

Specific comments:

P.6742, l. 12 and after: It is an error from the edition. In the original manuscript there is no problem, viscosity is in Pa.s.

Section 3.1.2: The current version of the manuscript indeed refers to snowdrift as a process responsible for snow compaction. "In the new version of the snow scheme, the evolution of snow density in each layer is due to snow compaction resulting from changes in snow viscosity (Brun et al., 1989) and snow densification of surface layers during snowdrift events (Brun et al., 1997)." The reviewer comment indicates that this formulation may be ambiguous. Indeed, what is referred to here is more precisely "wind-induced densification of near surface snow layers", which is related to the occurrence of snowdrift events but not blowing snow itself. In the revised manuscript, the term "snowdrift" is replaced everywhere by "wind-driven compaction" or "wind-induced densification of near surface snow layers ». This process is indeed responsible for the build-up of "wind crusts". Accordingly, some variable names were changed : ρ_{drift} to ρ_{wmax} , τ_{drift} to τ_w , and Γ_{drift} to Γ_w ,

P. 6754, l. 17 and after: "Global " scores was replaced by "Total" everywhere.

P. 6755, l. 6-9: The sentence "In terms of snow temperature, SNL induces warmer bottom snow layers in winter due to the densification of the lowest snow model layers and the insulation from other overlying layers. This explains the skill scores improvement found in winter soil temperature in Table 2" is indeed not clear. In fact, the soil temperature is warmer in winter with SNL (12 layers) compared to CTL (3 layers) because the bottom snow layer is better insulated from the atmosphere when using many layers due to the thermal gradients better represented. So this last bottom layer is warmer during winter in SNL (Figure 6). Because the temperature of the first layer of the soil remains close to the bottom snow temperature, the soil temperature is also warmer in SNL than in CTL. The previous sentence was changed to "In addition, the better representation of the vertical profile of density, that induces less dense and thus more insulating surface snow layers from November to February, permits also to better insulate the bottom snow layer from the atmosphere and results in higher bottom snow and top soil temperature. This explains the skill scores improvement found in winter soil temperature in Table 2"

P. 6755, l. 12:., We agree that SWE is not relevant here and it has been removed.

P. 6759, l. 1: The text is already too long and details about topographic indexes are not important for this study. Details are given in Decharme et al. (2013) Appendix B as it was said on page 6758 line 17. I prefer to suppress this sentence: " , and the topographic indexes used in the TOPMODEL runoff parameterization are given by the 1 km resolution HYDRO1K product (http://eros.usgs.gov/#/Find_Data/Products_and_Data_Available/gtopo30/hydro)" in the manuscript because descriptions of subgrid hydrological parameterizations are not the subject of this article and are given in Decharme et al. (2013).

P. 6759, l. 22 and after: Ablation or melting can be said for snow but I have changed all in "melting".

P. 6760, l. 15-17: We add "Figure 9" and a new paragraph starts after this sentence. The meaning of the sentence "In winter, regardless of the experiments, ISBA underestimates snow

depth measurements at many stations, especially in the Northern and Western parts of the domain” is that all simulations with ISBA underestimate snow depth measurements in the Northern and Western parts of Siberia.

P. 6760, ll. 22-23: “even if a slight cold bias appears at the subsurface” means that even if *NEW-SOC* is in better agreement with observations compared to *NEW* in Figure 11, it is not the case at the subsurface where the bias is negative (cold) especially at 320cm depth (this is not the case for *NEW*). The sentence was change to “ In summer (Figure 11), as was already shown in Figure 8, *NEW-SOC* is in better agreement with observations compared to *NEW* regardless of the soil horizon (lower c-rmse) even if a slight cold bias appears at the subsurface as shown by the negative total bias found at 320cm depth.

P. 6760, l. 28: Correct and the sentence is now : ”The best total scores found on Figures 8 and 11 without soil organic carbon by the *NEW* experiment...”

P. 6761, ll. 14-19: I agree with the entire comment of reviewer 1. So, Figure 12 was changed and the CALM data was added (see next page). However, as you can see, the areas categorized as continuous and discontinuous permafrost exhibit generally active layers shallower than 1m or 1.5m in the CALM data (see next page or new manuscript). For this propose, the simulations have been extended to 2013 because the CALM period of observation is 1990-2015. So we have added in section 5.1 the description of the CALM data : “The second dataset gives access to in-situ observations on active layer thickness collected by the Circumpolar Active Layer Monitoring (CALM; <http://www.gwu.edu/~calm/>) since the 1990s to 2015 [Brown et al. 2000]. Over the studied domain, 233 monitoring sites are available. To compare with simulations performed at a 0.5° by 0.5° resolution, 89 virtual stations have been computed from the 233 original sites by averaging all stations in each 0.5° by 0.5° grid-cells.” In section 5.3, results are now commented as follow: “However, this comparison with the limits of different permafrost types does not allow to determine which simulation leads to the most accurate active layer thicknesses. The comparison with the CALM data given in Figure 12 seems to show that *NEW-SOC* simulates a more accurate spatial distribution of the active layer thickness. This result is confirmed by Figure 13 that shows the estimated and simulated active layer thicknesses over the Yakutia region”. The caption of Figure 12 is changed to “Distribution of permafrost characteristics. The NSIDC estimated limits of continuous, discontinuous, sporadic and isolated permafrost regions are shown in the top panel. In each panel the red lines correspond to the observed boundary of the entire permafrost region. In the middle and the bottom panels, the mean active layer thicknesses simulated over the 1990-2013 period by the *NEW* and the *NEW-SOC* experiments are shown and compared to in-situ observations from the CALM network (circles). Total skill scores are given for each experiment.”

P. 6765, l. 7: “soil carbon degradation” means soil organic carbon decomposition. The end of the sentence was changed to “...due to the strong interaction between soil thermal processes and soil organic carbon decomposition with release of greenhouse gases.”

