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Importance of soil thermal regime in terrestrial ecosystem carbon dynamics in the circumpolar north

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ABSTRACT

In the circumpolar north (45–90°N), permafrost plays an important role in vegetation and carbon (C) dynamics. Permafrost thawing has been accelerated by the warming climate and exerts a positive feedback to climate through increasing soil C release to the atmosphere. To evaluate the influence of permafrost on C dynamics, changes in soil temperature profiles should be considered in global C models. This study incorporates a sophisticated soil thermal model (STM) into a dynamic global vegetation model (LPJ-DGVM) to improve simulations of changes in soil temperature profiles from the ground surface to 3 m depth, and its impacts on C pools and fluxes during the 20th and 21st centuries. With cooler simulated soil temperatures during the summer, LPJ-STM estimates ~0.4 Pg C yr^{-1} lower present-day heterotrophic respiration but ~0.5 Pg C yr^{-1} higher net primary production than the original LPJ model resulting in an additional 0.8 to 1.0 Pg C yr^{-1} being sequestered in circumpolar ecosystems. Under a suite of projected warming scenarios, we show that the increasing active layer thickness results in the mobilization of permafrost C, which contributes to a more rapid increase in heterotrophic respiration in LPJ-STM compared to the stand-alone LPJ model. Except under the extreme warming conditions, increases in plant production due to warming and rising CO_2, overwhelm the e nhanced ecosystem respiration so that both boreal forest and arctic tundra ecosystems remain a net C sink over the 21st century. This study highlights the importance of considering changes in the soil thermal regime when quantifying the C budget in the circumpolar north.

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1. Introduction

Permafrost is an important control on vegetation and soil carbon (C) dynamics by affecting hydrological and soil thermal conditions in northern high-latitude ecosystems (Wania et al.,2009a; Schaphoff et al., 2013), which account for a large portion of the global C stocks (Hugelius et al., 2014). Recent climate warming has caused significant thawing of the near-surface permafrost across the circumpolar region (Romanovsky et al., 2015), including Alaska (Jorgenson et al., 2006; Osterkamp, 2007), Canada (Camill, 2005), and Russia (Streletskiy et al., 2015). Projected warming over the 21st century is expected to greatly reduce the areal extent of permafrost and seasonally frozen ground (Lawrence et al., 2012). If permafrost thaws, a fraction of soil organic C (SOC) in previously frozen layers will decompose and be released as CO_2 and CH_4 (Olefeldt et al., 2012; Hayes et al., 2014; Johnston et al., 2014; Walter-Anthony et al., 2014). Furthermore, organic matter decomposition rates in unfrozen soils are sensitive to soil temperatures, which vary non-linearly across the soil column. Therefore, careful consideration of soil thermal regime changes (i.e. soil temperatures across the soil column from surface to deep soil layers) is important when simulating the potential future C loss from soils.

In addition to soil C, vegetation C pools are also sensitive to changes in soil thermal dynamics by permafrost thaw and rates of associated biogeochemical processes (Euskirchen et al., 2006). Field studies have indicated that permafrost thaw increases aboveground net primary production via increased nutrient availability (e.g., Schuur et al., 2007). While not explicitly testing the effects of thaw, other model studies have indicated a net C gain in circumpolar ecosystems because increased vegetation productivity more than compensated belowground C losses under a warming climate (Hartley et al., 2012; Koven, 2013). However, this net C gain may be optimistic as the effects of water stress and disturbances (e.g. insect infestations, wildfires) on permafrost-

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region biomass are not adequately incorporated in current models (Abbott et al., 2016). Thus, large uncertainties currently exist regarding the magnitude and timing of this permafrost-C feedback to the climate system (Schuur et al., 2013), due in part to the complexity of ecosystem C processes in areas of degrading permafrost (e.g., thermokarst, thermal erosion) and their heterogeneity across regions.

To date, a number of studies have used process-based land surface models to estimate permafrost-C feedbacks under projected thaw scenarios (Koven et al., 2011; Harden et al., 2012; MacDougall et al., 2012; Schneider von Deimling et al., 2012). Soil thermal dynamics within these models vary in their complexity, from calculations of the cumulative active layer thickness distribution (e.g., Harden et al., 2012) to more sophisticated parameterizations of soil physics (e.g., MacDougall et al., 2012). Thaw dynamics are typically considered in a top-down one-dimensional manner, and heat transfer via conduction is the primary mode considered. However, the importance of soil water in non-conductive heat transfer (e.g., latent heat exchange, convection) has long been recognized as an important control on soil thermal dynamics (e.g., Romanovsky and Osterkamp, 2000). In this regard, our recent modeling studies that incorporate interactions between heat and water transport (Jiang et al., 2012a) have shown improvements in simulating the soil thermal regime changes in both tundra and boreal forest ecosystems in the northern high latitudes (Jiang et al., 2015).

To provide a better quantification of the ecosystem C budget in northern high latitudes and how this budget may change in the future, we use the Lund-Potsdam-Jena Dynamic Global Vegetation Model (LPJ-DGVM, Sitch et al., 2003; Gerten et al., 2004) coupled with a sophisticated soil thermal model (STM) by Jiang et al. (2012a) to conduct a set of simulations for both historical and future periods. The standard version of LPJ-DGVM has been widely used to simulate the global C budget and its response to climate change (e.g., Sitch et al., 2008; Jiang et al., 2012b). However, the model has used simplified soil temperature parameterizations for high-latitude regions (Sitch et al., 2003). Although later studies by Wania et al. (2009a, b) have taken steps to improve soil thermal and hydrologic parameterizations for permafrost conditions in northern peatlands, soil temperature and water dynamics are still modeled separately. As a consequence, the modeled active layer thickness exhibits poor agreement with observations (Wania et al., 2009b). Moreover, the effects of the vertical soil C distribution on decomposition have not been considered. In an extended LPJ, which includes managed land (LPJmL), Schaphoff et al. (2013) have recently coupled the interactions of soil water and heat transport, and considered vertically differentiated soil C stocks based on an organized soil C dataset in Jobbagy and Jackson (2000). However, the discrete vertical differentiation of the soil temperature profile in LPJmL is relatively coarse and the vertical distribution of SOC down to 3 m is highly concentrated in the uppermost soil layers (e.g., 0–20 cm) and does not consider long-term SOC accumulation in deep soil layers (Hugelius et al., 2014).

