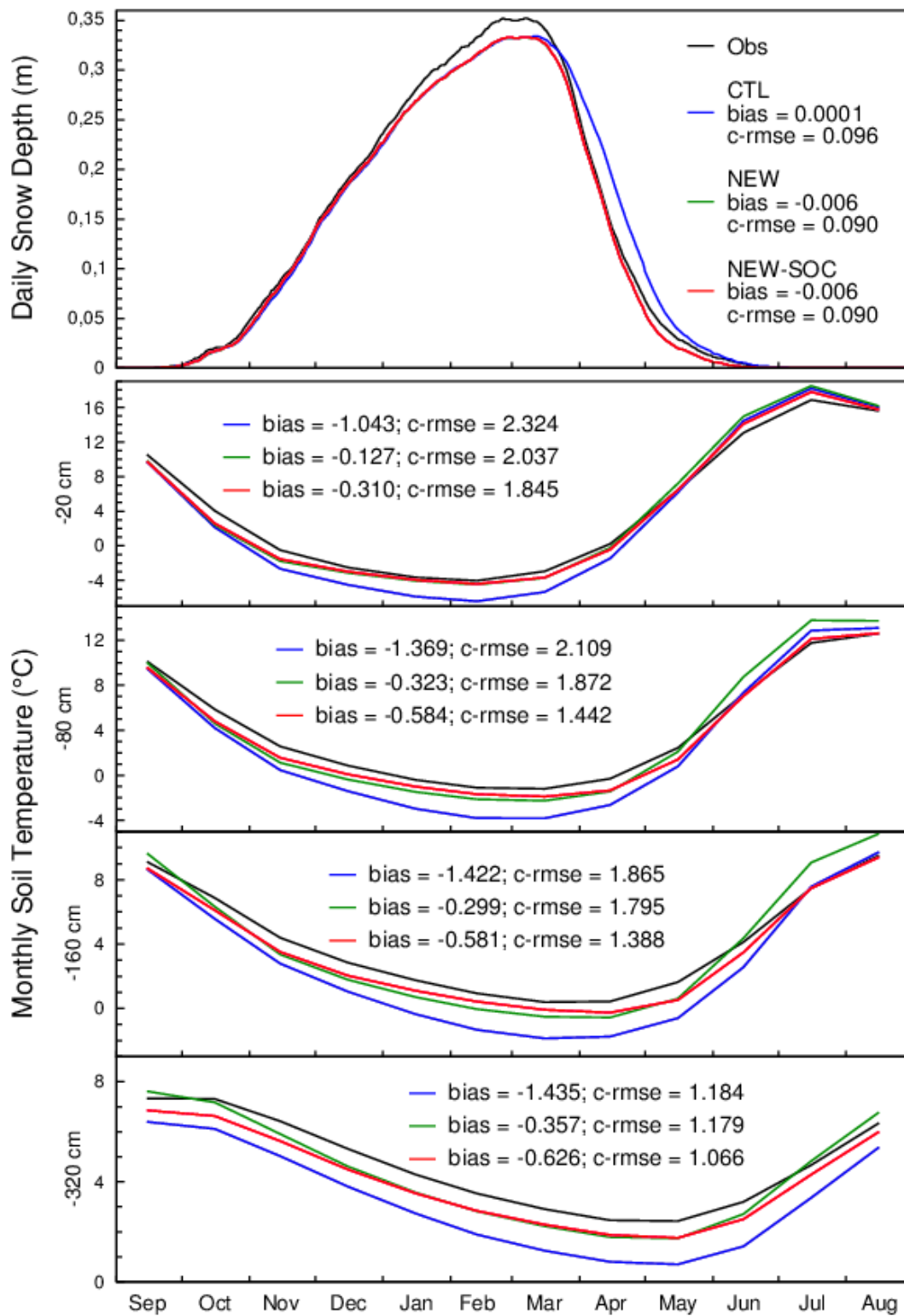




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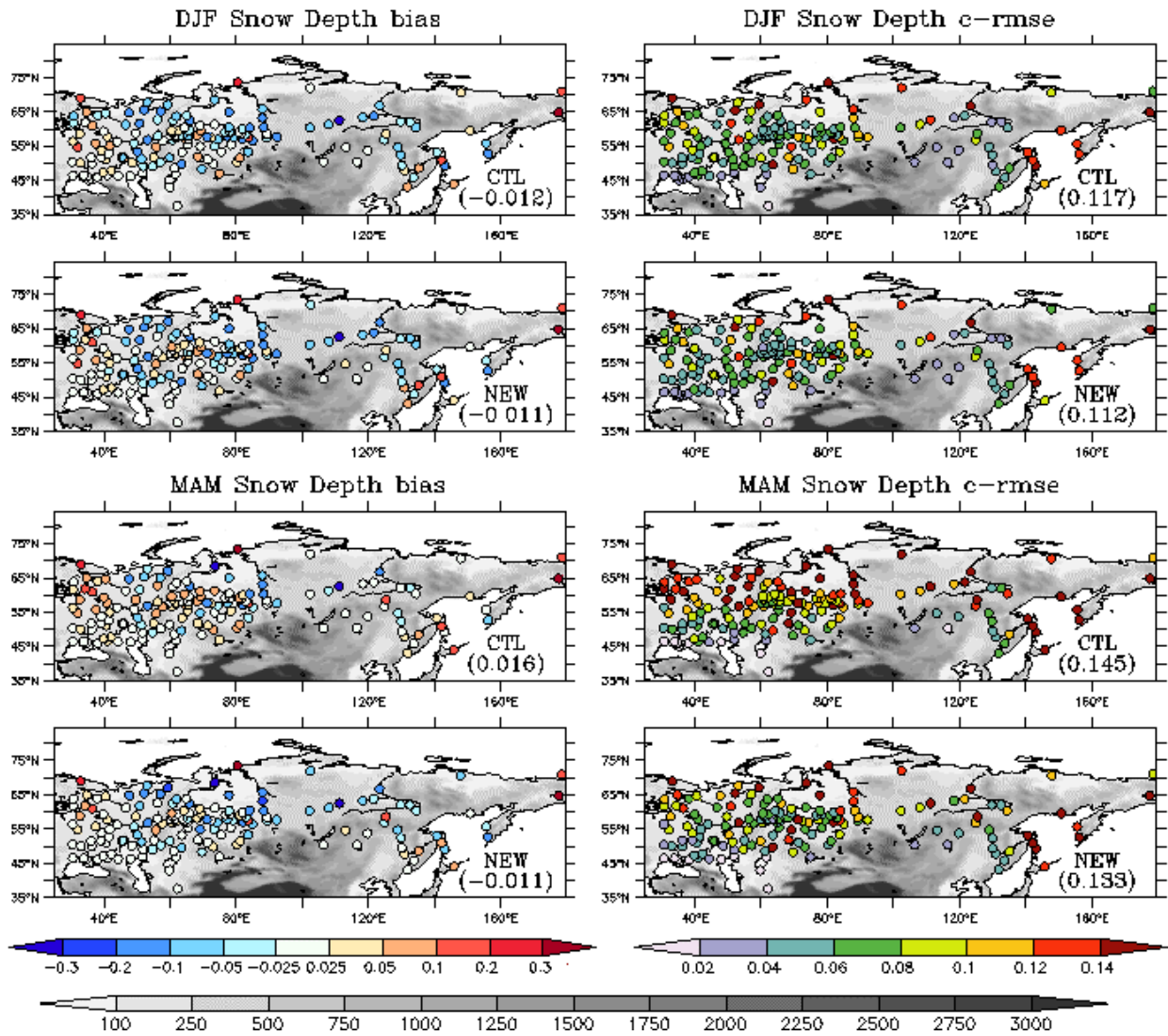
1121 **Figure 7** – Quantitative comparison between observed (plain circles) and simulated (plain fields)  
 1122 daily snow depth and monthly soil temperature at 20cm and 160cm depths over the Northern-  
 1123 Eurasia. Results from the bare soil sub-grid patch alone of the *NEW-SOC* simulation are presented  
 1124 because in-situ measurements have been collected in open areas following the WMO standards as  
 1125 mentioned in section 5.1.