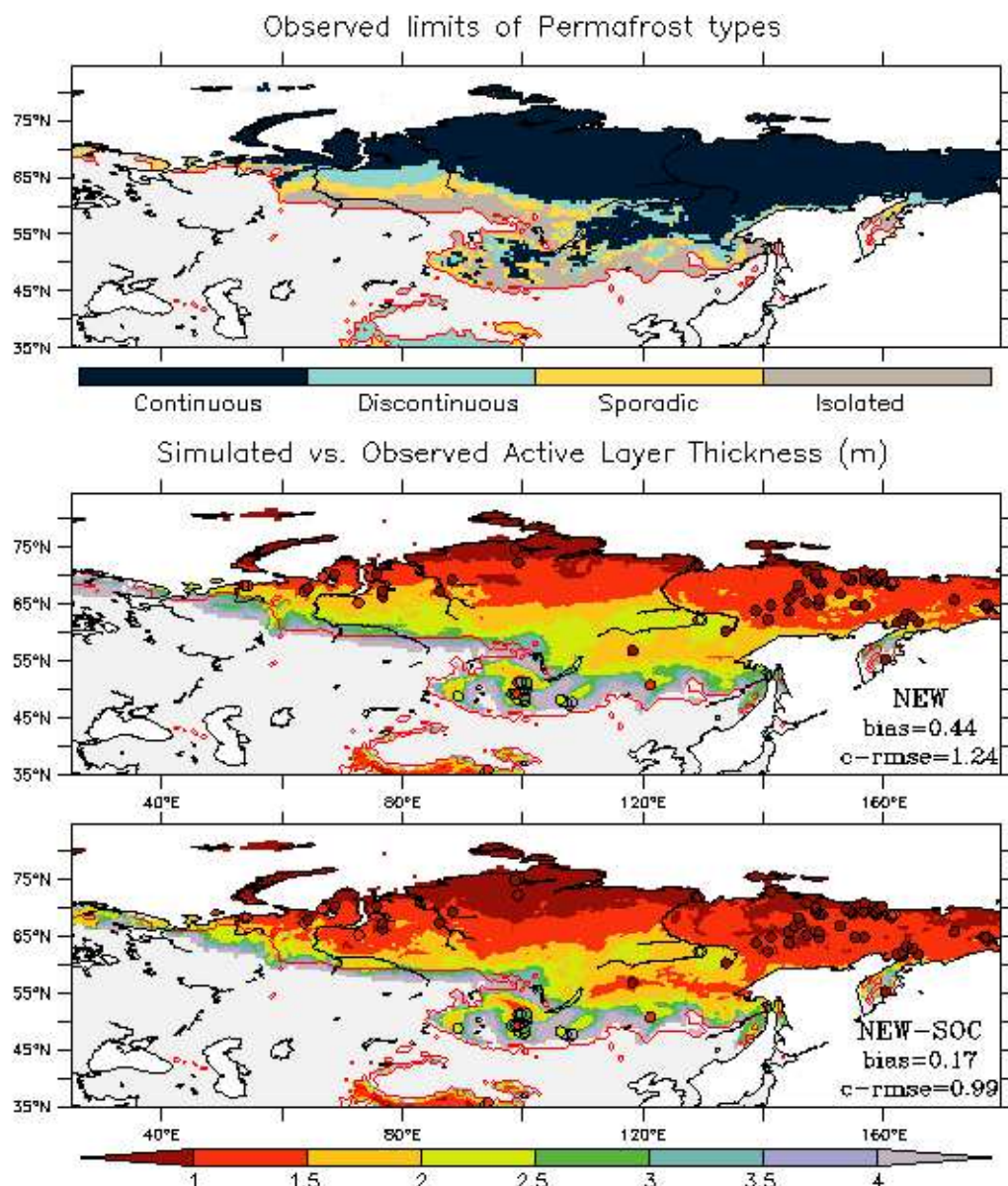
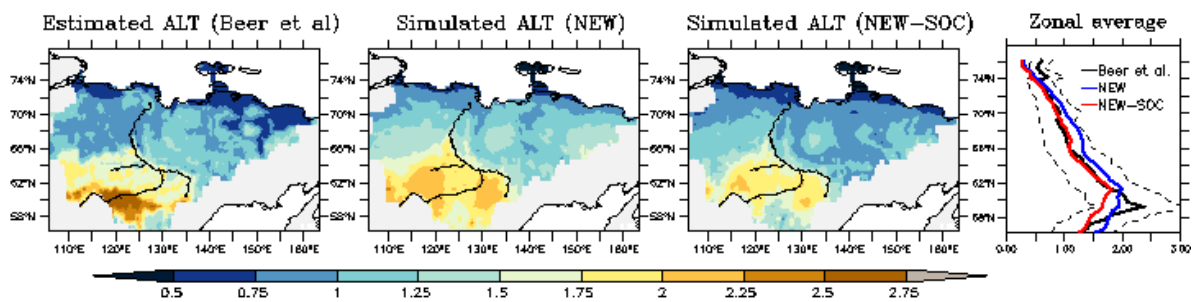


Figure 12 – Distribution of permafrost characteristics. The NSIDC estimated limits of continuous, discontinuous, sporadic and isolated permafrost regions are shown in the top panel. In each panel the red lines correspond to the observed boundary of the entire permafrost region. In the middle and the bottom panels, the mean active layer thicknesses simulated over the 1990-2013 period by the *NEW* and the *NEW-SOC* experiments are shown and compared to observations from the CALM network (circles). Total skill scores are given for each experiment.

Technical corrections: All corrections have been done

Other main author's changes in manuscript :

- (1) Reference added : Brown J., Hinkel K.M., & Nelson F.E. 2000. The Circumpolar Active Layer Monitoring (CALM) program: historical perspectives and initial results. *Polar Geography* 24(3): 165-258.
- (2) Wang et al. 2015 was changed to Wang et al. 2016 (new published article: Wang, W., Rinke, A., Moore, J. C., Cui, X., Ji, D., Li, Q., Zhang, N., Wang, C., Zhang, S., Lawrence, D. M., McGuire, A. D., Zhang, W., Delire, C., Koven, C., Saito, K., MacDougall, A., Burke, E., and Decharme, B.: Diagnostic and model dependent uncertainty of simulated Tibetan permafrost area, *The Cryosphere*, 10, 287-306, doi:10.5194/tc-10-287-2016, 2016)
- (3) The color legend was added in Figure 13 last panel :



**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

1
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.

19

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 CO₂ and CH₄ 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

97 generally 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, ψ (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 ψ_{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 ψ 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 Δz 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, ρ_{sn} ($\text{kg}\cdot\text{m}^{-3}$) in each layer, i , is the sum of snow compaction
 199 due to change in snow viscosity, η ([Pa.s](#)), and settling due to freshly fallen snow, ξ (s^{-1}),
 200 following 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 σ (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](#), $v_1 = 0.081 \text{ K}^{-1}$, $v_2 = 0.018 \text{ m}^3 \cdot \text{kg}^{-1}$, $s_0 = 2.8 \cdot 10^{-6} \text{ s}^{-1}$, $s_1 = 0.04 \text{ K}^{-1}$, $s_2 =$
 206 $0.046 \text{ m}^3 \cdot \text{kg}^{-1}$, and $\rho_d = 150 \text{ kg}\cdot\text{m}^{-3}$ are empirical parameters calibrated by Anderson [1976]. The
 207 minimum density of snow is constrained to be $50 \text{ kg}\cdot\text{m}^{-3}$. The snowfall density, ρ_{snow} ($\text{kg}\cdot\text{m}^{-3}$), is
 208 expressed as a function of wind speed, V_a ($\text{m}\cdot\text{s}^{-1}$), 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 $a_\rho = 109 \text{ kg}\cdot\text{m}^{-3}$, $b_\rho = 6 \text{ kg}\cdot\text{m}^{-3} \cdot \text{K}^{-1}$, and $c_\rho = 26 \text{ kg}\cdot\text{s}^{1/2} \cdot \text{m}^{-7/2}$.

