

Incorporating organic soil into a global climate model

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Abstract Organic matter significantly alters a soil's thermal and hydraulic properties but is not typically included in land-surface schemes used in global climate models. This omission has consequences for ground thermal and moisture regimes, particularly in the high-latitudes where soil carbon content is generally high. Global soil carbon data is used to build a geographically distributed, profiled soil carbon density dataset for the Community Land Model (CLM). CLM parameterizations for soil thermal and hydraulic properties are modified to accommodate both mineral and organic soil matter. Offline simulations including organic soil are characterized by cooler annual mean soil temperatures (up to $\sim 2.5^{\circ}\text{C}$ cooler for regions of high soil carbon content). Cooling is strong in summer due to modulation of early and mid-summer soil heat flux. Winter temperatures are slightly warmer as organic soils do not cool as efficiently during fall and winter. High porosity and hydraulic conductivity of organic soil leads to a wetter soil column but with comparatively low surface layer saturation levels and correspondingly low soil evaporation. When CLM is coupled to the Community Atmosphere Model, the reduced latent heat flux drives deeper boundary layers, associated reductions in low cloud fraction, and warmer summer air temperatures in the Arctic. Lastly, the insulative properties of organic soil reduce interannual soil temperature variability, but only

marginally. This result suggests that, although the mean soil temperature cooling will delay the simulated date at which frozen soil begins to thaw, organic matter may provide only limited insulation from surface warming.

Keywords Soil carbon · Soil temperature · Land-surface scheme · Permafrost · Climate modeling

1 Introduction

Organic material is prevalent in soils in most vegetated regions of the world. It is found in varying amounts in different regions and accumulates (or decays) at varying rates according to the balance between the accumulation and decomposition of dead vegetation. In the Arctic, where cold temperatures inhibit decomposition of dead vegetation, soil organic material can build up over time forming peat deposits. The thermal and hydraulic properties of organic soil differ significantly from those of mineral soil. Organic material acts as an insulator, with its low thermal conductivity and relatively high heat capacity modulating the transfer of energy down into the soil during spring and summer and out of the soil during fall and winter, typically leading to cooler soil temperatures (Bonan and Shugart 1989). Partly as a consequence of this, permafrost, defined as soil or rock that remains below 0°C for two or more consecutive years, is present at warmer annual mean air temperatures when below organic soils than would be the case under mineral soil alone. An additional fundamental characteristic of organic or peat soil is its high porosity, much higher than that of mineral soils, and its correspondingly high hydraulic conductivity and weak suction. These characteristics generate soil conditions in peatlands that are typified by saturated sub-surface soils with shallow

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depths to the water table but highly variable surface soil wetness (Hinzman et al. 1991).

Prior modeling studies have shown that inclusion of the thermal and hydraulic effects of organic soil can have a significant impact on surface energy fluxes, air temperature (Peters-Lidard et al. 1998), soil temperature (Beringer et al. 2001; Mölders and Romanovsky 2006; Yi et al. 2006; Nicolsky et al. 2007) and soil moisture (Letts et al. 2000). Despite its importance to the surface energy and hydrologic budgets, organic material is not typically included in surface datasets and parameterizations for land surface schemes used in global climate models (GCMs). Normally, model soil thermal and hydraulic properties are defined via mineral soil parameterizations that are dependent on soil texture (e.g. after the work of Clapp and Hornberger 1978; Cosby et al. 1984). Soil sand, clay, and silt percentages are provided as a model input dataset to characterize soil texture. Across much of the world, neglecting soil organic material is a reasonable simplification as it accounts for only a small fraction of total soil mass. In the Arctic and boreal zones, however, low vegetation decomposition rates have led to the accumulation of large quantities of soil organic matter. It is in precisely these regions that warming associated with rising greenhouse gas concentrations is being realized most acutely (Serreze et al. 2000; Holland and Bitz 2003; Hinzman et al. 2005).

Among the many transformations being observed in the Arctic is widespread permafrost warming and thawing (Carmill 2005; Smith et al. 2005b; Jorgenson et al. 2006; Osterkamp and Jorgenson 2006). Permafrost thaw is projected to continue and maybe even accelerate in the coming decades (Stendel and Christensen 2002; Lawrence and Slater 2005) although the rate at which it will thaw remains highly uncertain (Burn and Nelson 2006; Lawrence and Slater 2006). If large-scale thawing of near-surface permafrost occurs, it is likely to invoke a number of feedbacks in the Arctic and global climate system (McGuire et al. 2006). For example, permafrost thaw alters soil structural and hydrologic properties with potential knock-on effects on the spatial extent of lakes and wetlands (Smith et al. 2005a), freshwater fluxes to the Arctic Ocean, ecosystem functioning and structure (Payette et al. 2004; McGuire et al. 2006), and the surface energy balance. Warming of the soil is also expected to enhance decomposition of soil organic matter, possibly producing vast quantities of carbon dioxide and/or methane that could be released to the atmosphere (Walter et al. 2006; Zimov et al. 2006) and at the same time releasing nitrogen which, in nutrient limited Arctic ecosystems, may prompt shrub growth (Sturm et al. 2001; Sturm et al. 2005; Tape et al. 2006). Expansion of shrub cover has its own positive albedo feedback on climate (Chapin et al. 2005).

Improved understanding, simulation, and prediction of the rich complexity and interrelationships between climate-

change-induced thawing of permafrost and the global carbon and hydrologic cycle is a challenging problem that will require investigation with a hierarchy of models, including GCMs which can represent, through parameterizations, the broad changes in climate, hydrology, and vegetation that can occur with climate change. Progress towards better understanding and simulation of these feedbacks depends critically upon land-surface scheme simulation of soil temperature and soil moisture and their response to warming. As a step towards improved representation of high-latitude ground climate, we describe in this paper the incorporation of the thermal and hydraulic properties of organic soil into the Community Land Model (CLM), which is the land-surface scheme used in the NCAR Community Climate System Model (CCSM). In Sect. 2, we introduce a global soil carbon profile dataset for CLM that is based on global estimates of soil carbon content. Also in Sect. 2, we describe modifications to CLM soil thermal and hydraulic property parameterizations to reflect properties of organic soil material. In Sect. 3, results from offline and land-atmosphere coupled simulations both with and without soil organic matter are presented. The emphasis of the sensitivity analysis is on the influence of organic soil on soil temperature, soil moisture, and the surface energy balance. In Sect. 4, we discuss the implications for transient climate change simulations and review further potential improvements to the model. We conclude with a summary in Sect. 5.

2 Model description

2.1 CLM and CAM

We use the Community Land Model version 3 (CLM3, Oleson et al. 2004) in both offline mode and coupled to the Community Atmosphere Model version 3 (CAM3, Collins et al. 2006). CLM3 includes a 5-layer snow model that sits atop a 3.43 m-deep soil layer with ten layers of exponentially increasing depth. The thermal and hydrologic properties of the soil are functions of soil liquid and ice water content, soil texture, and soil temperature (further details in Sect. 2.3). The effects of phase change are included. Snow processes include accumulation, melt, compaction, and water transfer across layers. Sub-grid scale surface heterogeneity is represented through satellite-derived fractional coverage of lakes, wetland, bare soil, glacier, and up to four plant functional types in each grid box (dataset derived from MODIS, Lawrence and Chase 2007). Fluxes of energy and moisture are modeled independently for each surface type and aggregated before being passed to the atmosphere model.

The released version of CLM3 suffers from poor partitioning of evapotranspiration into its components (tran-

spiration, soil evaporation, and canopy evaporation) and generally dry and unvarying deep soil moisture (Lawrence et al. 2007). These deficiencies have largely been eliminated through a series of modifications to the released version and include a revised surface dataset (Lawrence and Chase 2007), reductions in canopy interception and incorporation of a two-leaf model (Thornton and Zimmerman 2006), and a reworking of the soil hydrology scheme (Niu et al. 2007). Of direct relevance to this paper is the inclusion of a freezing-point depression expression that permits liquid water to coexist with ice at temperatures below 0°C and enhances permeability through partially ice-filled soil (Niu and Yang 2006).

