Frozen soil parameterization in SiB2 and its validation with GAME-Tibet observations

Xin Li,*, Toshio Koike

*Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, 260 Donggang West Road, Lanzhou 730000, PR China

bDepartment of Civil Engineering, the University of Tokyo, Tokyo, Japan

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Abstract

This paper presents a new frozen soil parameterization in the land surface scheme (LSS). A modified approximation Stefan solution is incorporated in the framework of the land surface model-SiB2 to calculate the frost/thaw depth over time and to estimate the soil moisture and temperature profiles during the freeze/thaw cycle. The structure of the soil model in Simple Biosphere Model 2 (SiB2) is kept, but the governing equations of water balance and surface heat balance are modified to account for soil freezing/thawing. The model is calibrated and validated using the GEWEX Asian Monsoon Experiment (GAME)-Tibet observations at MS3608 site, Tibetan Plateau. First, the original SiB2 without frozen soil parameterization is calibrated using the observations in summer; an optimized parameter set for the short grassland in the Tibetan Plateau was obtained. Then, the modified SiB2 with frozen soil parameterization is validated using the soil moisture and temperature observations in winter. The results show that the new model estimates the frost depth precisely with an error less than 9% and predicts the soil temperature reasonably and phase transition time realistically. It also improves the soil moisture estimation in the surface layer and the root zone significantly. The mean absolute errors of simulated soil moisture using the new model are 0.020 and 0.013 in the surface layer and root zone, respectively, far below than those without using the frozen soil parameterization.

Keywords: Frozen soil parameterization; Land surface model; Tibetan Plateau

1. Introduction

Frozen soil plays an important role in the terrestrial portion of the hydrological cycle because the water phase transition results in significant change of liquid soil moisture content and hence impacts on soil evaporation and canopy transpiration. In addition, when soil is frozen, its hydraulic conductivity dramatically decreases, making frozen soil an impermeable layer of water flow. Frozen soil also has a great impact on heat balance of land surface due to fusion heat released (consumed) by a freezing (thawing) process. Thermal properties change simultaneously with this process of heat transfer. When soil is frozen, its thermal conductivity increases because the thermal conductivity of ice is about four times that of water.

* Corresponding author. Tel.: +86-931-8275593; fax: +86-931-8273894.
and its heat capacity decreases because the heat capacity of water is the largest among all materials. Any decrease of liquid water content in soil will result in a decreased heat capacity.

Besides its specific hydraulic and thermal characteristics, frozen soil covers a large proportion of global land area. Permafrost (permanently frozen soil) underlies approximately 24% of Northern Hemisphere land area (Zhang et al., 1999). Seasonal frozen soil occupies about 30% of global land area (Williams and Smith, 1989). Therefore, frozen soil is obviously a dominant factor in land surface processes because of its effects on thermal and hydrological regimes and its large area extent. To incorporate frozen/thaw process in a land surface scheme (LSS) is very important and essential. Lack of frozen soil parameterization in land surface models will result in great uncertainty of soil moisture simulation (Pitman et al., 1999), large diurnal temperature cycles, and the resulting soil cooling in winter (Viterbo et al., 1999).

However, the frozen soil parameterization in land surface schemes is not given sufficient attention until recent years. Most of the literature began to appear in later 1990s. These parameterizations are different in detail, but summed up, they can be assorted into three categories. The first kind only account for the hydraulic and thermal properties of frozen soil, e.g. SSiB and SiB2 use a linear function to calculate the decreasing hydraulic conductivity when soil temperature is below 0 °C (Sellers et al., 1996a; Xue et al., 1996); BATS assumes that soil moisture freezes uniformly between 0 and −4 °C and the thermal diffusivity of soil is limited to 1.4 × 10^{-6} m^2 s^{-1} (Dickinson et al., 1993).

The second kind calculates ice production rates based on the amount of heat consumed (released) and left to be consumed (released) in a unit volume of soil. The frozen soil parameterization in the Best Approximation of Surface Exchanges (BASE) (Slater et al., 1998) and the CCSR/NIES GCM (Takata and Kimoto, 2000) belong to this category. The origin of these methods can be found in a coupled heat and moisture flow model, FROSTB (Berg et al., 1980; Shoop and Bigl, 1997). In this parameterization, the amount of heat for ice production is calculated by

$$
\Delta Q = \min(C_s(T_s - T_f), L_f(\theta_t - \theta_n)) \quad (1)
$$

and the ice production rate is then calculated by

$$
\frac{\rho_l}{\rho_t} \frac{\partial \theta_t}{\partial t} = \frac{1}{L_f} \frac{\Delta Q}{\Delta t} \quad (2)
$$