212 2.2.3. Transmission of solar radiation and Snow albedo

213 The absorption of incident shortwave solar radiation, R_{sw} ($\text{W}\cdot\text{m}^{-2}$), 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} ($\text{W}\cdot\text{m}^{-2}$)
 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 α_{sn} is the dimensionless snow albedo, and β_{sn} (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 1m , $\delta_{10} = 0.5\text{m}$, $\delta_{11} = 0.1\text{m}$, 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 9-11 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 ~~and snowdrift events~~

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](#) ~~snow densification of surface layers during snowdrift~~
 269 ~~events~~ [Brun et al. 1997]. [This wind-driven compaction process](#) ~~Snowdrift~~ is assumed to occur
 270 when wind velocity exceeds a threshold value that depends on snow surface characteristics. This
 271 process is especially important for simulating the evolution of the snow density over polar regions.
 272 Brun et al. [1997] pointed out that this process is also critical for reproducing the snow thermal
 273 conductivity and the snow temperature profile over these regions. Therefore, the time tendency of
 274 snow density in each layer is computed as follows:

$$275 \quad \frac{\partial \rho_{sn}(i)}{\partial t} = \rho_{sn}(i) \frac{\sigma(i)}{\eta(i)} + \max\left(0, \frac{\rho_{wmax} - \rho_{sn}(i)}{\tau_w(i)}\right) \quad (13)$$

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

$$281 \quad \left| \begin{aligned} \sigma(1) &= \frac{g \Delta z(1) \rho_{sn}(1)}{2} \\ \sigma(i) &= g \sum_{j=1}^{i-1} [\Delta z(j) \rho_{sn}(j)] \quad \forall i > 1 \end{aligned} \right. \quad (14)$$

282 The snow viscosity is a function of snow density, temperature, and liquid water content, W_l
 283 (kg.m⁻²), and it is given as follows:

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

285 where η_0 ([Pa·s](#)) is a reference viscosity equal to 7622370 [Pa·s](#), ρ_0 ($\text{kg}\cdot\text{m}^{-3}$) is a reference
286 density equal to $250\text{kg}\cdot\text{m}^{-3}$, W_{lmax} ($\text{kg}\cdot\text{m}^{-2}$) represents the maximum liquid water holding capacity
287 (e.g. section 2.2) and the constants $a_\eta = 0.1\text{K}^{-1}$, $b_\eta = 0.023 \text{ m}^3\cdot\text{kg}^{-1}$, and $\Delta T_\eta = 5\text{K}$. The viscosity
288 dependence on snow temperature is limited according to Schlee et al. [2014] who pointed out that
289 the impact of snow temperature on snow densification becomes negligible at low temperatures.
290 The last dimensionless function, f_w , describes the decrease of viscosity in presence of liquid water.
291 Finally, the snowfall density is computed as previously (Equation 7).

292 3.1.3. Transmission of solar radiation and Snow albedo

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

$$298 \quad Q_s(i) = R_{sw} \sum_{k=1}^3 \left[\omega(k) (1 - \alpha_{sn}(k)) \exp\left(-\sum_{j=1}^i [\beta_{sn}(k, j) \Delta z(j)]\right) \right] \quad (16)$$

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

306 [Warren 1984]. In each band, both the albedo and the extinction coefficient of snow are computed
 307 according to Brun et al. [1992] as follows:

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

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

$$313 \quad d_{opt}(i) = \min \left[d_{max}, g_1 + g_2 \times \rho_{sn}(i)^4 + g_3 \times \min(15, A_{sn}(i)) \right] \tag{18}$$

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

317 The snow age for each layer is the time, in days, since the snow has fallen. When a
 318 snowfall event occurs, the fresh snow characteristics including its age (0 at time of snowfall) are
 319 averaged out with the snow already present in the first layer according to their respective masses.
 320 Finally, when the layer thicknesses of the entire snowpack are recalculated with Equation 11 and
 321 12, the snow age is redistributed accordingly. For example, the age of snow in the first layers can
 322 remain from 0 day to a week during winter but aging largely in spring, while the last layers age
 323 continuously.

324 3.2. Effects of soil organic carbon on soil hydraulic and thermal properties

325 North-Eurasian soils are rich in organic carbon as shown in Figure 1. This figure represents
 326 the soil organic carbon content of two soil horizons (0-30cm and 30-70cm) aggregated at a 0.5° by

327 0.5° horizontal resolution and estimated from the Harmonized World Soil Database (HWSD;
328 <http://webarchive.iiasa.ac.at/Research/LUC/External-World-soil-database/HTML/>) at a 1 km
329 resolution from the Food and Agricultural Organization [FAO 2012]. The parameterization of the
330 impact of soil organic carbon on hydraulic and thermal properties in ISBA is based on
331 pedotransfer functions of Boelter [1969], and on the work by Letts et al. [2000] and Lawrence and
332 Slater [2008]. The pedotransfer functions of Boelter [1969] link the soil water retention at different
333 pressure levels to the fiber content of a peat soil. Letts et al. [2000] describe the vertical profile of
334 hydraulic properties such as soil matric potential and hydraulic conductivity at saturation for a
335 typical organic soil. The hydraulic properties change sharply from the near surface where peat is
336 weakly decomposed (fibric soil) to the sub-surface with moderately and well decomposed peat
337 (hemic and sapric soils respectively). Lawrence and Slater [2008] proposed a linear combination
338 of such soil organic properties with the standard mineral soil properties.

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

$$342 \quad \alpha_{peat}(i) = \alpha_{fibric} z(i)^\beta \quad \text{with} \quad \beta = \frac{\ln(\alpha_{sapric} / \alpha_{fibric})}{\ln(d_{sapric} / d_{fibric})} \quad (19)$$

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

347 To determine the organic fraction of soil, the density profile of the soil carbon must be
348 known for the entire soil grid. Using the HWSD database, the soil carbon densities in the first
349 0.3m, ρ_{top} (kg.m⁻³), and the remaining 0.7m below, ρ_{sub} (kg.m⁻³), are known:

$$350 \quad \rho_{top} = \frac{S_{top}}{\Delta d_{top}} \quad \text{and} \quad \rho_{sub} = \frac{S_{sub}}{\Delta d_{sub}} \quad (20)$$

351 where S_{top} and S_{sub} (kg.m^{-2}) are the topsoil and subsoil organic carbon contents respectively, Δd_{top}
352 and Δd_{sub} (m) the thicknesses of each observed soil horizon (0.3 and 0.7m respectively). We
353 extrapolate the density present below 1m from this observed near-surface profile (Equation 20).
354 The extrapolation assumes that the carbon profile decreases sharply with soil depth according to a
355 power function. The shape of this function is given by the observed profile if the topsoil organic
356 carbon density is superior to the subsoil density. Otherwise, the density of soil carbon below a 1m
357 depth, ρ_{deep} (kg.m^{-3}), is taken equal to the subsoil density:

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

359 where Δd_{deep} (m) is an infinite soil thickness taken arbitrarily equal to 1000m.

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

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

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

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

372 4.1. Experimental data set

373 The Col de Porte field site (45°17'N, 05°45'E) is located at an elevation of 1325m in the
374 French Alps near Grenoble [Morin et al. 2012]. It consists in a 50m by 50m square covered by
375 grass, mowed approximately once a month in summer depending on its growth rate. Soil textures
376 (30% clay, 60% sand) are characteristic of a sandy-clay-loam soil that is very poor in organic
377 carbon. For this reason, this site is only used to evaluate the effect of changes in snow
378 parameterizations while changes in soil physics can be not tested. The atmospheric forcing
379 variables (air temperature, rain and snow rates, air humidity, atmospheric pressure, wind speed,
380 long-wave and short-wave incident radiation) are available at a one hour time step from August 1st,
381 1993 to July 31, 2011. It consists of a combination of in-situ measurements, roughly from
382 September to June each year, and the regional reanalysis SAFRAN from June to September each
383 year (see Morin et al. [2012] for details).