2.2 Global soil carbon content profile dataset

Global information on soil carbon density has been compiled by the Global Soil Data Task (2000) under the auspices of the International Satellite Land-Surface Climatology Project II (ISLSCP II). The soil carbon density dataset is an outgrowth of an initial effort by Zinke et al. (1986) to gather and compile soil carbon content measurements. This data is used to derive a gridded soil carbon content of the upper 150 cm of soil at $1^\circ \times 1^\circ$ resolution. Soil carbon content is high throughout much of the northern circumpolar region, with values exceeding 50 kg m^{-2} found in Siberia, Canada, and the Nordic countries (see map, Fig. 1a). Outside of the northern high-latitudes, soil carbon content in vegetated areas typically ranges between 10 and 20 kg m^{-2} while semi-arid or arid regions are characterized by soil carbon contents of less than 10 kg m^{-2} . We regrid the source data to the relevant

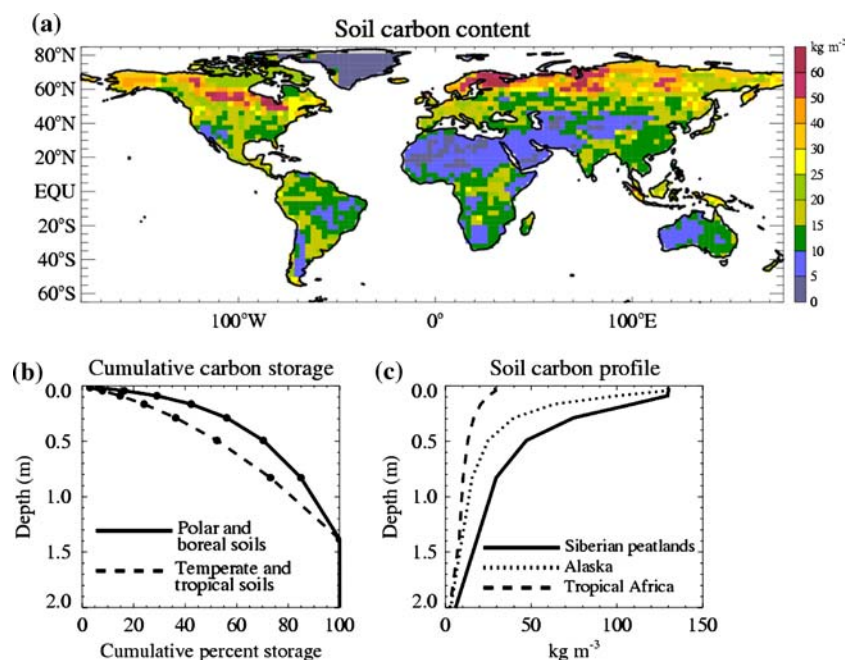
CLM resolution (2.8° longitude by 2.8° latitude for this study) and distribute the soil carbon across the top seven CLM soil layers (surface to 1.38 m depth) according to typical soil carbon profiles shown in Zinke et al. (1986). The profiles differ according to ecozone with carbon content concentrated towards the surface in boreal and polar soils and more evenly distributed throughout the upper soil column in tropical and temperate soils (Fig. 1b). A check in the vertical distribution calculation requires that the carbon density in any individual soil layer does not exceed the bulk density of peat (130 kg m^{-3} , Farouki 1981). The result is a new surface dataset for CLM that provides carbon density (kg m^{-3}) for each soil layer for each land grid location. Sample carbon density profiles are shown in Fig. 1c.

A revised estimate of high-latitude soil carbon content is being derived as part of the International Permafrost Association Carbon Pools in Permafrost (CAPP) project. This new dataset, known as the Northern Circumpolar Soil Carbon Database, may provide an improved estimate of soil carbon content in the northern high-latitudes at some point in the near future (C. Tarnocai, personal communication).

2.3 Thermal and hydraulic parameterizations for organic and mineral soil

In order to incorporate the impacts of organic material on soil thermal and hydraulic properties, we assume that each soil layer is made up of a combination of organic material, as provided by the dataset described in Sect. 2.2, and mineral material, as given in the standard CLM soil dataset

Fig. 1 **a** Global Soil Data Task (2000) soil carbon content regridded onto CLM $2.8^\circ \times 2.8^\circ$ grid. **b** Cumulative carbon storage with depth for the two major classes of soils identified in Zinke et al. (1986). These profiles are used to determine vertical soil carbon distribution in new CLM soil carbon dataset. **c** Sample derived soil carbon profiles for Siberian peatlands ($60^\circ\text{--}70^\circ\text{N}$, $70^\circ\text{--}80^\circ\text{E}$), Alaska ($60^\circ\text{--}70^\circ\text{N}$, $140^\circ\text{--}160^\circ\text{W}$), and Tropical Africa (Eq– 10°N , $25^\circ\text{--}35^\circ\text{E}$). Note that depth refers to the depth of the bottom boundary of each individual soil layer (solid circles in **b**)



which contains information on the mineral soil texture as percentage sand and clay contents for each of the ten model soil layers. We define the soil carbon or organic fraction for a particular soil layer as

$$f_{sc,i} = \rho_{sc,i} / \rho_{sc,max} \quad (1)$$

where $\rho_{sc,i}$ is the soil carbon density for soil layer i and $\rho_{sc,max} = 130 \text{ kg m}^{-3}$ is the maximum soil carbon density (equivalent to a standard bulk density of peat; Farouki 1981). Physical properties of the soil are assumed to be a weighted combination of values for mineral soil and values for pure organic soil. For example, the volumetric water content at saturation (porosity) for mineral soil, $\Theta_{sat,min}$, is defined in CLM as

$$\Theta_{sat,min,i} = 0.489 - 0.00126(\%sand)_i \quad (2)$$

after Clapp and Hornberger (1978) and Cosby et al. (1984). The porosity of organic material is much higher than that of mineral soil with typical values ranging between 0.8 and 0.95 (Hinzman et al. 1991; Letts et al. 2000). Incorporating organic soil, the equation for Θ_{sat} becomes

$$\Theta_{sat,i} = (1 - f_{sc,i})\Theta_{sat,min,i} + f_{sc,i}\Theta_{sat,sc} \quad (3)$$

where $\Theta_{sat,sc}$ is assumed to be 0.9. Hence, for low soil carbon density, porosity is similar to that of pure mineral soil whereas for high soil carbon density porosity approaches that of pure organic soil or peat. For mid-range carbon densities, porosity is a weighted combination of mineral and organic values. Note that thermal and hydraulic properties of organic material depend on the state of decomposition of the organic material (Letts et al. 2000; Quinton et al. 2000). For the sake of simplicity and because we do not have information on the extent of decomposition of the soil carbon material, the parameters used in this model and listed in Table 1 are standard literature values for undecomposed organic material (also known as fibric peat).

This strategy for implementation of organic soil differs somewhat from some previous approaches wherein organic soil has been incorporated by assuming that one or more layers of pure organic soil overly mineral soil layers at depth (e.g. Letts et al. 2000; Beringer et al. 2001). The advantage of our weighted combination approach is that it is more general in that it can be applied at all locations, even those without high soil carbon contents. Furthermore, it can be implemented straightforwardly within the existing CLM framework and it can be used when the model is run with an active terrestrial carbon cycle with prognostic soil carbon pools (Thornton and Rosenbloom 2005). For locations with high soil carbon content, e.g. much of the

northern high-latitudes, the first 0.1–0.5 m (1–4 layers) of soil are 100% organic and therefore approximate the implementations of Letts et al. (2000) and Beringer et al. (2001) except with a more smoothly varying transition from organic to mineral soil at mid-level soil layers. Our assumption that the thermal and hydraulic properties of coexisting mineral and organic soil material can be approximated as a weighted combination of the separate mineral soil and organic soil properties is difficult to validate or refute as measured properties for coexisting mineral and organic soils are not readily found in the literature, either because mineral and organic soils do not tend to coexist or because measurements have simply not been taken because there was not a perceived need for such information. Further details on the new parameterizations are given in the remainder of this section.

2.3.1 Thermal conductivity

Soil organic material significantly alters heat conduction into and through the soil due to its low thermal conductivity and relatively high heat capacity. Soil heat conduction is solved numerically in CLM via the diffusion equation,

$$c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left[\lambda_s \frac{\partial T}{\partial z} \right], \quad (4)$$

where c is the volumetric soil heat capacity ($\text{J m}^{-3} \text{K}^{-1}$), λ is soil thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), T is temperature (K), t is time (s) and z is depth (m). The CLM thermal conductivity parameterizations are predicated on the work of Johansen (1975) and Farouki (1981). Soil thermal conductivity is calculated as a combination of the dry λ_{dry} and saturated λ_{sat} thermal conductivities weighted by a normalized thermal conductivity (K_e , the Kersten number):

$$\lambda_i = K_{e,i}\lambda_{sat,i} + (1 - K_{e,i})\lambda_{dry,i}. \quad (5)$$

The dry and saturated thermal conductivities are dependent on the type of soil material, e.g. sand, clay, loam, or organic. Dry thermal conductivity for mineral soils, $\lambda_{dry,min}$, is given by the following empirical equation derived by Johansen:

$$\lambda_{dry,min,i} = \frac{0.135\rho_{d,i} + 64.7}{2700 - 0.947\rho_{d,i}} \quad (6)$$

where $\rho_{d,i} = 2700(1 - \Theta_{sat,min,i})$ is the bulk density of mineral soil. Farouki calculates that a typical dry thermal conductivity of organic soil, $\lambda_{dry,sc}$, is $0.05 \text{ W m}^{-1} \text{K}^{-1}$. The dry thermal conductivity for the combination of mineral and organic soil is then:

Table 1 Soil parameters used in CLM for two representative mineral soil texture classes, sand and clay, and the equivalent soil parameters for organic soil that are used in the new model

Soil type	λ_s (W m ⁻¹ K ⁻¹)	λ_{sat} (W m ⁻¹ K ⁻¹)	λ_{dry} (W m ⁻¹ K ⁻¹)	c_s (J m ⁻³ K ⁻¹ × 10 ⁶)	Θ_{sat}	k_{sat} (m s ⁻¹ × 10 ⁻³)	Ψ_{sat} (mm)	b
Sand (92% sand, 5% silt, 3% clay)	8.61	3.12	0.27	2.14	0.37	0.023	-47.3	3.4
Clay (22% sand, 20% silt, 58% clay)	4.54	1.78	0.20	2.31	0.46	0.002	-633.0	12.1
Peat (100% soil carbon)	0.25 ^a	0.55	0.05 ^a	2.5 ^a	0.9 ^{a,b}	0.100 ^b	-10.3 ^b	2.7 ^b

The soil parameters are λ_s is the thermal conductivity of soil solids, λ_{sat} is the unfrozen saturated thermal conductivity, λ_{dry} is the dry soil thermal conductivity, c_s is the soil solid heat capacity, Θ_{sat} is the saturated volumetric water content (porosity), k_{sat} is the saturate hydraulic conductivity, Ψ_{sat} is the saturated matric potential, and b is the Clapp and Hornberger parameter

^a Farouki (1981)

^b Letts et al. (2000)

$$\lambda_{\text{dry},i} = (1 - f_{\text{sc},i})\lambda_{\text{dry,min},i} + f_{\text{sc},i}\lambda_{\text{dry,sc}} \quad (7)$$

The saturated thermal conductivity is given by

$$\lambda_{\text{sat},i} = \lambda_{s,i}^{1-\Theta_{\text{sat},i}} \lambda_{\text{liq}}^{\Theta_{\text{sat},i}} \lambda_{\text{ice}}^{\Theta_{\text{sat},i}-\Theta_{\text{liq},i}} \quad (8)$$

where λ_{liq} and λ_{ice} are the liquid water and ice thermal conductivities, $\Theta_{\text{liq},i}$ is the volumetric liquid water content, and $\lambda_{s,i}$ is the soil solid thermal conductivity which is derived from the following equations:

$$\lambda_{s,\text{min},i} = \frac{8.80(\%\text{sand})_i + 2.92(\%\text{clay})_i}{(\%\text{sand})_i + (\%\text{clay})_i} \quad (9)$$

$$\lambda_{s,i} = (1 - f_{\text{sc},i})\lambda_{s,\text{min},i} + f_{\text{sc},i}\lambda_{s,\text{sc}} \quad (10)$$

where $\lambda_{s,\text{sc}} = 0.25 \text{ W m}^{-1} \text{ K}^{-1}$ is the thermal conductivity of organic soil solid (Farouki 1981) and $\lambda_{s,\text{min},i}$ is the empirically derived thermal conductivity of mineral soil solid.

The Kersten number $K_{e,i}$ is a function of the degree of saturation S_r and phase of water, for unfrozen soils,

$$K_{e,i} = \log(S_{r,i}) + 1, \quad (11)$$

and for frozen soils,

$$K_{e,i} = S_r. \quad (12)$$

The resulting soil thermal conductivity values for frozen and unfrozen soil are shown in Fig. 2 as a function of soil carbon fraction and percent saturation.

2.3.2 Heat capacity

Heat capacities of organic soil are typically higher than in mineral soils. Soil heat capacity is a function of the heat capacities of the soil solid, liquid water, and ice constituents

$$c_i = c_{s,i}(1 - \Theta_{\text{sat},i}) + w_{\text{liq},i}C_{\text{liq}} + w_{\text{ice},i}C_{\text{ice}} \quad (13)$$

where $w_{\text{liq},i}$ and $w_{\text{ice},i}$ are the liquid water and ice contents (kg m⁻³) for a layer and C_{liq} and C_{ice} are specific heat capacities (J kg⁻¹ K⁻¹) of liquid water and ice, respectively, and $c_{s,i}$ is given by

$$c_{s,i} = (1 - f_{\text{sc},i})c_{s,\text{min},i} + f_{\text{sc},i}c_{s,\text{sc}} \quad (14)$$

with

$$c_{s,\text{min},i} = \left(\frac{2.128(\%\text{sand})_i + 2.385(\%\text{clay})_i}{(\%\text{sand})_i + (\%\text{clay})_i} \right) \times 10^6 \quad (15)$$

and the heat capacity of organic soil, $c_{s,\text{sc}}$, equal to $2.5 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ (Farouki 1981).

2.3.3 Saturated hydraulic conductivity

The soil hydraulic characteristics are also affected by the organic content. The saturated hydraulic conductivity of peat is considerably higher than that of mineral soil. In the new model, saturated hydraulic conductivity k_{sat} is

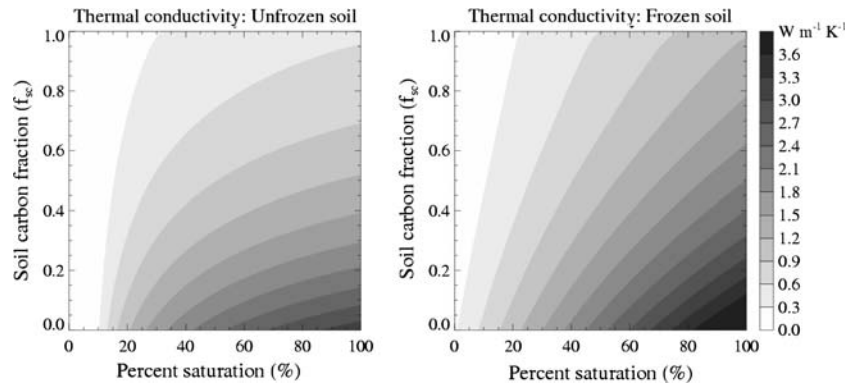
$$k_{\text{sat},i} = (1 - f_{\text{sc},i})k_{\text{sat,min},i} + f_{\text{sc},i}k_{\text{sat,sc}} \quad (16)$$

with $k_{\text{sat,min}}$ obtained from Clapp and Hornberger (1978) and Cosby et al. (1984) and given by

$$k_{\text{sat,min},i} = 0.0070556 \times 10^{-0.884+0.0153(\%\text{sand})_i} \quad (17)$$

As noted in Letts et al. (2000), literature values for peat saturated hydraulic conductivity vary by several orders of magnitude. A median value for undecomposed peat is $2.8 \times 10^{-4} \text{ m s}^{-1}$. Initial tests of the new model were conducted with $k_{\text{sat,sc}}$ values of $2.8 \times 10^{-4} \text{ m s}^{-1}$ but this introduced numeric instabilities so we settled on a slightly lower value ($k_{\text{sat,sc}} = 1.0 \times 10^{-4} \text{ m s}^{-1}$). Importantly, this

Fig. 2 Soil thermal conductivity as a function of percent saturation and soil carbon fraction. Base mineral soil is a sandy soil (92% sand, 5% silt, 3% clay). Thermal conductivities for unfrozen and frozen soil are shown. For these calculations, it is assumed that all soil water is either liquid or ice even though CLM retains a small fraction of liquid water at temperatures below 0°C



value remains about an order of magnitude larger than that of typical mineral soil.

2.3.4 Soil water retention

CLM follows the Campbell (1974) parameterization for soil water retention curves for mineral soils. In this parameterization, the soil matric potential (suction) (mm) is given by

$$\Psi_i = \Psi_{\text{sat},i} \left(\frac{\Theta_{\text{liq},i}}{\Theta_{\text{sat},i}} \right)^{-b_i} \quad (18)$$

where the saturated soil matric potential for mineral soil is

$$\Psi_{\text{sat,min},i} = -10.0 \times 10^{1.88 - 0.0131(\% \text{ sand})_i}$$

and the exponent b is the Clapp and Hornberger exponent. Letts et al. (2000) derive parameters for $\Psi_{\text{sat,sc}}$ (10.3 mm) and b_{sc} (2.7) for undecomposed peat that generate a Campbell soil water retention curve that fits available water retention data for peat. Incorporating the Letts parameters into our combined organic/mineral soil model, the equations for saturated matric potential and the b parameter become

$$\Psi_{\text{sat},i} = (1 - f_{\text{sc},i}) \Psi_{\text{sat,min},i} + f_{\text{sc},i} \Psi_{\text{sat,sc}} \quad (19)$$

$$b_i = (1 - f_{\text{sc},i}) b_{\text{min},i} + f_{\text{sc},i} b_{\text{sc}} \quad (20)$$

where $b_{\text{min},i} = 2.91 + 0.159(\% \text{ clay})_i$. The large pore spaces in organic soils lead to weak suction and freely draining soil over a wide range of moisture contents. The soil water retention curve (suction versus soil wetness) for peat is similar to that for sandy soil (Fig. 3).