where the function \( \min \) is to get the minimum value of a set of elements; \( C_s \) is the volumetric heat capacity of soil (J m\(^{-3}\) K\(^{-1}\)); \( T_s \) is the soil temperature (K); \( T_f \) is the freezing point of water (K); \( L_f \) is the latent heat of fusion (J kg\(^{-1}\)); \( \theta_t \) is the liquid water content (m\(^3\) m\(^{-3}\)); \( \theta_i \) is the ice content (m\(^3\) m\(^{-3}\)); \( \theta_n \) is the minimum volumetric unfrozen water content (m\(^3\) m\(^{-3}\)); \( \rho_l \) is the density of ice (kg m\(^{-3}\)); \( \rho_t \) is the density of liquid water (kg m\(^{-3}\)); \( \Delta t \) is the time step.

The third kind of frozen soil parameterization uses soil matric potential to define maximum liquid water content when soil temperature is below the freezing point. The frozen soil parameterization in the Meso-scale Analysis and Prediction System (MAPS) (Smirnova et al., 2000), the Variable Infiltration Capacity (VIC) model (Cherkauer and Lettenmaier, 1999), and the mesoscaleEta model (Koren et al., 1999) belong to this category. In this parameterization, liquid water content is a function of soil matric potential, and any additional water is ice (Fuchs et al., 1978; Flerchinger and Saxton, 1989).

$$
\theta_t = \theta_s \left( \frac{\psi}{\psi_e} \right)^{-1/b} = \theta_s \left[ \frac{L_f(T_s - T_f)^{1/b}}{gT_s\psi_e} \right]^{-1/b} \quad (3)
$$

where \( \theta_s \) is the saturated water content (m\(^3\) m\(^{-3}\)); \( \psi \) is the soil matric potential (m); \( \psi_e \) is the air entry potential (m); \( b \) is a pore-size distribution index; \( g \) is the acceleration of gravity (m s\(^{-2}\)).

In categories 2 and 3, the phase transition is modeled by an isothermal approach, with an assumption that the phase transition occurs in the middle of each soil layer. This assumption, when applied to very thick soil layers (in most land surface schemes, this is the case), will result in delayed or rapid freezing/thawing. Therefore, it is natural to expect that the predictability of frost/thaw depth can improve the frozen soil parameterization because the soil temperature profile within a soil layer can be estimated according to the position of freezing front. The Stefan solution, which is frequently used by permafrost scientists to predict frost depth and simulate heat transfer with water phase transition in frozen soil, is capable to implement this objective. The classic sol-
ution of Stefan problem can be documented as early as 1890s (Stefan, 1890). Recent literatures of modified solutions can be classified as numerical solutions (Li et al., 1996; Zhang and Stamnes, 1998), analytic solutions (Romanovsky and Osterkmap, 1997), and approximation solutions (Fox, 1992; Hinkel and Nicholas, 1995). All these solutions have their advantage in explicitly incorporating phase change in heat and moisture transfer and in predicting frost/thaw depth over time, but the numeric solutions usually need very fine resolution of soil layers (from < 1 cm to a few centimeters). This can explain why only a few efforts have been made to couple Stefan solution with land surface or large-scale water balance models (Fox, 1992), which usually employ a soil column of no more than three layers below surface. Too many soil layers and a numerical solution of coupled heat and moisture transfer in a land surface model result in large computation cost and incompatibility with the structure of current land surface schemes.

In this paper, we attempt to find a trade-off between the accuracy of Stefan solution in predicting frost/thaw depth and the simple structure of land surface models in manipulating soil column. A new frozen soil parameterization is presented, in which a modified approximation Stefan solution is incorporated in the framework of the land surface model-SiB2. We introduce the detailed frozen soil parameterization scheme in Section 2, calibrate and validate the offline-uncoupled model using the GAME-Tibet observations in Section 3. In Section 4, we give some concluding remarks.

2. Frozen soil parameterization

SiB2 is a revised version of the Simple Biosphere Model (SiB) developed by Sellers and Mintz (1986) and Sellers et al. (1996a). It is chosen to work as the framework model for the new sub-layered frozen soil parameterization because of its improved performance in simulating continental hydrometeorology (Sato et al., 1989). The new parameterization is focused on the heat and moisture transfer with phase transition so that consideration is mainly given to the soil and the surface hydrological submodels of SiB2.

