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Representing permafrost properties in CoLM for the Qinghai–Xizang (Tibetan) Plateau

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ABSTRACT

Most land surface models (LSMs) used in climate models do not perform well in modeling the permafrost processes. Due to the complex permafrost distribution characteristics and landscapes of the Qinghai–Tibet Plateau (QTP), the LSMs simulations over QTP are even worse. In this study, we revised the permafrost scheme in the original Common Land Model (CoLM) to improve its capability of simulating permafrost processes. We adopted a new frozen soil parameterization scheme, in which maximum unfrozen water content is defined as a function of soil matric potential. In addition, we extended the model's bottom to a depth below that without annual variations in temperature and replaced the zero-flux lower boundary condition with a constant geothermal heat flux based on literature and temperature gradient measurements in a 34.5-m-deep borehole. What's more, we revised the original snow cover fraction parameterization scheme of CoLM according to the special snow cover distribution characteristics over QTP. We calibrated and validated the modified model against observations from 2005 to 2008. The results indicate that the modified model produced more reasonable simulations of radiation balance components and significantly improved the simulation of soil liquid water content. It also shows an improved capability of reproducing soil temperatures from the top to the bottom of soil layers. The modified CoLM provides a useful tool for understanding and predicting the fate of permafrost in QTP under a warming climate.

1. Introduction

Permafrost, defined as ground temperature that remains at or below 0 °C continuously for two or more years (Muller, 1947), is widespread in high latitudes and alpine regions, occupying approximately 24% of the land area in the Northern Hemisphere (Brown et al., 1997; Zhang et al., 1999, 2000) and 53% of the area on the Qinghai–Tibet Plateau (QTP) (Cheng, 1984). Widespread permafrost degradation due to climate warming during the last several decades has already initiated a number of feedbacks in the global climate system (Anisimov and Reneva, 2006; Guglielmin and Dramis, 1999; Isaksen et al., 2007; Lachenbruch and Marshall, 1986; Liu and Chen, 2000; Nelson et al., 2001; Osterkamp, 2007; Pavlov, 1994; Smith and Riseborough, 2002; Smith et al., 2005), and may likely have significant impacts on local hydrology, ecosystems and carbon cycle, soil biogeochemistry, and engineering infrastructure (Cheng and Wu, 2007; Hinzman et al., 2005; Lemke et al., 2007). Therefore, more and more attention has been paid to understanding, assessing, and predicting the changes of permafrost over the QTP regions in recent years (Cheng and Wu, 2007; IPCC, 2007; Lemke et al., 2007; Wu and Zhang, 2008; Wu et al., 2009; Zhao et al., 2004, 2010).

Process-based modeling is an important tool for studying the interactions between the atmosphere, vegetation, and soil processes (Sridhar et al., 2002). Processes occurred in permafrost are unique and complex because ice, liquid water, and water vapor may coexist in the frozen ground and lead to complex interactions in the active layer during freezing and thawing phases (Romanovsky and Osterkamp, 2000; Zhao et al., 2000). The existence of ice not only changes the physical characteristics of the soil, but also affects the migration and distribution of liquid water in the soil surface and subsurface (Jame and Norum, 1980; Li and Koike, 2003). Numerous studies of permafrost modeling were focused on evaluating permafrost responses to climate change (Anisimov and Nelson, 1996, 1997; Goodrich, 1982; Hinzman et al., 2005; Malevsky-Malevich et al., 2001; Nelson et al., 2002; Zhang et al., 2003, 2008; Zhou et al., 2008). However, the importance of permafrost property parameterization in these models was not given adequate attention until 2000s.