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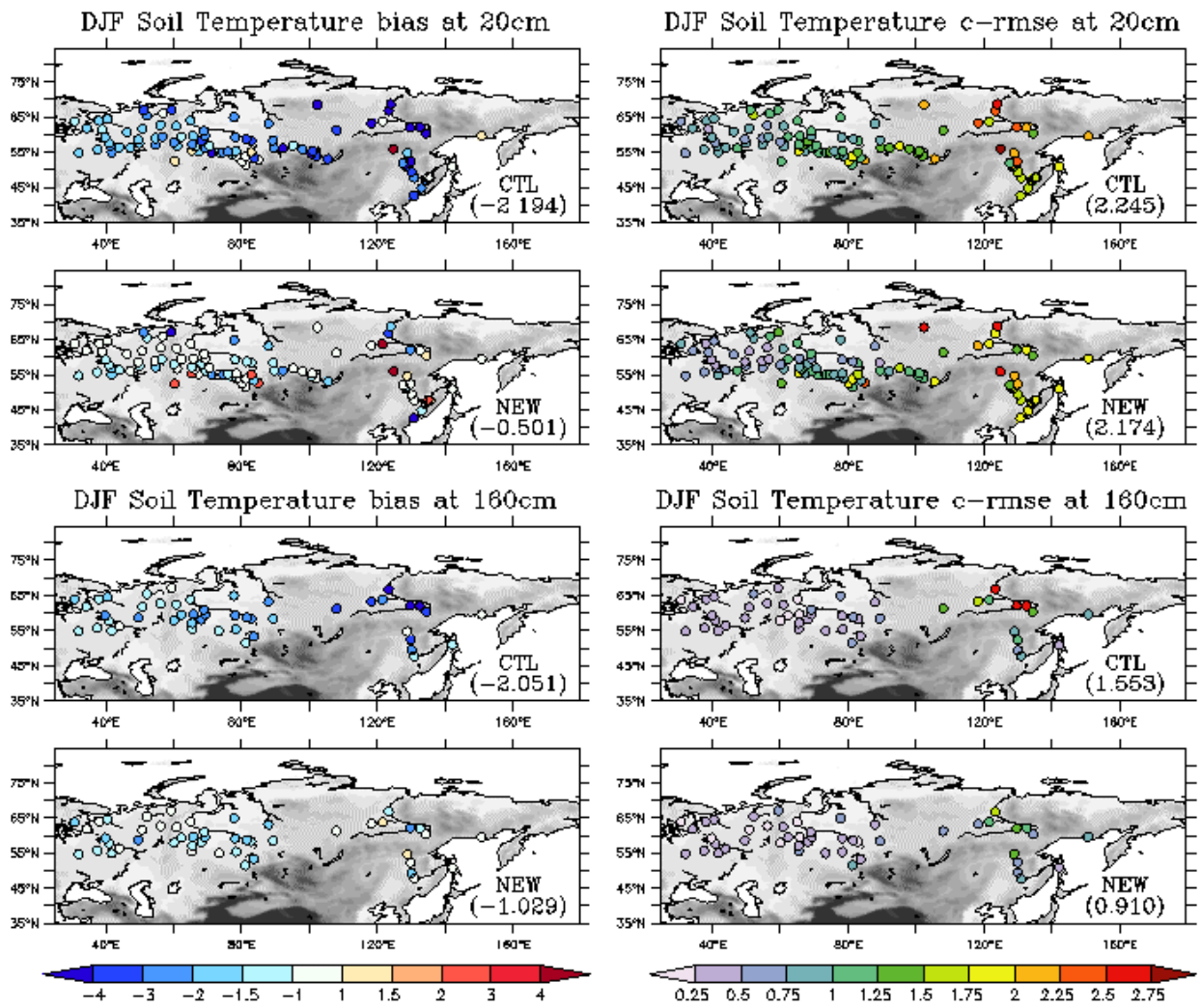
1127 **Figure 8** – Mean annual cycles of observed and simulated daily snow depth and monthly soil  
 1128 temperature profiles. The mean cycles are computed by averaging all simulated or observed mean  
 1129 annual cycles at each station. However, total skill scores (bias and c-rmse) found in each panel are  
 1130 computed merging together all simulated or observed time series of all stations over the entire  
 1131 observed periods.



1132

1133 **Figure 9** – Daily snow depth skill scores (bias and c-rmse) simulated by the *CTL* and the *NEW*  
 1134 experiments during winter (DJF) and spring (MAM) over the Northern-Eurasia and expressed in  
 1135 meters. Total scores given between parentheses are computed by merging together all simulated or  
 1136 observed daily time series of all stations for each season.