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

399 4.2. *Model configuration*

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

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

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

423 4.3. *Results*

424 Figure 3 and 4 show an overview of the four simulations performed at the Col de Porte in
425 terms of snow depth, SWE, surface albedo and soil temperature at 10cm. A quick look at the time
426 series indicates that all of the model versions match the observations relatively well. However,
427 annual statistics show a clear hierarchy between the four experiments. The snow depth statistics
428 shows that the new snow compaction/densification algorithm has a positive impact on the
429 snowpack simulation. Indeed, both the *CPT* and *NEW* experiments exhibit the lowest bias and c-
430 rmse for twelve of the eighteen years. However, the comparison to SWE data does not allow a
431 discrimination between the four simulations, even if the c-rmse of the *NEW* experiment is the best
432 for seven of the ten years. The surface albedo from the *NEW* simulation is clearly better than the
433 albedo from the other experiments: bias and c-rmse are the best for all years (Figure 4). The soil
434 temperature bias and c-rmse are also reduced by the *NEW* experiment (for ten of seventeen years)
435 compared to the other simulations. Thus, accounting for different spectral bands within the snow
436 albedo calculation has a significant positive impact on the energy balance of the snow-soil system.

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

450 variables during this period. They induce a sharp springtime snowmelt with a strong decrease in
451 snow depth and SWE. The snow insulation during spring is thus less important and allows the soil
452 surface to warm up faster. As a result, the model is capable of reproducing the strong soil warming
453 observed in April (Figure 5). Not surprisingly, the soil temperature skill scores for spring and the
454 whole period are drastically improved although there is a slight degradation in winter.

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

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

474 5. Simulations over North-Eurasia

475 5.1. Numerical experiment design and observational dataset

476 The experimental design used here is close to that proposed by Brun et al. [2013]. The
477 region considered (35°N to 85°N, 25°E to 180°E) covers Eastern-Europe, Russia and Siberia
478 (Figure 7). The ISBA land surface model is run at a 0.5° by 0.5° spatial resolution using the
479 Interim Re-Analysis (ERA-I; <http://www.ecmwf.int/en/research/climate-reanalysis/era-interim>)
480 [Dee et al. 2011]. ERA-I meteorological variables are extracted with a 3-hourly frequency in order
481 to represent the diurnal cycle. This reanalysis covers the time period from 1979 to the present.
482 Many details about ERA-I can be found in Dee et al [2011] and an evaluation of its performance is
483 provided in Berrisford et al. [2011]. For precipitation, the monthly ERA-I precipitation are
484 rescaled to match the observed Global Precipitation Climatology Center (GPCC) Full Data
485 Product V5 (<http://gpcc.dwd.de>) as proposed by Decharme and Douville [2006a]. This method
486 conserves the 3-hourly chronology of the ERA-I precipitation but ensures a reasonable monthly
487 amount [Szczypta et al. 2012]. Brun et al. [2013] pointed out the significantly better performance
488 of this ERA-I scaled GPCC forcing product in simulating North-Eurasian snowpack variables
489 compared to the ERA-I precipitation or other “state of the art” global scale atmospheric forcings.

490 To evaluate snow and soil temperature simulations, several in-situ dataset are used. As in
491 Brun et al. [2013], the Historical Soviet Daily Snow Depth (HSDSD; http://nsidc.org/data/docs/noaa/g01092_hsd/d/index.html) compiled by Armstrong [2001] was used
492 in the current study. It consists in daily snow depth measurements taken at synoptic stations
493 following the World Meteorological Organization (WMO) standards. WMO requires the
494 measurements to be taken in bare ground open areas or clearings with regular grass cutting. These
495 snow depth data are therefore representative of open areas of bare ground or those covered with
496 very short grass. This dataset starts in 1881 with a few stations and ends in 1995. Considering that
497 ERA-I starts in 1979, the model simulations are done from 1979 ~~to 1993~~ according to Brun et al.
498 [2013]. 263 HSDSD stations are available over this period with approximately half of them
499 without any missing data. We chose to use only the stations where the difference between the local
500 and the ERA-I elevation is less than 100m to avoid temperature biases for instance that would be

502 directly due to the low resolution of ERA-I. We also only kept the stations where the number of
503 days with a non zero snow depth measurement over the entire period is superior to 100 days, and
504 that have at least 8 days with snow measurement per year. With this filter, the number of available
505 stations decreases to 158, which remains acceptable. Most stations are located in Russia and
506 Western-Siberia with only a few in Eastern-Siberia (Figure 7).

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

519 To quantify the capability of the model to simulate the permafrost characteristics, ~~threetwo~~
520 datasets are used. The first dataset is the Circum-Artic Map of Permafrost and Ground Ice
521 Conditions (<http://nsidc.org/data/ggd318>) edited by Brown et al. [2002]. This dataset is available at
522 a 0.5° by 0.5° resolution and shows the continuous, discontinuous, isolated and sporadic
523 permafrost boundaries. [The second dataset gives access to in-situ observations on active layer](#)
524 [thickness collected by the Circumpolar Active Layer Monitoring \(CALM;](#)
525 <http://www.gwu.edu/~calm/>) since the 1990s to 2015 [Brown et al. 2000]. Over the studied
526 domain, 233 monitoring sites are available. To compare with simulations performed at a 0.5° by
527 0.5° resolution, 89 virtual stations have been computed from the 233 original sites by averaging all

528 [stations in each 0.5° by 0.5° grid-cells.](#) The ~~second-last~~ dataset is an estimate of the active layer
529 thickness over North-West-Siberia [before the 1990s.](#) This data set is based on the map of
530 landscapes and permafrost conditions in Yakutia
531 (<http://doi.pangaea.de/10.1594/PANGAEA.808240>). It gives access to the mean and standard
532 deviation of the most probable active layer thickness in each grid box at 0.5° by 0.5° resolution.
533 All details can be found [in](#) Beer et al. [2013].

534 5.2. *Model configuration*

535 Three experiments using the ISBA land surface model forced by the ERA-I scaled GPCC
536 atmospheric dataset are performed using the same configuration. In addition to the *CTL* (old snow
537 scheme) and *NEW* (new snow scheme) experiments already described in section 4, we performed
538 one simulation using the parameterization of the impact of the soil organic carbon on the
539 hydrologic and thermal soil properties. This last experiment, called *NEW-SOC*, uses the new snow
540 and soil-property schemes described in section 3.1 and 3.2, respectively. As previously, the model
541 determines the temperature, liquid water and ice content evolution in each of the 14 soil layers
542 corresponding to a total soil depth of 12m. The model is run with a 15-minute time step from
543 January 1st, 1979 to Decembre 31, ~~1993~~[2013](#). The model's spin-up uses twenty iterations of the
544 first five years (1979 to 1983) of the atmospheric forcing, representing a total of one hundred
545 years.