SiB2 includes three soil layers (Fig. 1): a surface soil layer of a few centimeters, which acts as a significant source of direct evaporation when moist; a root zone, which is the supplier of soil moisture to the roots and accounts for transpiration; and a deep soil layer, which acts as a source for hydrological base flow and upward recharge of the root zone. This structure of the soil model in SiB2 is kept in the new frozen soil parameterization, but the three governing equations of water balance and the governing equation of surface heat balance are modified to involve the soil freezing/thawing process.

2.1. Modifications on the governing equations of water balance

Assuming that Darcy’s law of liquid water flow can be applied analogously to the frozen soil and that the vertical ice flow can be negligible, the equations of water balances that incorporate ice content in the three soil layers can be expressed by

\[
\frac{\partial \theta_{1,1}}{\partial t} + \frac{\rho_1}{\rho_l} \frac{\partial \theta_{1,1}}{\partial t} = \frac{1}{D_1} \left[ (D_c + D_d - R_{01}) - Q_{1,2} - \frac{1}{\rho_1} E_{gs} \right]
\]

(4)

\[
\frac{\partial \theta_{1,2}}{\partial t} + \frac{\rho_1}{\rho_l} \frac{\partial \theta_{1,2}}{\partial t} = \frac{1}{D_2} \left[ Q_{1,2} - Q_{2,3} - \frac{1}{\rho_1} E_{cl} \right]
\]

(5)

\[
\frac{\partial \theta_{1,3}}{\partial t} + \frac{\rho_1}{\rho_l} \frac{\partial \theta_{1,3}}{\partial t} = \frac{1}{D_3} \left[ Q_{2,3} - Q_3 \right]
\]

(6)

where \( \theta_{1,j} \) and \( \theta_{1,j} \) (\( j = 1,2,3 \)) are the liquid water content and the ice content (m\(^3\) m\(^{-3}\)) of each soil.
layer, respectively; \( D_j \) \((j = 1,2,3)\) is the thickness of each soil layer (m); \( Q_{j\,j+1} \) is the water flow between \( j \) and \( j+1 \) layers (m s\(^{-1}\)); \( Q_3 \) is the gravitational drainage from recharge soil moisture store (m s\(^{-1}\)); \( D_c \) is the canopy drainage rate (m s\(^{-1}\)); \( D_t \) is the canopy throughfall rate (m s\(^{-1}\)); \( R_{o1} \) is surface runoff rate due to excess water infiltration in the surface soil (m s\(^{-1}\)). Because of the existence of the ice in the soil, the equation of excess water that contribute to surface runoff is modified as

\[
\text{excess} = \theta_{l,1} + \theta_{l,1} - \theta_s \tag{7}
\]

\( E_{gs} \) and \( E_{ct} \) are the evaporation rate from soil surface layer (m s\(^{-1}\)) and the canopy transpiration rate though the stomata (m s\(^{-1}\)), respectively.

The unit of liquid and frozen soil moisture is defined as volumetric water content in the new parameterization instead of soil moisture wetness in SiB2 because the modification can achieve a simplicity in the calculation of moisture transfer with phase transition. The relationship between the original \( W \) and the new prognostic variables \( (\theta_l \) and \( \theta_t ) \) of soil moisture is

\[
W = \frac{\theta}{\theta_s} \tag{8}
\]

where \( \theta \) is the total volumetric water content of each soil layer (m\(^3\) m\(^{-3}\))

\[
\theta = \theta_l + \theta_t \frac{\rho_i}{\rho_l} \tag{9}
\]

Experiments showed that the unfrozen water content is approximately a simple power function of soil temperature (Andersland and Anderson, 1978; Li and Cheng, 1995; Nakano et al., 1982; Romanovsky and Osterkamp, 2000)

\[
\theta_l = a(T_f - T_s)^b \tag{10}
\]

where \( a \) and \( b \) are two empirical coefficients associated with soil type.

The changing rate of liquid soil moisture (or ice) to soil temperature is thus expressed by the derivative of the Eq. (10)

\[
\frac{\partial \theta_l}{\partial T_s} = - \frac{\rho_l}{\rho_i} \frac{\partial \theta_l}{\partial T_s} = (-ab)(T_f - T_s)^{b-1} \tag{11}
\]

The liquid other than the total water content is adopted to calculate the water flow between soil layers. For hydraulic conductivity, the progressive reduction factor defined in SiB2 is kept in the new frozen soil parameterization, but instead of using a simple estimation of soil temperature at each soil layer, the soil temperature solution from the thermal model described in the next section is used to calculate the factor \( f_{ice} \)

\[
f_{ice} = \text{max}[0, (T_{s,j} - (T_f - 10)/10)] \quad j = 1, 2, 3 \tag{12}
\]

where the function max is to get the maximum value of a set of elements.