3 Results

3.1 Offline simulations

Analysis of the new model is conducted first in offline mode, where CLM is forced with observed precipitation,

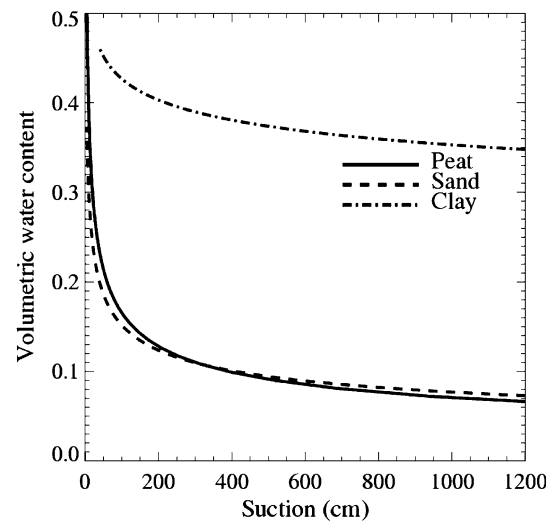


Fig. 3 Soil water retention curves for organic, sand, and clay soils

temperature, downward solar and longwave radiation, surface wind speed, specific humidity, and air pressure from a historical forcing dataset (Qian et al. 2006). Two 35-year (1970–2004) simulations are performed, a control simulation (CONTROL_{offn}) and a simulation incorporating the new soil carbon dataset and parameterizations (SOIL-CARB_{offn}). The deepest soil layer moisture and temperature fields reach equilibrium within the first 10–15 years of both simulations. The remaining 20 years are used to form climatological averages of soil temperature, soil moisture, and the surface energy balance.

3.1.1 Soil temperature

The most prominent impact of organic soil is a reduction in mean annual soil temperature. Reduced temperatures scale approximately linearly with increasing grid box soil carbon content although there is considerable spread due to the influence of local climatological and hydrological conditions (Fig. 4). The impact on soil temperature, however, is not uniform throughout the year. While strongly cooler soil

temperatures are seen during summer, the soils are often actually warmer in SOILCARB_{offn} during winter. The reasons for the seasonal differences in the impact of soil carbon on soil temperature are addressed later in this section. The relationships between soil carbon and soil temperature simulated by CLM are consistent with relationships reported in Harden et al. (2006). In that study, Harden et al. use soil temperature data collected from sites close to Delta Junction, Alaska that span a range of organic layer thicknesses to show that daily minimum July soil temperatures decrease roughly linearly with organic layer thickness (approximately $-0.5^{\circ}\text{C cm}^{-1}$ of organic mat thickness at 5 cm depth). Winter soil temperature was not collected, but data from the shoulder seasons (May and September) show a much weaker relationship between organic layer thickness and soil temperature, a result that is consistent with that seen in CLM (not shown).

Cooler annual mean soil temperatures in the northern high-latitudes reduce what appears to be a warm soil temperature bias in CLM3. We compare simulated soil temperature to a collection of Russian soil temperature data obtained from 418 sites spanning most of Siberia (43° – 69°N , 30° – 180°E , Zhang et al. 2001). Although the Russian soil temperature data extends back as far as 1891 for a few locations, we only analyze data from 1985 to the end of the dataset in 2000. Prior to analysis, the data is mapped onto the model's latitude–longitude grid. To minimize the influence of regridding edge effects, we only analyze gridded data from an interior portion of the domain (48° – 64°N , 45° – 165°E). Average soil carbon content in this region is relatively high (27 kg m^{-2}). The annual mean soil temperature bias is reduced at soil depths both near the surface (0.6 m, 0.73 to -0.04°C) and deeper in the soil (2.9 m, 0.48 to -0.27°C) (see Fig. 5). These improvements should be considered within the context that other limitations in CLM, such as the zero flux bottom boundary condition at the relatively shallow depth of 3.43 m, are also likely to affect soil temperature. Extension of the soil

column, and its impact on soil temperature and hydrology, is being considered as a potential further improvement to CLM (Alexeev et al. 2007; Nicolsky et al. 2007).

The change in soil temperature is not uniform across the seasons. Representative soil temperature and thermal conductivity vertical profiles over a climatological annual cycle, averaged over a region of high soil carbon content in north-west Siberia, are shown in Fig. 6. In CONTROL_{offn}, soils warm rapidly to depth after the April to May snowmelt before cooling off more slowly in late summer–early fall. Thermal conductivity, although slightly lower during the summer because of the lower thermal conductivity of liquid water relative to ice, stays above $2.0 \text{ W m}^{-1} \text{ K}^{-1}$ throughout the year. In contrast, in SOILCARB_{offn} the insulation provided by the organic content in near-surface soil layers, prevents the soil from warming as rapidly after snowmelt. Additionally, the warming front does not penetrate as deep into the soil column. The low thermal conductivity of the uppermost organic soil layers alters the annual cycle of ground heat flux (Fig. 7). From May to August, when the uppermost soil layer is warmer than the layers beneath it, ground heat flux is positive and heat is gained by the soil. The amount of heat gained is lower in SOILCARB_{offn} compared to CONTROL_{offn} due to the much lower thermal conductivity of the surface layers ($\sim 0.4 \text{ W m}^{-1} \text{ K}^{-1}$ compared to $\sim 2.0 \text{ W m}^{-1} \text{ K}^{-1}$). During September and October, air temperatures cool and heat is efficiently transferred out of the soil column, at least until snow begins to accumulate and insulate the soil from the cold atmospheric air. The same insulative properties of organic soil that prevent soil temperatures from warming during early summer restrict the extent of soil cooling in early fall. During winter, the highly insulative properties of snow and the relatively weak vertical soil temperature gradient limits heat loss from the soil in both experiments. In many locations (although not for the Siberia region shown in Figs. 6 and 7), and especially at lower latitudes (e.g. 30° – 60°N) where snow cover develops later in the

Fig. 4 Scatter diagram showing soil carbon content versus change in simulated column mean soil temperature (SOILCARB_{offn}–CONTROL_{offn}). Only data from grid points north of 45°N are plotted

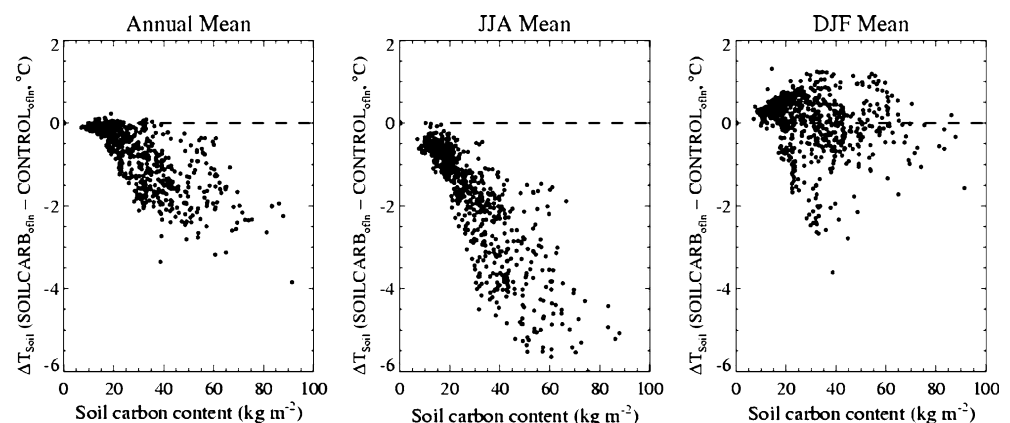
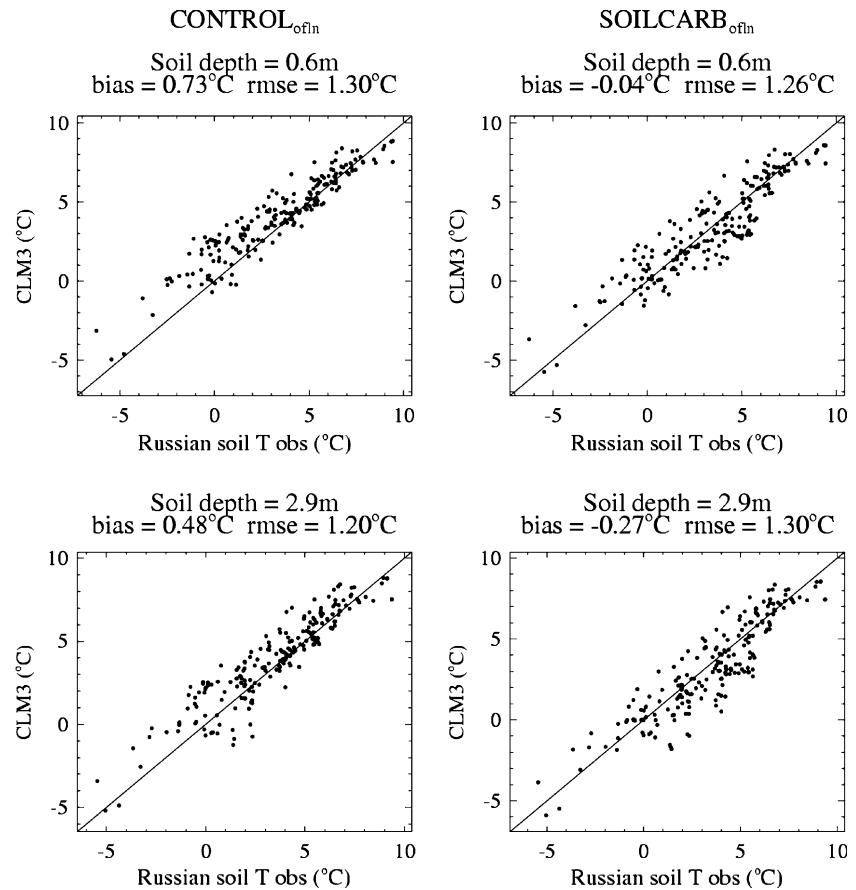