546 In ISBA, we use a series of twelve sub-grid independent patches per grid cell in order to
547 account for land cover heterogeneity. Land cover parameters such as Leaf Area Index (LAI),
548 vegetation height, vegetation/soil albedos, and rooting depth are prescribed for each sub-grid
549 patch. The dominant patches present in the model over the Northern-Eurasian region are bare soil,
550 grassland/tundra, deciduous forest, coniferous boreal forest, and C3 crops in the South. The
551 fraction of each surface type within each grid box is used to compute the grid box average of the
552 water and energy budgets. Some other processes, such as surface runoff, dripping from the canopy

553 reservoir, and soil infiltration account for sub-grid parameterizations. More details can be found in
554 Decharme and Douville [2006b] and Decharme et al. [2013].

555 For all of the simulations, the grid-cell fraction covered by snow evolves according to the
556 simulated snow depth and is different for bare soil and vegetated areas (Appendix C) in each land
557 cover patch. As was the case for the Col de Porte experiment, the effective roughness length of
558 snow retains its usual value of 0.001m. The land surface parameters used by ISBA are specified
559 according to the 1-km resolution ECOCLIMAP-II database [Faroux et al., 2013]. LAI, vegetation
560 height, and vegetation/soil albedos are prescribed for the twelve vegetation sub-grid patches based
561 on a mean annual cycle at a 10-day time step. The rooting depth is specified for each vegetation
562 type according to Canadell et al. [1996]. It ranges from 0.5m to 1.5m for tundra and temperate
563 grassland, and from 2m to 3m for forest. The soil textural properties are given by the ~~HSWD~~
564 ~~HSWD~~ database at 1 km resolution ~~while. The~~ topographic information is specified according to
565 the 30-arcsecond resolution GTOPO30 data set, ~~and the topographic indexes used in the~~
566 ~~TOPMODEL runoff parameterization are given by the 1 km resolution HYDRO1K product~~
567 (http://eros.usgs.gov/#/Find_Data/Products_and_Data_Available/gtopo30/hydro).

568 5.3. Results

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

579 The seasonal cycles of daily snow depths and monthly soil temperatures (Figure 8) clearly
580 show the biases of the *CTL* simulation and the improvements due to the new snow and soil
581 representations. The seasonal cycles and the [globaltotal](#) skill scores are computed using the
582 measurements and simulations for all stations over the entire observed periods. ISBA globally
583 underestimates the snow depth from December through February with no clear difference between
584 *CTL* and *NEW* (or *NEW-SOC*). However, the springtime snow [ablationmelting](#) is drastically
585 improved by the new snow scheme inducing a better simulated seasonality. This fact is confirmed
586 by some other quantitative comparisons. The average number of days per year with observed snow
587 on the ground for all in-situ stations is 150.7 days. *CTL* simulates 158.7 days against 151.5 days
588 for *NEW*. On average, the last day of the snow season is day number 281.6 when starting on July
589 first. *CTL* goes beyond this date by more than 9 days while for *NEW* it is only 2 days (day number
590 283). These results are consistent with the model evaluation at the Col de Porte field site (section
591 4). As could be expected also, the new physical soil properties (*NEW-SOC*) play a minimal role in
592 the snow depth simulation. The seasonal cycle of the soil temperature profile confirms that the
593 new snow scheme induces a warmer soil in winter compared to *CTL*, and it strongly reduces the
594 cold bias of *CTL*. The effect of soil organic carbon is especially observable during spring and
595 summer. *NEW* exhibits a warm bias for each soil horizon while *NEW-SOC*, with more insulating
596 soils, reduces this weakness.

597 These improvement in snow depth and soil temperature are confirmed by the spatial
598 distributions of their seasonal skill scores (bias and c-rmse). Figure 9 shows the spatial
599 distributions of snow depth seasonal skill scores (bias and c-rmse) during winter and spring. No
600 clear [discrepancies differences](#) among these simulations appear in winter while the bias and c-rmse
601 of many stations are improved in spring by the new snow scheme. The springtime snow depth is
602 simulated in an acceptable manner by *NEW*, while *CTL* exhibits a significant overestimation. This
603 fact is confirmed by [globaltotal](#) scores given in each of the panels. In winter, regardless of the

604 experiments, ISBA underestimates snow depth measurements at many stations, especially in the
605 Northern and Western parts of the domains [\(Figure 9\)](#).

606 The spatial distribution of soil temperature seasonal skill scores simulated at 20 cm and 160
607 cm depth during winter is given in Figure 10. Regardless of the region, the generalized cold bias
608 found over all stations with *CTL* is drastically reduced with the new snow scheme and the
609 interannual variability (c-rmse) is largely improved. In summer [\(Figure 11\)](#), as was already shown
610 in Figure 8, *NEW-SOC* is in better agreement with observations [compared to *NEW*](#) regardless of
611 the soil horizon [\(lower c-rmse\)](#) even if a slight cold bias appears at the subsurface. ~~[\(as shown by](#)~~
612 ~~[the negative total bias found Figure 11 at 320cm depth\)](#)~~. The *NEW* experiment overestimates the
613 temperature profile measurements at many stations near the surface, but less-so at a 320 cm depth.
614 So, it seems that the subsurface cooling in the *NEW-SOC* experiment is too intensive. But in fact at
615 320 cm depth, the simulated soil temperature in the western part of the domain remains quasi
616 unchanged between *NEW-SOC* and *NEW*. The best ~~global~~[total](#) scores ~~(or seasonal cycles in Figure~~
617 ~~9)~~[found on Figures 8 and 11](#) without soil organic carbon by the *NEW* experiment are in fact due to
618 error compensation between the cold and warm biases simulated in the western and eastern part of
619 the domain, respectively.

620 The effect of soil organic carbon content on soil temperature profile is also especially
621 observable in terms of the simulated permafrost characteristics. The observed and simulated
622 locations of permafrost boundaries are compared in Figure 12. Regardless of the experiment, ISBA
623 generally simulates acceptable boundaries even if the permafrost limit extends slightly too far
624 south in the western part of the domain. This figure also shows the spatial distribution of active
625 layer thicknesses simulated by the *NEW* and the *NEW-SOC* experiments. The active layer
626 thickness in the model is computed as the maximum depth reached each year by the 0°C isotherm
627 in the soil approximated via a linear [interpolation between the last positive temperature node going](#)
628 [down from the surface and the first negative temperature node](#)~~[extrapolation between the last node](#)~~
629 ~~[from the surface where the soil temperature is positive and the first where it is negative.](#)~~ As

630 expected from the lower summer soil temperatures with *NEW-SOC* (Figure 9 and 11), the active
631 layer is shallower. However, this comparison with the limits of different permafrost types does not
632 allow ~~the determination of~~ to determine which simulation leads to the most accurate active layer
633 thicknesses ~~_, unless it is considered that the continuous and discontinuous permafrost exhibit~~
634 ~~generally active layers shallower than 1.5m or 2m. With this very subjective criterion~~ The
635 comparison with the CALM data given in Figure 12 seems to show that *NEW-SOC* ~~seems to~~
636 simulates a more accurate spatial distribution of the active layer thickness. ~~To partially verify this~~
637 ~~hypothesis,~~ This result is confirmed by Figure 13 that shows the estimated and simulated active
638 layer thicknesses over the Yakutia region. Estimations from Beer et al. [2013] present a strong
639 latitudinal gradient with an increase in active layer thickness from the north to the south. Both
640 experiments exhibit such profiles. However, the active layer thickness simulated by *NEW-SOC* is
641 in better agreement with these estimations than those by *NEW*. The latitudinal zonal average
642 confirms this result.