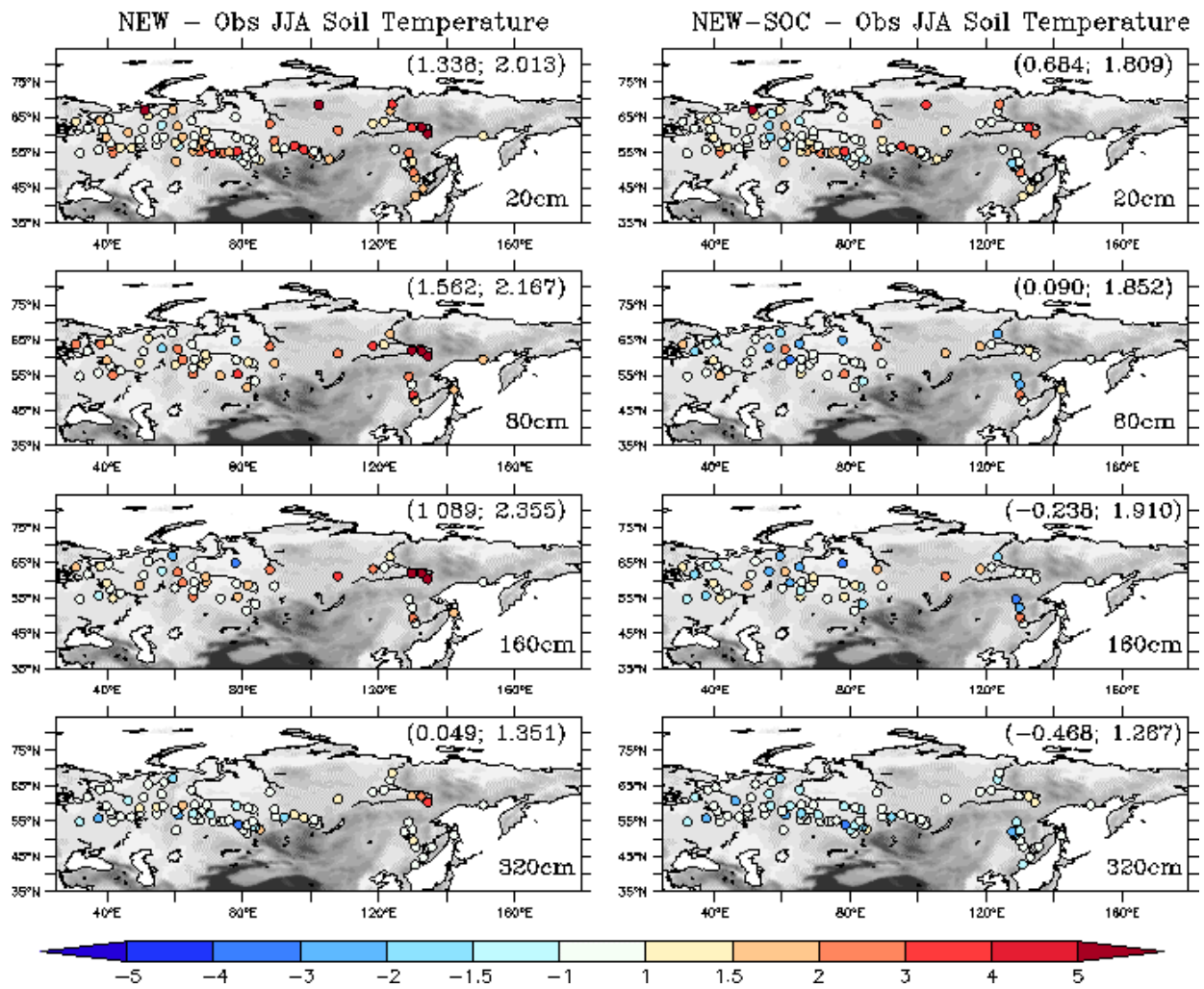
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1138 **Figure 10** – Monthly soil temperature skill scores at 20cm and 160cm depths simulated by the  
 1139 *CTL* and the *NEW* experiments during winter and expressed in degrees Celsius. Total scores (bias  
 1140 and c-rmse) are given for each panel.