Fig. 5 Scatter diagram of observed Russian soil temperature data versus CLM soil temperature data for CONTROL_{offn} and SOILCARB_{offn} experiments at two soil depths



season or does not persist throughout the winter, the insulative properties of organic soil prevents the soils from cooling off during fall and winter resulting in warmer mean wintertime soil temperatures in SOILCARB_{offn} (Fig. 4).

For the period 1985–2004, annual mean ground heat flux in SOILCARB_{offn} is nearly identical to that in CONTROL_{offn} (difference $< 0.04 \text{ W m}^{-2}$), yet annual mean soil temperatures are on average 1.5°C cooler in SOILCARB_{offn}. If ground heat fluxes are the same, what explains the cooler soil in SOILCARB_{offn}? The difference in soil temperature, it turns out, develops during the first 5–10 years of the simulation when the effects of soil organic material restrict the heat gained or lost from the soil as the simulations come into balance. During this time period, annual mean heat flux into the ground is roughly $0.5\text{--}1.0 \text{ W m}^{-2}$ lower in SOILCARB_{offn} compared to CONTROL_{offn}, and soil temperatures in SOILCARB_{offn} steadily cool relative to CONTROL_{offn} (note that both simulations start from the same initial conditions). As the simulations reach equilibrium, annual mean heat flux in the two simulations converges to a value that reflects the energy imbalance forcing due to transient climate change. In experiments run with repeated single year forcing annual mean ground heat flux approaches zero after about 10 years.

3.1.2 Hydrology

The inclusion of soil organic matter also affects the hydrologic cycle in the form of changes in soil water content, runoff, and evaporation. The impact of organic soil on soil water storage is strongly tied to its impact on soil temperature. We have identified three distinct situations that summarize the impact on soil water content and saturation at high latitudes: (a) no perpetually frozen soil layer in either experiment, (b) no perpetually frozen soil layer in CONTROL_{offn} but at least one perpetually frozen soil layer in SOILCARB_{offn}, and (c) at least one perpetually frozen soil layer in both experiments. All land grid points north of 50°N and with soil carbon contents greater than 20 kg m^{-2} in each category are identified and averaged (weighted by grid box area) to determine annual cycles of the vertical profile of soil water content and percent saturation (Fig. 8). For locations where there is not a perpetually frozen soil layer in either experiment, the upper portion of the soil column (0.1–1.5 m deep) is slightly wetter in SOILCARB_{offn} as water fills the larger pore space of organic soil. In places where both experiments contain at least one perpetually frozen soil layer, lower temperatures and enhanced ice content coupled with higher porosity in SOIL-

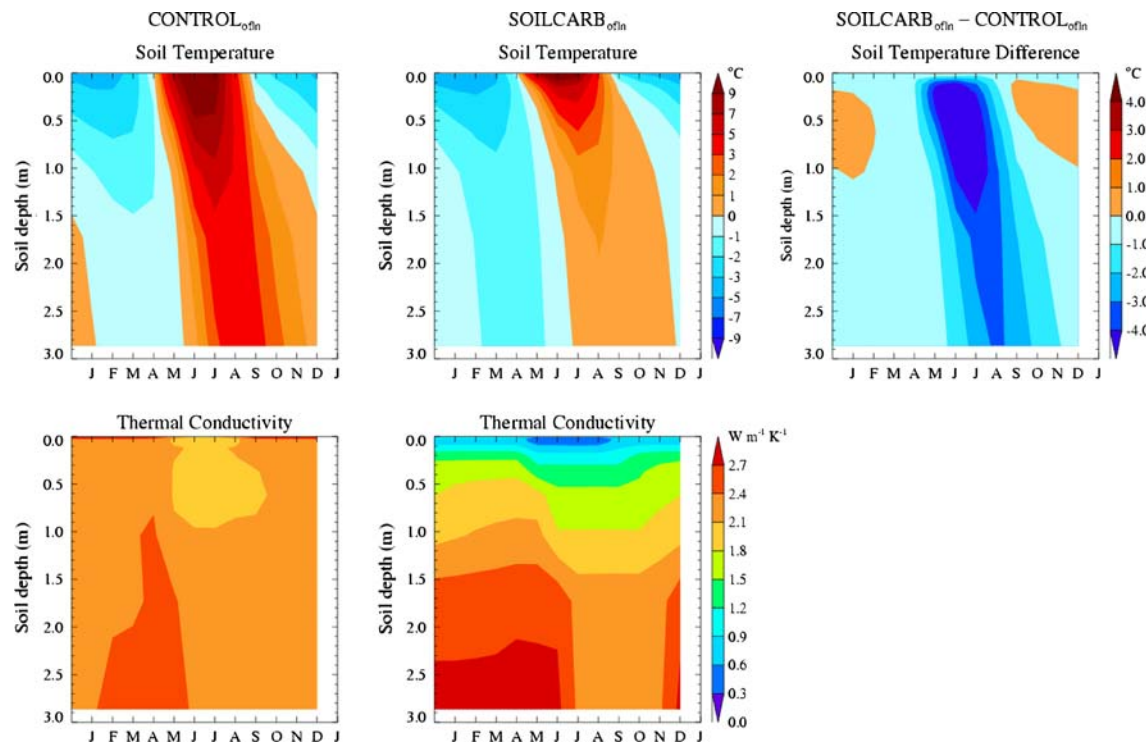


Fig. 6 Climatological mean vertical profiles of soil temperature and thermal conductivity over an annual cycle for $\text{CONTROL}_{\text{offn}}$ and $\text{SOILCARB}_{\text{offn}}$ averaged over region of high soil carbon content in

north-west Siberia (60° – 80°N , 70° – 90°E) where area mean soil carbon content = 49 kg m^{-2} . Also shown is vertical profile of change in soil temperature ($\text{SOILCARB}_{\text{offn}} - \text{CONTROL}_{\text{offn}}$)

$\text{CARB}_{\text{offn}}$ lead to an increase in total soil water content. Most of the column remains saturated throughout the year as the high ice content severely restricts movement of water out of the soil column. The biggest change in soil water content occurs for points that have new perpetually frozen layers in $\text{SOILCARB}_{\text{offn}}$. In this situation the soil column fills with soil water due to the sharp reduction in permeability through the new perennially frozen soil layers.

Although water content in the interior of the soil column is broadly higher in $\text{SOILCARB}_{\text{offn}}$, at the surface, the efficient transport of water down into the column due to the high hydraulic conductivity and low suction of organic soil results in a relatively dry surface soil layer. The low surface water content coupled with the high porosity yields sharply reduced percent saturation values (e.g. for category (a) above, July saturation levels are 69% for $\text{CONTROL}_{\text{offn}}$, 27% for $\text{SOILCARB}_{\text{offn}}$). In nature, organic soils are characterized in summer by nearly saturated sub-surface conditions with a drier, highly variable, surface layer (Hinzman et al. 1991). Strong variations in surface water content are a consequence of the observed efficient transport of water down into peatland soil columns (Quinton and Gray 2003). In the model, reduced surface layer saturation levels result in lower soil evaporation and associated reductions in total latent heat flux (Fig. 9). The reduction in latent heat flux (LH) is compensated for by a

rise in sensible heat flux (SH) and a corresponding rise in the Bowen ratio (SH/LH). This impact, and associated changes in surface air temperature, has an influence on climate in coupled atmosphere-land simulations (see Sect. 3.2). Note that although we show the impact of organic soil on surface hydrology for a particular region, e.g. the same north-west Siberia region considered previously and a region that in CLM is predominantly free of perpetually frozen layers (category (a) above), similar characteristics are seen for high soil carbon content locations throughout the northern high-latitudes.