643 **6. Discussion and Conclusion**

644 In this study, the impact of improved representation of snowpack and soil properties in the
645 ISBA LSM to simulate snow characteristics and soil temperature profiles over cold regions was
646 analysed. ISBA's representations of snow layering, albedo, and compaction were updated by
647 incorporating some parameterizations of the detailed Crocus snowpack model. In addition, a
648 simple parameterization of the soil organic carbon effect on hydraulic and thermal soil properties
649 was introduced based on previous work [Boelter 1969; Letts et al. 2000; Lawrence and Slater
650 2008]. The model is evaluated first over the Col de Porte field site in the French Alps [Morin et al.
651 2012] in order to isolate the changes in the snowpack parameterization, and second over the North-
652 Eurasian region to analyze the model's ability to simulate snow depth, soil temperature profile and
653 permafrost characteristics.

654 Changes in the snowpack parameterizations induce noticeable improvements in the
655 simulated snow depth, SWE, surface albedo and soil temperature at the Col de Porte (field) site.

656 The new snow layering algorithm with 12 layers permits a refinement of the vertical distribution of
657 density and temperature in the snowpack leading to slight improvements in simulated snow depth,
658 SWE, and soil temperature during winter. The densification of the snowpack with the new
659 compaction scheme, which increases the density contrast between the top and the bottom snow
660 layers, has a significant positive impact on snow depth and winter soil temperature. Finally, the
661 new spectral albedo scheme clearly improves the simulation of the springtime surface albedo that
662 allows a better simulation of the snowpack characteristics and soil temperature during
663 [ablationmelting](#) at the end of the snow season.

664 It must be noted that the large improvement in snow albedo in spring is mainly due to the
665 use of snow age in the diagnostics of the optical diameter of snow (Equation 186). Without this
666 parameterization, the surface albedo is strongly overestimated in winter and, to a lesser extent in
667 spring at the Col de Porte field site, with a larger bias and c-rmse for all variables compared to the
668 new version of ISBA (not shown). The optical diameter of snow strongly controls the near-infrared
669 albedo, while impurities mostly affect the albedo in the visible spectrum [Wiscombe and Warren
670 1981]. This increase of snow optical diameter with time is necessary to represent well the decrease
671 in spectrally integrated albedo with age. However, the increase of snow optical diameter is not
672 only a function of snow density as parameterized by Anderson [1976] in Equation (810), but it is
673 also due to snow metamorphism, which is macroscopically driven by snow temperature and snow
674 thermal gradients. Several complex parameterizations exist to explicitly represent the evolution of
675 snow optical diameter according to these processes [e.g. Carmagnola et al. 2014]. Nevertheless,
676 for the sake of simplicity, we just use a snow age dependency in the diagnostic of snow optical
677 diameter with a limitation at fifteen days (Equation 186). This simple diagnostic allows the model
678 to reasonably match the explicit computation of the optical diameter of snow simulated in the
679 Crocus model (not shown). The good results of the ISBA model at the Col de Porte field site
680 reinforce this choice.

681 The positive impacts of the new ISBA snow scheme are confirmed when tested over the
682 North-Eurasian region with an important number of open field in-situ snow depth and soil
683 temperature stations. Winter snow depths are slightly better simulated with the new version and
684 the winter soil temperature cold bias obtained with the old version of ISBA is clearly reduced. This
685 fact confirms that the physics used in snow schemes is of primary importance for adequately
686 simulating the snow insulating effect that prevents soil from getting too cold in winter [Slater et al.
687 2001; Luo et al. 2003; Gouttevin et al. 2012; Paquin and Sushama 2015]. Another important
688 impact of changes in the ISBA snow scheme over the North-Eurasian region is seen in spring
689 when the snowmelt is well reproduced. As shown over the Col de Porte (field) site, this is mainly
690 due to the new parameterization of spectral snow albedo.

691 Nevertheless, regardless of the model version used, simulated winter snow depths are
692 generally underestimated compared to in situ observations. The cause of this underestimation is
693 not trivial. The first source of uncertainty can be attributable to the GPCC precipitation
694 measurements that do not account for wind undercatch leading to a possible underestimation of
695 solid precipitation during winter [Adam and Lettenmaier 2003, Brun et al., 2013]. Besides
696 uncertainties related to the atmospheric forcing, the snow depth underestimation can be due to the
697 non-explicit representation of snow metamorphism. Indeed, in similar experimental conditions
698 over the Northern Eurasian region, the winter snow depth simulated by the detailed Crocus
699 snowpack model did not exhibit the same problem [Brun et al. 2013] and the main remaining
700 difference between Crocus and ISBA is now restricted almost entirely to the explicit simulation of
701 snow metamorphism. In Crocus, the viscosity of layers composed of faceted crystals and depth
702 hoar snow types is increased [Vionnet et al., 2012], which leads to reducing the overall
703 compaction rate of snowpack undergoing temperature conditions conducive to such snow types,
704 and this is consistent with the situation described above.

705 Taking into account soil organic carbon in soil physical properties logically plays a
706 minimal role in the simulated snowpack behaviour. However, this process has drastic impacts on

707 the summer soil temperature profile because it allows the soil to remain cool during spring and
708 summer as shown in previous studies [Bonan and Shugart 1989; Lawrence and Slater 2008;
709 Dankers et al. 2011]. Consequently, the spatial distribution of the permafrost active layer thickness
710 simulated by the new version of ISBA is in better agreement with estimations from Beer et al.
711 [2013] over the Yakutia region. This result is in agreement with Paquin and Sushama [2015] who
712 showed that the hydraulic and thermal properties of soil organic carbon partly control the thickness
713 of the active layer during summer. However, spatial observations of permafrost characteristics on
714 the global scale are still very scarce, and if available, they are static and don't allow the study of
715 long term trends and inter-annual variability.

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

- 719 • An adequate simulation of snow layering and snow compaction/densification is important
720 in order to represent well winter snowpack characteristics and the soil temperature profile.
- 721 • Snow albedo strongly controls the simulation of the springtime snow characteristics and the
722 ~~ablation~~melting timing.
- 723 • To account for soil organic carbon in terms of the soil physical properties drastically
724 impacts the simulation of summer soil temperature profile and hence the permafrost active layer
725 thickness and its spatial distribution.