Existence of unfrozen water at sub-zero temperatures can change the thermal and hydrological properties of the soil (Kane and Stein, 1983). Neglecting unfrozen moisture in land surface models would induce great uncertainty in the estimation of soil moisture and temperature (Koren et al., 1999; Pitman et al., 1999; Viterbo et al., 1999). In recent years, some frozen soil parameterization schemes

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have been developed for land surface models (Koren et al., 1999; Li and Koike, 2003). Li and Koike (2003) assorted the schemes into three categories and pointed out that all these schemes have their own advantages in explicitly incorporating phase change in heat and moisture transfers. The parameterization in which the liquid water content in frozen soil is defined as a function of soil matric potential (Clapp and Hornberger, 1978; Fuchs et al., 1978) has been proved reasonable in many recent land surface models (Cherkauer and Lettenmaier, 1999; Flerchinger and Saxton, 1989; Luo et al., 2009; Niu and Yang, 2006; Zhang and Lu, 2002).

However, most of the modeling studies were conducted in high-latitude regions (Bowling et al., 2003; Henderson-Sellers et al., 1995; Nijssen et al., 2003), and few were carried out in the QTP region. In such a low latitude alpine region, the characteristics of solar radiation, air temperature, wind speed, and snow cover are different from those in high-latitudes. The permafrost of the plateau is much thinner and more sensitive to changes in climate and surface conditions (Cheng, 1998), and the frozen period of the surface active layer is much shorter (Wu, 2004; Ye and Gao, 1979). Therefore, the permafrost schemes in land surface models need to be further developed for use over the plateau. In addition, few permafrost schemes have been developed for the QTP due to the paucity of long-term, high-quality, continuous observational data, which are also essential for the study of permafrost degradation over the plateau and assessment of its impact on climate change.

This paper describes the representation of several permafrost property schemes to CoLM to improve its capability of simulating soil temperature and liquid water content, and the partitioning of radiative energy into sensible and latent heat fluxes in the permafrost regions of QTP. We calibrated and validated the modified model over QTP against the observational data collected at the Tanggula site from 2005 to 2008, and discussed the importance and necessity of considering frozen soil parameterization and geothermal flux in land surface models.

2. Modified common land model

2.1. Overview

The Common Land Model (CoLM) is a state-of-the-art land surface model developed for use in climate studies (Dai et al., 2001; Zeng et al., 2002). Dai et al.(2003) provided a detailed description of the model, which has 10 unevenly spaced vertical soil layers extending to a depth of 3.43 m and up to five snow layers (depending on the total snow depth). A two-big-leaf model for canopy temperature, photosynthesis and stomatal resistance was coupled by Dai et al. (2004). Coupling CoLM with the NCAR Community Climate Model (CCM3), Zeng et al. (2002) compared the simulation results with those resulting from the NCAR LSM (Bonan, 1996) and BATS, and found that CoLM improved the simulation of land surface climate. Steiner et al. (2005) coupled CoLM to a regional climate model (RegCM) and obtained significant improvements in their simulations.

Xiao et al. (2011) investigated the applicability of CoLM to permafrost regions of QTP by comparing the simulations against observations. It showed that CoLM could successfully simulate water and heat exchanges between the land and the atmosphere but needs further developments in parameterizing the permafrost processes. Also, extension of the simulation depth to a greater depth is necessary for modeling the permafrost on QTP.

2.2. Frozen soil parameterization schemes

2.2.1. Original frozen soil considerations in CoLM

In the original CoLM, the soil temperature T_i^{n+1}, firstly calculated without consideration of phase change, are then evaluated to determine if phase change will take place at soil layer i (Dai et al., 2003):

$$\begin{array}{lll} T_\mathrm{i}^{n+1} > T_\mathrm{f} \text{ and } w_\mathrm{ice,i} > 0 & & \text{melting} \\ T_\mathrm{i}^{n+1} < T_\mathrm{f} \text{ and } w_\mathrm{liq,i} > 0 & & \text{freezing} \end{array} \tag{1}$$

where $w_\mathrm{ice,i}$ and $w_\mathrm{liq,i}$ are the mass of ice and liquid water content (kg m^{-2}) in each soil layer, respectively, and T_f is the freezing point of water (T_f=273.16 K). The excess or deficit of energy H_i, needed to change T_i to the freezing point (T_f), is determined as follows