Another interesting impact of organic soil seen in the model is on surface and sub-surface runoff. In $\text{SOILCARB}_{\text{offn}}$ there is a distinct increase in surface runoff during the spring snowmelt season. This increase is driven by higher surface layer soil ice content that prevents infiltration in $\text{SOILCARB}_{\text{offn}}$. The increase in ice content is a function of higher porosity and cooler soil temperatures in $\text{SOILCARB}_{\text{offn}}$. Once the surface layers have thawed, surface runoff nearly ends as most water infiltrates into the soil column. The annual mean increase in surface runoff is partially compensated for by a decrease in annual mean sub-surface runoff so that total runoff is relatively unaffected. In regions where the insulating properties of the soil organic matter lead to greater ice contents, the associated low hydraulic conductivities and deeper diagnosed water

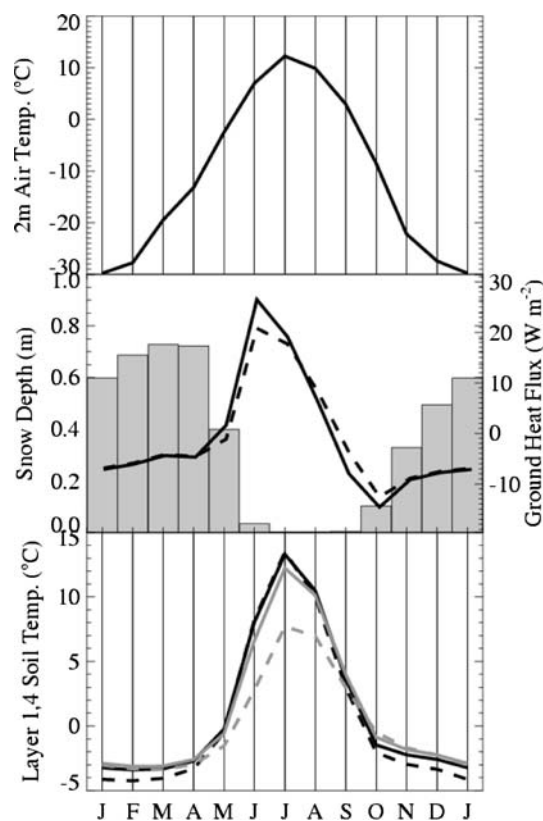


Fig. 7 Climatological annual cycle of 2 m air temperature, snow depth, ground heat flux, and layer 1, 0.01 m (black) and layer 4, 0.12 m (gray) soil temperature averaged over north-west Siberia (60°–80°N, 70°–90°E). Solid lines are CONTROL_{offn}; dashed lines are SOILCARB_{offn}. Note that 2 m air temperature and snow depth are almost identical in both experiments so only CONTROL_{offn} is shown

table due to the higher ice fraction work together to inhibit sub-surface runoff in SOILCARB_{offn}. In reality, sub-surface runoff through the upper organic layer can be very efficient due to its high hydraulic conductivity (Hinzman et al. 1991; Evans et al. 1999). The interactions between water table depth, sub-surface runoff, and evapotranspiration in peatlands (Fraser et al. 2001; Lafleur et al. 2005) and icy soils (Woo and Winter 1993) are clearly complex and likely require further consideration in CLM. It is worth noting, however, that the simulated monthly hydrograph of major high-latitude rivers such as the Yenisey is not sensitive to the inclusion of organic soil (not shown).

Although our focus has been on the high-latitudes where peatlands and other regions of characterized by soils with high organic matter contents tend to be found, including organic matter also impacts the simulations at middle and low latitudes where soil carbon content is generally lower. The impacts on soil hydrology are correspondingly weaker but, as seen for high latitudes, summertime and/or wet season soil evaporation (and latent heat flux) is generally lower in SOILCARB_{offn}. In some locations (e.g. in India,

see Fig. 10), the reduced soil evaporation can lead to enhanced soil water accumulation and storage during the wet season followed by increased soil evaporation during the dry season.

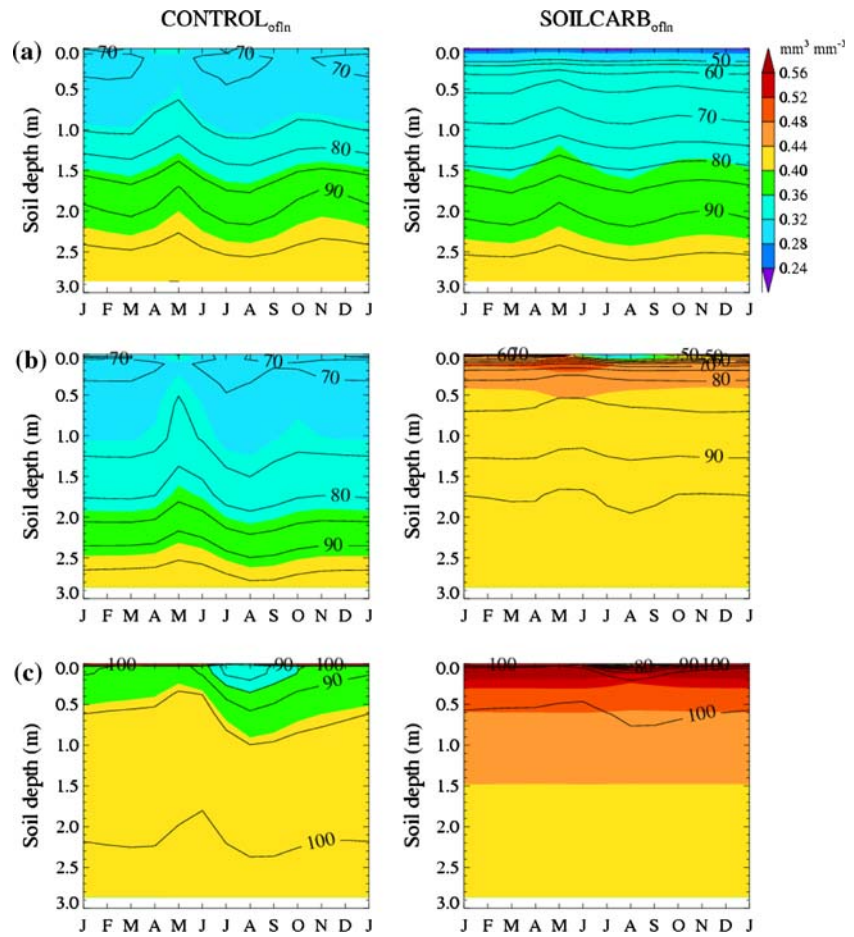
3.2 Coupled land–atmosphere simulations

To evaluate the impact of soil carbon on climate, we perform two AMIP-type CAM–CLM simulations forced with observed SSTs for the same 1970–2004 time period done for the offline experiments. Both experiments are initialized with spun-up land states from CONTROL_{offn}. The first 15 years are devoted to spin-up of the deep soil states and the remaining 20 years are analyzed for changes in climate.

As in the offline experiments, annual column mean soil temperatures are cooler in SOILCARB compared to CONTROL for soil carbon contents in excess of 40 kg m⁻², although the cooling is not as strong as in the offline experiments (Fig. 11). At soil carbon contents below 40 kg m⁻², soil temperatures are, on average, actually warmer in SOILCARB. The warmer soil temperatures are driven by warmer simulated summertime air temperatures. In Fig. 12 we present maps of climatological mean JJA fields for 2 m air temperature, column mean soil temperature, evaporative fraction (latent heat flux divided by latent heat flux plus sensible heat flux), boundary layer depth, low cloud fraction, and net radiation. In CAM3–CLM3, the presence of soil organic material contributes to warmer summer air temperatures across the Arctic region. The warming can be attributed to changes in the surface energy balance that affect the boundary layer, clouds, and the amount of incident solar radiation. Basically, a reduced latent heat flux from the moist but unsaturated surface peat layer (see Sect. 3.2) and associated increased sensible heat flux lead to a deeper and drier boundary layer and reduced occurrence of low level clouds. The reduction in low cloud fraction (mid and high cloud fractions are largely unchanged) permits more solar radiation to reach the earth's surface, thereby warming the surface and the near-surface air mass. The warmer air temperatures mitigate the summertime cooling of soil temperature that is induced by the insulative properties of organic soil material. Impacts on precipitation are negligible and not statistically significant (not shown). The changes in mean boreal winter climate are comparatively small.

The significance of these results are not the precise changes in air temperature but rather that soil organic material, through its impact on the surface energy balance, can alter the simulated climate. Due to climate models' differing response to changes in the surface energy balance, the details of the climate sensitivity to organic soil are likely to be dependent upon the characteristics of individual models.