726 Finally, these conclusions underscore the fact that the representation of snowpack characteristics
727 and soil thermal processes are of primary importance for studying permafrost vulnerability under
728 climate change conditions, especially if the continental carbon cycle is considered due to the
729 strong interaction between soil thermal processes and soil organic carbon degradation
730 decomposition with release of greenhouse gases. ~~and soil thermal processes.~~

731

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731

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740

APPENDIX A

Snow thermal conductivity

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

$$746 \quad \lambda_{sn}(i) = \lambda_{ice} \left(\frac{\rho_{sn}(i)}{\rho_w} \right)^{1.88} + \frac{P_0}{P_a} \times \max \left(0, k_1 - \frac{k_2}{T_{sn}(i) - k_3} \right) \quad (\text{A1})$$

747 where λ_{ice} ($\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$) is the thermal conductivity of ice equal to $2.2 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$, ρ_w ($\text{kg}\cdot\text{m}^{-3}$) the
748 water density, P_a (Pa) the air pressure, P_0 (Pa) a reference pressure equal to 1000hPa , and the
749 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

Wind-induced densification of near surface snow layers ~~Compaction rate during snowdrift events~~

750 Following Brun et al. [1997], the compaction rate, $\tau_{drift}\Gamma_w$, of wind-induced densification of
751 near surface snow layers ~~the snow during snowdrift events~~ is computed using several steps. First, a
752 mobility index, Γ_{mob} , that describes the potential for snow erosion for each snow layer is computed
753 as a function of snow density:

$$756 \quad \Gamma_{mob}(i) = a_{mob} \left[1.0 - \max \left(0, \frac{\rho_s(i) - \rho_{snmin}}{\rho_{mob}} \right) \right] \quad (\text{B1})$$

757 where $\rho_{snmin} = 50\text{kg}\cdot\text{m}^{-3}$ is the minimum density of snow, ρ_{mob} a reference density of $295\text{kg}\cdot\text{m}^{-3}$,
758 and the dimensionless constant $a_{mob} = 1.25$. Secondly, a wind-driven compaction ~~driftability~~ index,
759 $F_{drift}\Gamma_w$, combining the mobility index and the near surface atmospheric wind speed:

$$760 \quad \Gamma_w(i) = 1 - a_\Gamma \exp(-b_\Gamma \kappa_v V_a) + \Gamma_{mob}(i) \quad (\text{B2})$$

761 where $\kappa_v = 1.25$ is a dimensionless coefficient for gust diagnosis from average wind speed, and the
762 constants $a_\Gamma = 2.868$ and $b_\Gamma = 0.085 \text{ s}\cdot\text{m}^{-1}$. A positive value of $F_{drift}\Gamma_w$ indicates that wind-driven

763 ~~compactionsnowdrifting~~ can occur. Compaction rate from the surface is then propagated to the
 764 layers beneath, following an exponential decrease, until it meets a snow layer having a negative
 765 ~~wind-driven compactiondriftability~~ index. For each layer, this compaction rate is computed as
 766 follows:

$$767 \quad \tau_w(i) = \frac{2\kappa_v \pi_\tau}{f_\tau(i)} \quad \text{with} \quad f_\tau(i) = \max(0, \Gamma_w(i)) \times \exp\left(-a_\tau \sum_{j=1}^i (\Delta z(j)(b_\tau - \Gamma_w(j)))\right) \quad (\text{B3})$$

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

769 APPENDIX C

770 *Grid-cell snow fraction*

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

$$775 \quad p_{sn} = (1 - f_{veg})p_{sng} + f_{veg}p_{snv} \quad \text{with} \quad \begin{cases} p_{sng} = \min(1, h_{sn}/h_{sng}) \\ p_{snv} = h_{sn}/(h_{sn} + w_{snv}z_{0veg}) \end{cases} \quad (\text{C1})$$

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

$$782 \quad z_{0veg} = \max(0.001, 0.13 \times h_{veg}) \quad (\text{C2})$$

783 For woody vegetation, h_{veg} is assumed constant over time. It ranges from 30m for tropical forests
 784 and 20m for coniferous boreal forests to 15m, 10m or 5m for temperate forests and 2m for bushes.
 785 For herbaceous plants, $h_{veg} = LAI/6$, with LAI the leaf area index given by the ECOCLIMAP

786 database. It ranges approximately from 0.01m to 0.8m for grassland/tundra. Finally, the height of
787 crop types is related to an exponential function of LAI and has a height of 1m before maturity
788 defined as a LAI of $3.5 \text{ m}^2 \cdot \text{m}^{-2}$. More details on these physiographic parameters can be found in
789 Masson et al. [2003].

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1031

TABLE CAPTIONS

1031

1032 **Table 1** – The peat soil hydraulic and thermal parameter values used in ISBA for fibric and sapric
 1033 soil. w_{sat} ($m^3.m^{-3}$) is the porosity, w_{fc} ($m^3.m^{-3}$) the water content at field capacity specified as
 1034 matric potential at -0.1 bar for peat soil, w_{wilt} ($m^3.m^{-3}$) the water content at wilting point (matric
 1035 potential of -15 bar), b the dimensionless shape parameter of the soil-water retention curve, ψ_{sat}
 1036 (m) the soil matric potential, k_{sat} ($m.s^{-1}$) the soil hydraulic conductivity at saturation, c ($J.m^{-3}.K^{-1}$)
 1037 the soil heat capacity of organic matter, λ_s ($W.m^{-1}.K^{-1}$) the thermal conductivity of soil matrix, and
 1038 λ_{dry} ($W.m^{-1}.K^{-1}$) the dry soil thermal conductivity. For pedotransfer functions of Boelter [1969], the
 1039 fiber content in fibric soil is assumed to be equal to 76.8 % against 21.8 % in sapric soil in order to
 1040 reach soil porosity values close to Letts et al. [2000]. The method for averaging mineral soil
 1041 properties with peat soil values using the fraction of soil that is organic is also given for each
 1042 parameter.

α_{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
ψ_{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
λ_s	0.25	0.25	Farouki [1986]	Geometric
λ_{dry}	0.05	0.05	Farouki [1986]	Geometric

1043

1043 **Table 2** – Daily skill scores simulated by each experiment at Col de Porte for snow depth, SWE,
1044 albedo and soil temperature at 10cm over the number of point measurement, n . The bias, centred
1045 root mean square errors (c-rmse) and square correlation (r^2) described in section 4.1 are shown.
1046 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	0.074	0.089
		<i>c-rmse</i>	0.159	0.157	0.126	0.130
		r^2	0.863	0.870	0.907	0.900
	<i>MAM</i> ($n=1656$)	<i>bias</i>	0.165	0.127	0.077	0.027
		<i>c-rmse</i>	0.223	0.192	0.169	0.155
		r^2	0.845	0.878	0.884	0.900
	<i>All</i> ($n=4737$)	<i>bias</i>	0.102	0.082	0.053	0.041
		<i>c-rmse</i>	0.176	0.157	0.130	0.126
		r^2	0.889	0.908	0.923	0.927
<i>SWE (kg.m⁻²)</i>	<i>DJF</i> ($n=835$)	<i>bias</i>	12.329	6.196	4.934	8.887
		<i>c-rmse</i>	38.331	35.004	34.476	36.079
		r^2	0.901	0.913	0.915	0.911
	<i>MAM</i> ($n=887$)	<i>bias</i>	25.022	19.064	16.352	0.334
		<i>c-rmse</i>	61.138	57.204	55.699	49.583
		r^2	0.861	0.872	0.876	0.900
	<i>All</i> ($n=2310$)	<i>bias</i>	13.851	9.169	7.648	2.981
		<i>c-rmse</i>	45.641	42.267	41.134	38.100
		r^2	0.902	0.910	0.913	0.924
<i>Albedo (-)</i>	<i>DJF</i> ($n=1456$)	<i>bias</i>	0.047	0.047	0.047	0.045
		<i>c-rmse</i>	0.076	0.076	0.076	0.074
		r^2	0.528	0.535	0.533	0.506
	<i>MAM</i> ($n=1516$)	<i>bias</i>	0.077	0.077	0.076	0.023
		<i>c-rmse</i>	0.119	0.117	0.115	0.080
		r^2	0.768	0.785	0.792	0.889
	<i>All</i> ($n=4101$)	<i>bias</i>	0.048	0.046	0.045	0.026
		<i>c-rmse</i>	0.101	0.098	0.098	0.082
		r^2	0.858	0.869	0.871	0.905
<i>Soil temperature 10cm (K)</i>	<i>DJF</i> ($n=1323$)	<i>bias</i>	-1.082	-1.009	-0.962	-1.032
		<i>c-rmse</i>	0.892	0.837	0.797	0.811
		r^2	0.234	0.234	0.272	0.279
	<i>MAM</i> ($n=838$)	<i>bias</i>	-0.646	-0.624	-0.606	-0.199
		<i>c-rmse</i>	2.109	1.995	1.967	1.701
		r^2	0.827	0.848	0.852	0.896
	<i>All</i> ($n=2237$)	<i>bias</i>	-1.121	-1.079	-1.049	-0.936
		<i>c-rmse</i>	1.650	1.591	1.569	1.519
		r^2	0.871	0.880	0.883	0.894