$$H_\mathrm{i} = \alpha\left(F_\mathrm{i}^n - F_\mathrm{i-1}^n\right) + (1-\alpha)\left(F_\mathrm{i}^{n+1} - F_\mathrm{i-1}^{n+1}\right) - \frac{c_\mathrm{i}\Delta z_\mathrm{i}}{\Delta t}\left(T_\mathrm{f} - T_\mathrm{i}^n\right) \tag{2}$$

where $F_{\mathrm{i}-1}$ and F_i are heat fluxes as positive upwards, c_i is the specific heat capacity. If the melting or freezing criteria are met (Eq. (1)) and $|H_\mathrm{m}| = \left|\frac{H_\mathrm{i}\Delta t}{L_\mathrm{f}}\right| > 0$, then the ice mass is updated as

$$w_\mathrm{ice,i}^{n+1} = \left\{ \begin{array}{ll} \max\left(w_\mathrm{ice,i}^{n+1} - H_\mathrm{m}, 0\right) & H_\mathrm{m} > 0 \\ \min\left(w_\mathrm{ice,i}^{n+1} - H_\mathrm{m}, w_\mathrm{liq,i}^n + w_\mathrm{ice,i}^n\right) & H_\mathrm{m} < 0 \end{array} \right\} \tag{3}$$

where $\mathrm{L_f}$ is the latent heat of fusion (0.3336×106 J kg−1), and the liquid water mass is updated as

$$w_\mathrm{liq,i}^{n+1} = \max\left(w_\mathrm{liq,i}^n + w_\mathrm{ice,i}^n - w_\mathrm{ice,i}^{n+1}, 0\right). \tag{4}$$

2.2.2. Modified frozen soil parameterization scheme

The original CoLM, neglecting liquid water content in the frozen soil, would take much more water into the process of phase change during soil freezing/thawing periods. Thus there would be a significant error in estimation of soil temperature and moisture. We modified the frozen soil parameterization scheme in CoLM using the concept of supercooled soil water from Niu and Yang (2006). The supercooled soil water is the liquid water that coexists with ice over a wide range of temperatures below the freezing point and is implemented through a freezing point depression equation:

$$\psi = \left\{ \begin{array}{c} \psi_\mathrm{sat}\left(\frac{\theta\mathrm{liq}}{\theta_\mathrm{sat}}\right)^{-b} \\ \frac{10^3 L_\mathrm{f}(T - T_\mathrm{f})}{gT} \end{array} \right\} \tag{5}$$

where ψ is soil matric potential (mm), ψ_sat is saturated soil matric potential (mm) depending on soil texture, b is the Clapp–Hornberger parameter, and g is the acceleration of gravity (9.80616 m s^{-2}).

We used the following equation to calculate the maximum unfrozen liquid water mass,

$$\mathrm{W_{liqmax}} : \mathrm{w_{liq,max,i}} = \rho_\mathrm{water}\Delta z_\mathrm{i}\theta_\mathrm{sat,i}\left\{\frac{10^3 L_\mathrm{f}(T_\mathrm{i} - T_\mathrm{f})}{gT_\mathrm{i}\psi_\mathrm{sat,i}}\right\}^{-1/b}, T < T_\mathrm{frz} \tag{6}$$

where ρ_water is the density of water (103 kg m^{-3}). Then the melting/freezing criteria for phase change Eq. (1) are changed into:

$$\begin{array}{lll} T_\mathrm{i}^{n+1} > T_\mathrm{f} \text{ and } w_{ice,i} > 0 & & \text{melting} \\ T_\mathrm{i}^{n+1} < T_\mathrm{f} \text{ and } w_\mathrm{liq,i} > w_\mathrm{liq,max,i} & & \text{freezing.} \end{array} \tag{7}$$