In this study, we integrate STM into LPJ to improve simulations of soil temperature dynamics from the ground surface to a depth of three meters, and the consequent impact on soil organic C stabilization and release across the northern permafrost region. Because the newly coupled version of the model, LPJ-STM, considers a fine vertical differentiation of soil temperatures, interactions of soil water with heat transport, and a vertical distribution of SOC with more C in deeper soil layers, we expect LPJ-STM to provide a more accurate quantification of the C budget for historical periods and to improve projections of carbon dynamics under future scenarios. To examine if LPJ-STM improved estimates of soil temperatures and associated carbon dynamics over LPJ, we compare estimates of both models to site observations of soil temperatures and net ecosystem production (NEP). In addition, simulated atmospheric carbon dioxide concentrations determined by using the model NEP estimates combined with an atmospheric transport model are compared to atmospheric flask measurements. To examine how improvements in simulated soil thermal regime affect the estimates of the contemporary C budget of northern high latitude ecosystems and its projection into the 21st century, we compare estimates of net primary production (NPP), heterotrophic respiration (R_H), and net ecosystem production (NEP) as well as vegetation and soil C stocks between LPJ and LPJ-STM.

2. Methods

2.1. Model description

The LPJ model simulates large-scale vegetation structure and land-atmosphere C and water fluxes in a modular framework (Sitch et al., 2003; Gerten et al., 2004). In the standard LPJ soil model, two soil layers have a fixed depth (i.e. 0.5 m and 1.0 m). The vertical distribution of soil C within these two soil layers is not explicitly considered. For permafrost thermal dynamics, LPJ calculates soil temperature at a depth of 25 cm in a very simplistic way based on the surface air temperature seasonal cycle with a dampened oscillation about a common mean and with a temporal lag (Sitch et al., 2003). The magnitude of the damping and temporal lag is related to soil texture, and the applied function assumes a sinusoidal variation in seasonal temperatures (Campbell and Norman, 1998). In this study, we replace the existing soil temperature scheme in the original LPJ with STM (Jiang et al., 2012a) to produce the coupled LPJ-STM model. Meanwhile, we modified the hydrology scheme following Wania et al. (2009b) and the representation of soil carbon dynamics following Schaphoff et al. (2013) in the LPJ-STM model.

2.1.1. Soil thermal dynamics in LPJ-STM

While the original LPJ only includes a temperature calculation at a depth of 25 cm, LPJ-STM has a vertical differentiation of the soil temperature profile and provides a more complete depiction of soil thermal dynamics, including estimates of active layer thickness (ALT). Here, we estimate ALT using the 0 °C isotherm line as in Wania et al. (2009b). In LPJ-STM, the top three meters of soil is split into six layers with thicknesses of 10, 10, 10, 20, 50, and 200 cm (Fig. 1) that differ with respect to physical and hydraulic properties of boreal forest and tundra sites (Jiang et al., 2012a). These soil layer thicknesses are finer than the five soil layers used by Schaphoff et al. (2013) for the top 3 m: 20, 30, 50, 100, and 100 cm. In addition, unlike Schaphoff et al. (2013), LPJ-STM divides each of the soil layers into a number of depth-step increments to calculate soil temperatures. From the first to the sixth layer, the number of depth-step increments is respectively 10, 10, 10, 10, 20, and 40. To simulate perennially frozen ground, soil between 3 and 50 m is classified as the seventh layer with soil temperature simulated at 0.5 m depth-step increments.

Soil temperature is simulated numerically at each depth step by solving a modified Richards equation (Hansson et al., 2004; Saito et al., 2006). The governing equation for heat transport is as follows:

$$\frac{\partial C_p T}{\partial t} - L_f \rho_i \frac{\partial \theta_i}{\partial t} + L_0(T)\frac{\partial \theta_v(T)}{\partial t} = \frac{\partial}{\partial z}\left[\lambda \frac{\partial T}{\partial z}\right] - C_w \frac{\partial q_l T}{\partial z} - C_v \frac{\partial q_v T}{\partial z} - L_0(T)\frac{\partial q_v}{\partial z} - C_w S T \quad (1)$$

where C_p (J m^{-3} K^{-1}) is the volumetric heat capacity of the soil, C_w is the volumetric heat capacity of liquid water (4.18×10^6 J m^{-3} K^{-1}), C_v is the volumetric heat capacity of water vapor (1.2×10^3 J m^{-3} K^{-1}), θ_i is the volumetric ice content (%), θ_v is the volumetric unfrozen water content (%), t is time (units), z is depth (m), T is the absolute temperature (K), ρ_i is the density of ice (931 kg m^{-3}), L_0 is the volumetric latent heat of vaporization of liquid water (J m^{-3}), L_f is the latent heat of freezing (3.34×10^5 J kg^{-1}), λ is the apparent thermal conductivity of soil (J m^{-1} s^{-1} K^{-1}), q_l is the flux density of liquid water, q_v is the flux density of water vapor (m s^{-1}), and S is a sink term accounting for root water uptake (s^{-1}). While soil temperatures

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