Fig. 8 Climatological mean vertical profiles of volumetric soil water profile, liquid plus ice (*filled contours*) and percent saturation (*lined contours*). The profiles represent area averages across three different soil climate classes; **a** no perpetually frozen soil layer in either experiment, **b** no perpetually frozen soil layer in CONTROL_{offn} but at least one perpetually frozen soil layer in SOILCARB_{offn}, and **c** at least one perpetually frozen soil layer in both experiments. Only locations north of 50°N with soil carbon content exceeding 20 kg m⁻² in SOILCARB_{offn} are included in the averages



4 Discussion

Due to the central role that changes in permafrost are thought to have in the Arctic and global climate system, it is interesting to consider how the inclusion of organic material will alter the rate of permafrost thaw in transient climate change simulations. Clearly, the cooler annual mean soil temperatures will postpone permafrost thaw under Arctic warming as the soils have to warm further to even reach 0°C. One might anticipate also that the insulative properties of organic soil, coupled with the higher heat capacity of the wetter upper soil column, could to a certain degree protect permafrost from a transient surface warming (e.g. see processes described and analyzed in Hinkel et al. 2001). A rough analog to transient warming is the soil temperature response to interannual surface temperature variability. If organic material is insulating the soil from interannual surface temperature variations in CLM then one would expect to see reduced soil temperature variability in SOILCARB_{offn}. Indeed, interannual soil temperature variability is lower in SOILCARB_{offn}, although only slightly and only in the uppermost part of the soil (~10–15% lower in the top 1.0 m, Fig. 13). Interannual

variability of annual mean ground heat flux (σ_{GH}) is similarly reduced by about 5–20%: $\sigma_{GH} = 0.67 \text{ W m}^{-2}$ CONTROL_{offn}, 0.54 W m^{-2} SOILCARB_{offn} averaged for the same grid points considered for Fig. 13; $\sigma_{GH} = 0.80 \text{ W m}^{-2}$ CONTROL_{offn}, 0.77 W m^{-2} SOILCARB_{offn} averaged for all grid points north of 45°N with soil carbon content > 20 kg m⁻². This result suggests that organic soil should damp the response of the upper soil to surface warming, but not dramatically.

While organic soil appears to result in general improvements to the soil thermal regime, some discrepancies remain between CLM and observed soil temperatures in high latitude regions. These discrepancies may be due to a number of possible factors including the lack of deep cold layers that provide thermal inertia to soil temperatures (Alexeev et al. 2007; Nicolsky et al. 2007), CLM's tendency towards excessive autumn snow accumulation and early spring melt (Slater et al. 2006), and the absence of moss and lichen in the characterization of the land surface. Mosses are commonly found in boreal forest and tundra ecosystems and, like organic soil, significantly affect the surface energy balance and hydrologic cycles (Bonan and Shugart 1989). The soil thermal and hydraulic

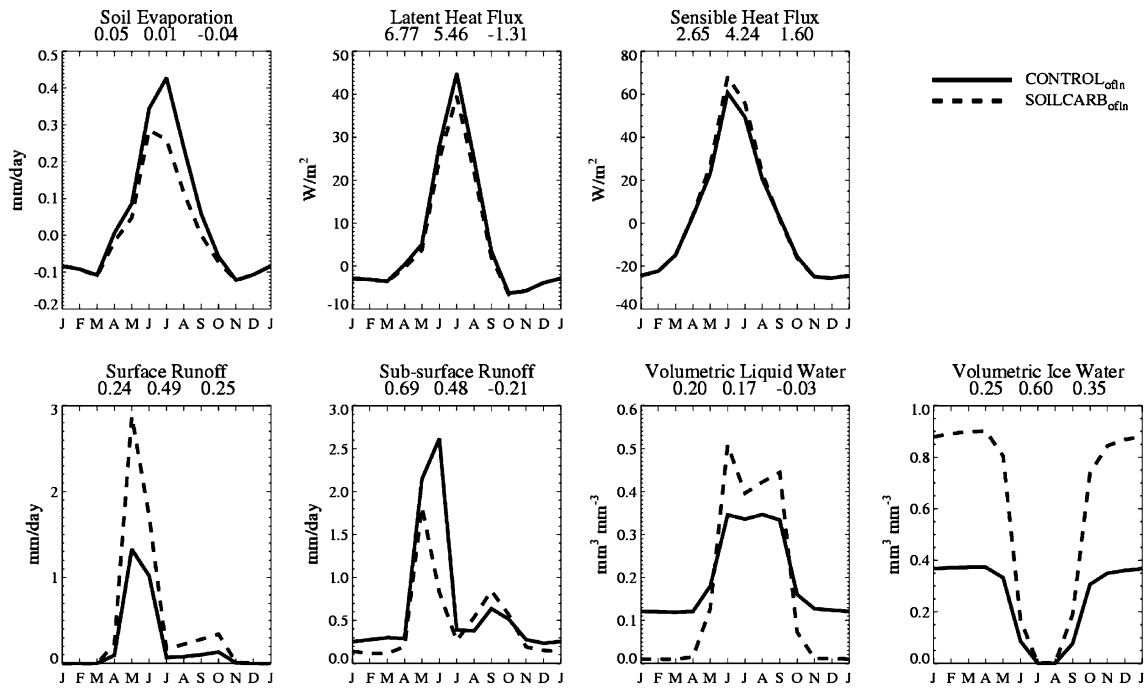


Fig. 9 Climatological annual cycle plots of selected surface energy and hydrologic cycle fields averaged over same north-west Siberia region shown in Figs. 6 and 7. Volumetric liquid and ice water are for

soil layer one. Values in *title bar* are annual mean quantities for (from left to right) CONTROL_{offn}, SOILCARB_{offn}, and SOILCARB_{offn}–CONTROL_{offn}

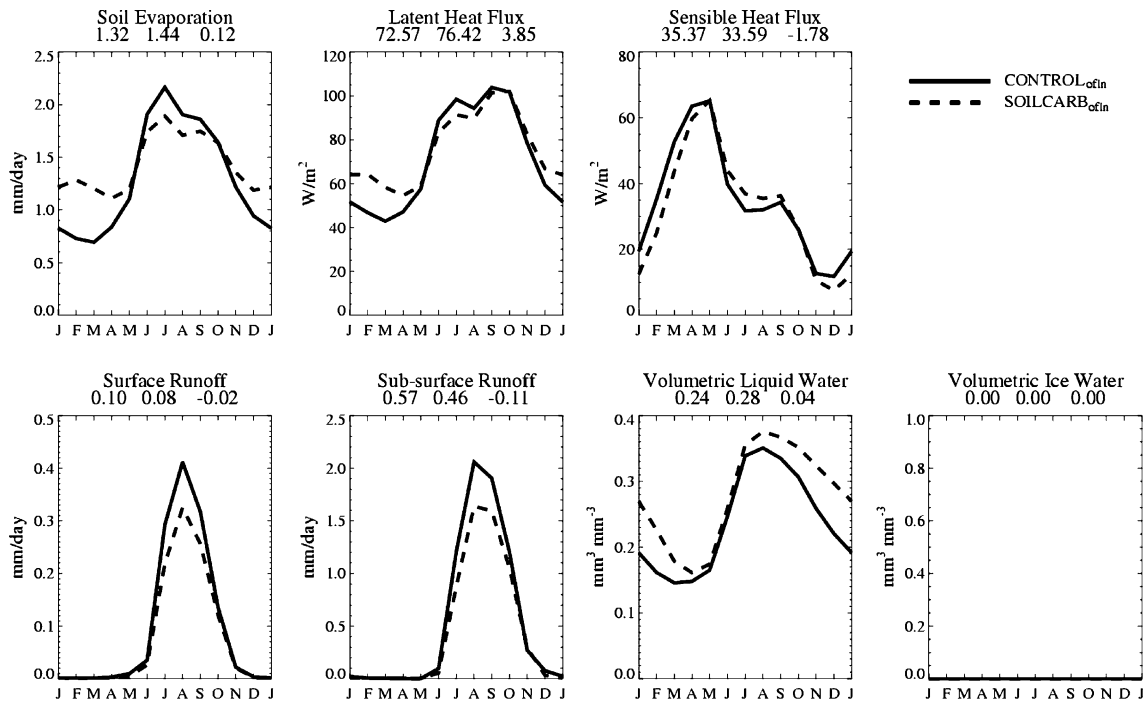


Fig. 10 Same as Fig. 9 except for Indian subcontinent (10°–25°N, 70°–90°E). Area mean soil carbon content = 14 kg m⁻²

parameters for moss and lichen are similar to those for peat and could be applied in a similar way to that done here for soil carbon, with moss thermal and hydrologic parameters

specified for the uppermost soil layers for locations with extensive moss cover. However, moss also transpires and respire (and is important in the carbon cycle) and therefore