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1047

FIGURE CAPTIONS

1048 **Figure 1** – Spatial distribution of the observed soil organic carbon content over two soil horizon
1049 (0-30cm and 30-70cm) at 0.5° by 0.5° resolution. Observations come from the Harmonized World
1050 Soil Database at 1 km resolution of the Food and Agricultural Organization.

1051 **Figure 2** – Parameterization of the effect of soil organic carbon (SOC) on soil hydraulic and
1052 thermal properties. The soil organic carbon density profile, ρ_{soc} , is given by Equation 21 using a
1053 top soil organic carbon content of 10 kg.m⁻², a sub soil content of 15 kg.m⁻², and via a simple
1054 linear interpolation at each soil grid nodes that conserves the total soil carbon mass. The fraction of
1055 the soil that is organic, f_{soc} , in each layer is determined assuming a simple relationship between this
1056 last soil organic carbon density profile and an idealized peat soil density profile (Equation 22).
1057 Examples for the soil porosity, w_{sat} , the soil saturated hydraulic conductivity, k_{sat} , and the soil heat
1058 capacity, c , are given. Dotted lines represent vertical homogeneous mineral soil properties, dashed
1059 lines the idealized peat soil properties, and plain lines the resulting combined soil properties using
1060 averaging method sums-up in Table 1.

1061 **Figure 3** – Overview of the four experiments performed at the Col de Porte field site. Daily
1062 simulated and observed data for snow depth (top) and SWE (bottom) are provided for 18 and 10
1063 years respectively. In-situ observations are in black, the *CTL* simulation in blue, *SNL* in green,
1064 *CPT* in orange, and *NEW* in red. The corresponding statistics are given in terms of annual bias and
1065 c-rmse for each year by measurements periods.

1066 **Figure 4** – As Figure 3 but for surface albedo (top) and soil temperature at 10 cm depth (bottom).

1067 **Figure 5** – Daily mean annual cycles of snow depth, SWE, surface albedo, and soil temperature at
1068 10 cm depth simulated (colours) and observed (black) at the Col de Porte field site. The
1069 corresponding skill scores are given in Table 2. Over all panels, the grey shadow corresponds to
1070 the uncertainty in in-situ measurements as discussed in section 4.1. The observed snow depth
1071 exhibits an accuracy of ±1cm, the soil temperature is measured with a precision of ±1K, while
1072 uncertainties in SWE and surface albedo is near ±10%.

1073 **Figure 6** – Daily mean annual cycles of snow density ($\text{kg}\cdot\text{m}^{-3}$) and snowpack internal temperature
1074 ($^{\circ}\text{C}$) simulated by the four experiments over 18 years at the Col de Porte field site.

1075 **Figure 7** – Quantitative comparison between observed (plain circles) and simulated (plain fields)
1076 daily snow depth and monthly soil temperature at 20cm and 160cm depths over the Northern-
1077 Eurasia. Results from the bare soil sub-grid patch alone of the *NEW-SOC* simulation are presented
1078 because in-situ measurements have been collected in open areas following the WMO standards as
1079 mentioned in section 5.1.

1080 **Figure 8** – Mean annual cycles of observed and simulated daily snow depth and monthly soil
1081 temperature profiles. The mean cycles are computed by averaging all simulated or observed mean
1082 annual cycles at each station. However, [globaltotal](#) skill scores (bias and c-rmse) found in each
1083 panel are computed merging together all simulated or observed time series of all stations over the
1084 entire observed periods.

1085 **Figure 9** – Daily snow depth skill scores (bias and c-rmse) simulated by the *CTL* and the *NEW*
1086 experiments during winter (DJF) and spring (MAM) over the Northern-Eurasia and expressed in
1087 meters. [Global-Total](#) scores given between parentheses are computed by merging together all
1088 simulated or observed daily time series of all stations for each season.

1089 **Figure 10** – Monthly soil temperature skill scores at 20cm and 160cm depths simulated by the
1090 *CTL* and the *NEW* experiments during winter and expressed in degrees Celsius. [Global-Total](#)
1091 scores (bias and c-rmse) are given for each panel.

1092 **Figure 11** – Monthly soil temperature profile bias simulated by the *NEW* (left) and the *NEW-SOC*
1093 (right) experiments during summer and expressed in degrees Celsius. [Global-Total](#) skill scores
1094 (bias; c-rmse) are given in the top-panel for each soil horizon.

1095 **Figure 12** – Distribution of permafrost characteristics. The NSIDC estimated limits of continuous,
1096 discontinuous, sporadic and isolated permafrost regions are shown in the top panel. In each panel
1097 the red lines correspond to the observed boundary of the entire permafrost region. [In the middle](#)
1098 [and the bottom panels, the mean active layer thicknesses simulated over the 1990-2013 period by](#)

1099 the *NEW* and the *NEW-SOC* experiments are shown and compared to observations from the
1100 CALM network (circles). Total skill scores are given for each experiment.~~The all-year mean~~
1101 ~~maximum active layer thicknesses simulated by the *NEW* and the *NEW-SOC* experiments are~~
1102 ~~shown in the middle and the bottom panels respectively.~~

1103 **Figure 13** – Estimated and simulated active layer thicknesses over the Yakutia region. Estimations
1104 before the 1990s are given by Beer et al. [2013] ~~and while~~ the *NEW* and the *NEW-SOC*
1105 experiments are ~~represented~~averaged over the 1979-1990 period. The estimated and simulated
1106 latitudinal zonal averages are shown over the last panel where Beer et al. [2013] estimations are in
1107 black, *NEW* in blue and *NEW-SOC* in red. Dashed lines correspond to uncertainties in active layer
1108 thicknesses estimations computed using standard deviations provided with the dataset.