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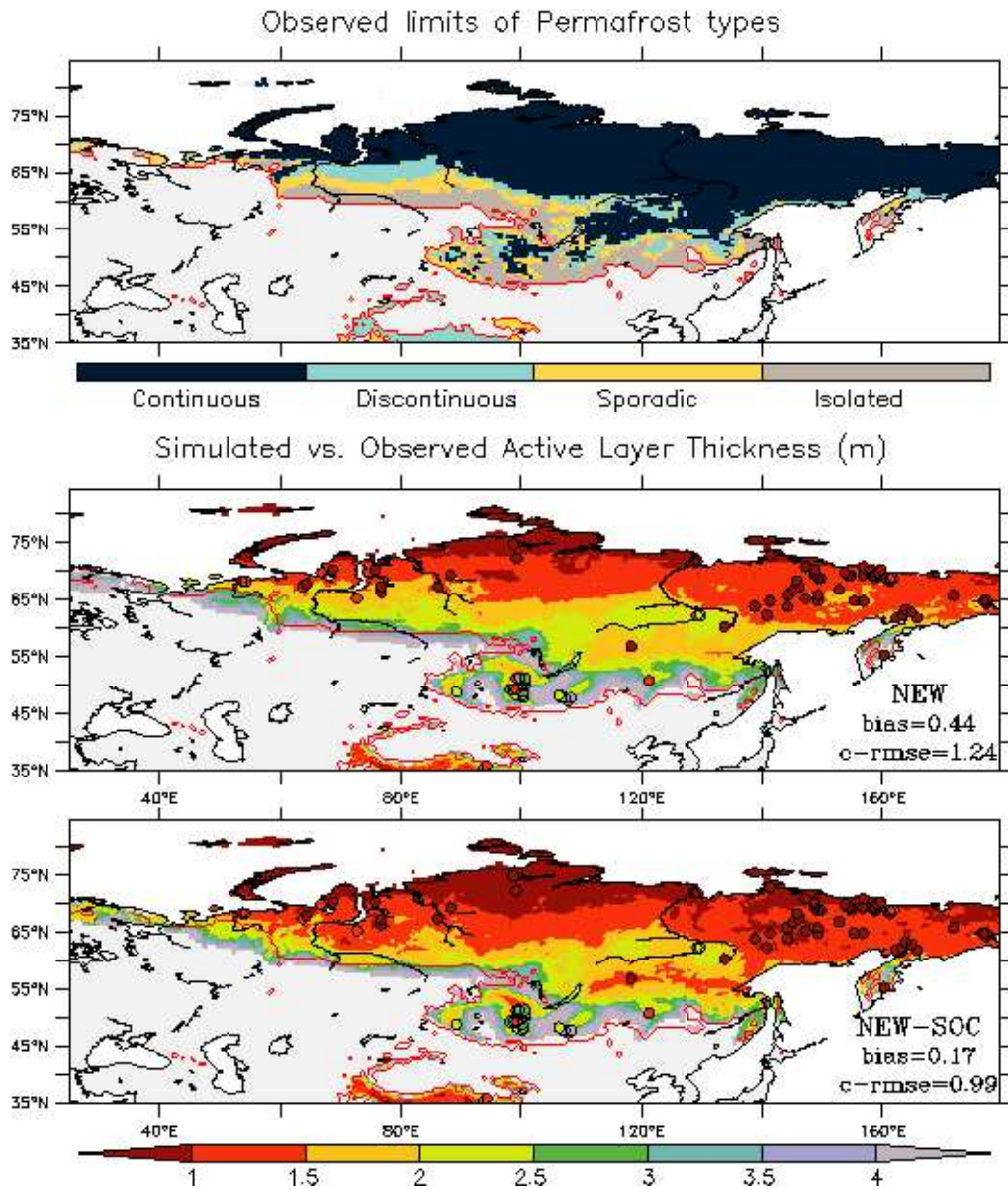


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1142 **Figure 11** – Monthly soil temperature profile bias simulated by the *NEW* (left) and the *NEW-SOC*  
 1143 (right) experiments during summer and expressed in degrees Celsius. Total skill scores (bias; c-  
 1144 rmse) are given in the top-panel for each soil horizon.

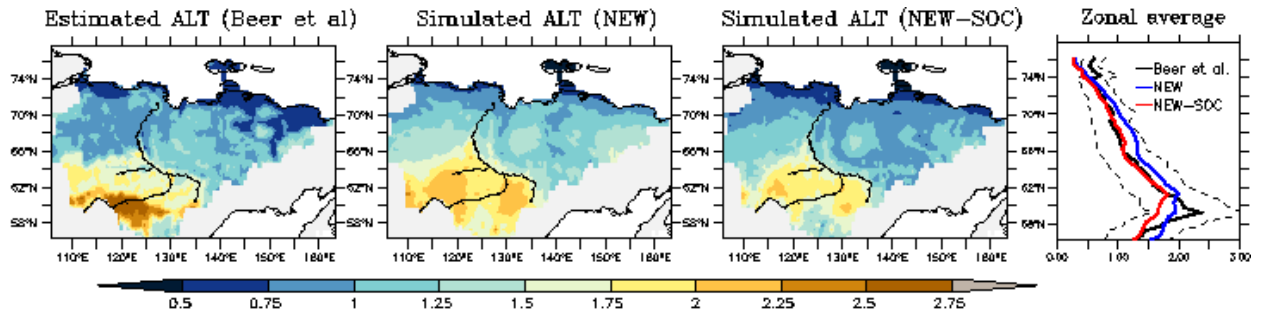
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1154 **Figure 13** – Estimated and simulated active layer thicknesses over the Yakutia region. Estimations  
 1155 before the 1990s are given by Beer et al. [2013] while the *NEW* and the *NEW-SOC* experiments  
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