Fig. 11 Same as Fig. 4, scatter diagram showing soil carbon content versus change in simulated column mean soil temperature, except for CAM3–CLM3 SOILCARB minus CONTROL. Only data for grid points north of 45°N are plotted

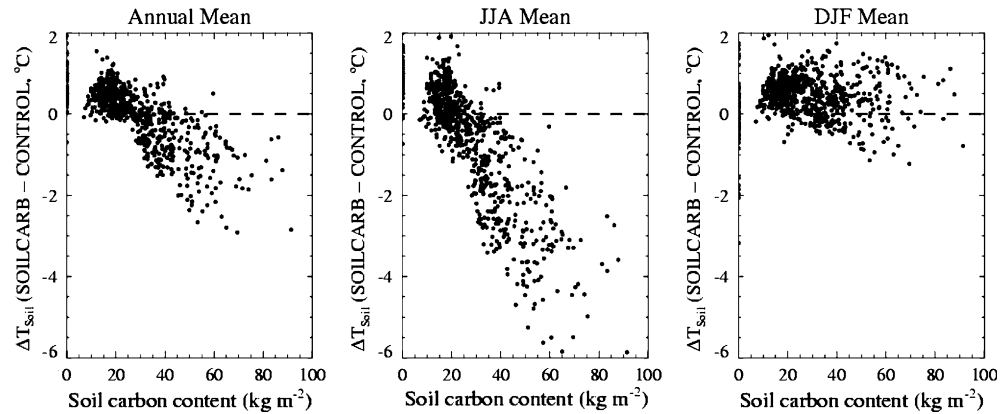
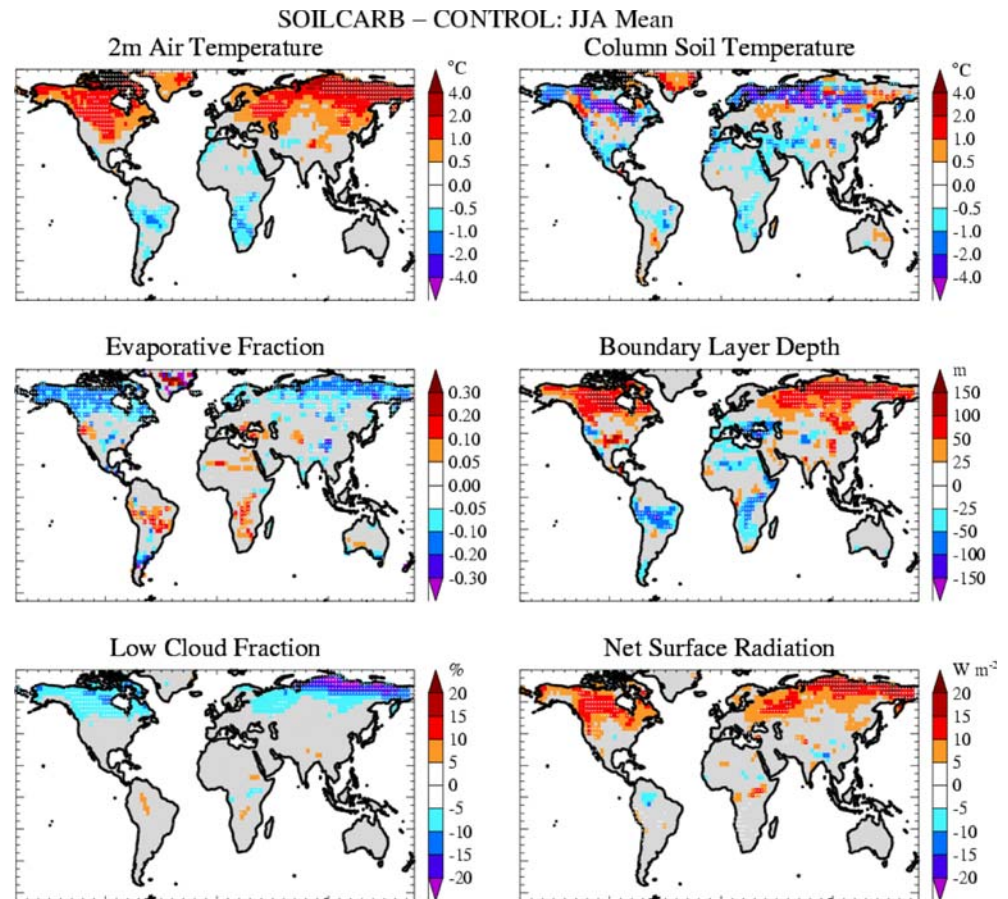


Fig. 12 Global maps of differences in JJA means between SOILCARB and CONTROL experiments for selected diagnostics. *Stippling* indicates that local difference passes the Student's *t* test for statistical significance at 95% confidence level. Evaporative fraction is latent heat flux over latent heat flux plus sensible heat flux



needs to be treated as a plant functional type, which complicates its inclusion into a GCM land-surface scheme. Another requirement, before moss and lichen can be incorporated into a global land-surface model, is a dataset detailing the distribution of moss and lichen around the world. A final point is that as we move towards higher and higher resolutions and consider incorporating a true moss/peatland/wetland ecosystem into the model, the parameterization described here may need to be revised to something more like the layered solution proposed in the Letts et al. (2000) study, at least for locations with thick peat

layers characterized by more decomposed peat typically found deeper in the soil column. These peatland ecosystems cover only about 6% of the world's surface but contain up to a third of the world's soil carbon.

5 Summary

The impact of soil organic matter on the thermal and hydraulic properties of soil is not typically accounted for in land surface schemes used in global climate models despite

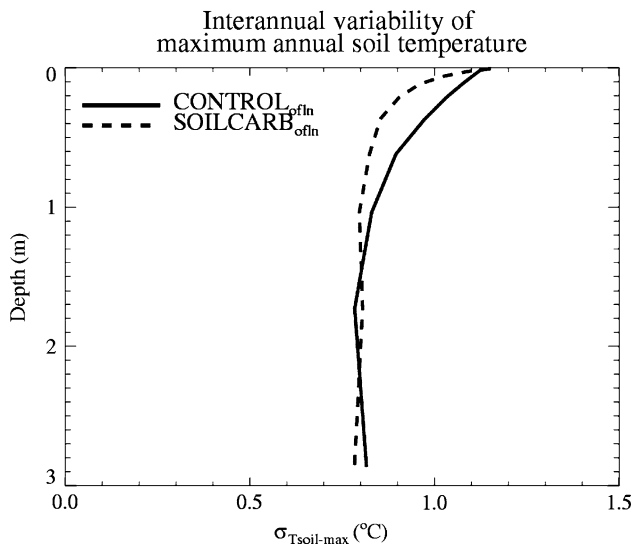


Fig. 13 Interannual variability of maximum annual soil temperature with depth for CONTROL_{offn} and SOILCARB_{offn} experiments. Variability represented as standard deviation of 20 years of model data. Lines represent average for all grid points poleward of 50°N with soil carbon contents greater than 20 kg m⁻² and where the max soil temperature variations are not damped by the phase change process in either simulation (193 grid points)

its strong influence on these properties. In the high-latitudes, where soil carbon content is high, this omission is not a valid simplification. Global data on soil carbon content, obtained through the Global Soil Data Task, has been used to derive a new geographically distributed, profiled CLM surface dataset of soil carbon density. New parameterizations that incorporate the impact of soil carbon on soil thermal and hydraulic properties are described. Analysis of offline and coupled simulations with and without the soil carbon dataset and parameterizations yield the following conclusions regarding the sensitivity of soil temperature, soil moisture, the surface energy balance, and climate to the inclusion of soil organic matter.

- Cooler annual mean soil temperatures are set up by ground heat flux differences during the first 5–10 years of the simulation. The amplitude of the cooling depends on column soil carbon content.
- The impact on soil temperature is not constant across the annual cycle. Strongly colder soil temperatures are seen in summer while somewhat smaller impacts are found during winter with some locations exhibiting weak warming and others weak cooling.
- Increases in column soil water content and saturation levels are seen, except at the surface where efficient transport of water through the organic layer results in highly variable and lower mean surface layer saturation levels and a corresponding decrease in soil evaporation and latent heat flux.

- Shifts in surface energy balance (lower latent heat flux, higher sensible heat flux) contribute to changes in high latitude climate. Summer surface air temperatures are warmer in most high-latitude locations in response to deeper boundary layers and associated reduced low cloud fraction and enhanced incident solar radiation. This amplitude of the surface air–temperature impact is likely to be model-dependent.
- The insulative properties of organic soil coupled with higher heat capacity of the wetter upper soil column combine to mildly interannual variability of soil temperature and ground heat flux, suggesting that organic soil may provide only weak protection of permafrost from transient surface warming.

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