2.2. Thermal model

2.2.1. Surface temperature

The force-restore model (Deardorff, 1977) of the heat balance in the soil surface is kept the same, but the effective heat capacity of surface soil is modified to account for the latent heat of fusion in the surface layer

\[
C_g \frac{\partial T_g}{\partial t} = R_{ng} - H_g - \lambda E_g - \frac{2\pi C_d}{\tau_d} (T_g - T_d) - \bar{\varphi}_{gs} \tag{13}
\]

where \( T_g \) is the surface temperature (K); \( T_d \) is the temperature of deep soil (K); \( R_{ng} \) is the absorbed net radiation (W m\(^{-2}\)); \( H_g \) is the sensible heat flux (W m\(^{-2}\)); \( E_g \) is the evapotranspiration rate (kg m\(^{-2}\) s\(^{-1}\)); \( C_g \) is the effective heat capacity (J m\(^{-2}\) K\(^{-1}\)) of surface soil layer; \( C_d \) is the effective heat capacity (J m\(^{-2}\) K\(^{-1}\)) of snow free soil; \( \lambda \) is the latent heat of vaporization (J kg\(^{-1}\)); \( \bar{\varphi}_{gs} \) is the energy transfer due to phase changes in snow on ground (W m\(^{-2}\)).

The new equation of \( C_g \) is

\[
C_g = d_s \left[ C_s + \rho_l L_t \left( \frac{\partial \theta_{l,1}}{\partial T_g} \right) \right] + \text{min}[0.05, (M_{gs} + M_{gw})]C_w \tag{14}
\]

where \( d_s \) is the effective depth that feels the diurnal variation of temperature (Stull, 1988); \( C_s \) is the volumetric heat capacity of soil (J m\(^{-3}\) K\(^{-1}\)); the term \( \rho_l L_t (\partial \theta_{l,1}/\partial T_g) \) accounts for the apparent heat
capacity of soil freezing in the surface layer; \( M_{gs} \) and \( M_{gw} \) are snow stored on the canopy and ground (m), respectively, \( C_w \) is the volumetric heat capacity of water (J m\(^{-3}\) K\(^{-1}\)).

2.2.2. Heat transfer in the surface soil layer

Stefan solution assumes a linear soil temperature profile in the frozen soil column, however the soil temperature near the surface usually varies greatly, especially in some cold regions such as the Tibet Plateau, where the diurnal variation of surface temperature can be larger than 40 K. This great variation makes the estimation of soil temperature profile very unstable. Accordingly, frost/thaw depth is calculated from 4 cm below the surface in this model, where the diurnal variation is much smaller and can be assumed as a prefect periodic relationship with time. The soil temperature at 4 cm \( (T_{s,D_1}) \) is then solved by an analytic solution of the thermal transfer equation (Gao et al., 2000a; Li and Cheng, 1995)

\[
T_{s,D_1} = \bar{T}_g - gD_1 + A\exp\left(-D_1\sqrt{\frac{\omega}{V_g}}\sin(\omega t) - D_1\sqrt{\frac{\omega}{V_g}}\right) \tag{15}
\]

where \( \bar{T}_g \) is the daily mean surface temperature (K); \( g \) is the thermal gradient (K m\(^{-1}\)), calculated by the daily averaged temperature difference from the surface to the deep soil; \( A \) is the amplitude of surface temperature variation (K); \( \omega = 2\pi/\lambda \) and \( \lambda = 86,400 \) is the period; \( V_g \) is the heat diffusivity \( (\text{m}^2 \text{s}^{-1}) \) and

\[
V_g = \frac{\kappa_s}{C_s'} \tag{16}
\]

where \( C_s' \) is the apparent volumetric heat capacity \( (\text{J m}^{-3} \text{K}^{-1}) \). When soil is frozen, \( C_s' \) is calculated by the equation below to take into account the thermal effect of frozen soil in the surface layer

\[
C_s' = C_s + L_c\rho_1 \frac{\partial \theta_{l,1}}{\partial T_{s,D_1}} \tag{17}
\]

The liquid water changing rate \( \partial \theta_{l,1}/\partial T_{s,D_1} \) in the surface layer is a function of \( T_{s,D_1} \), as described in the Eq. (11). Then, by substituting Eq. (17) into Eq. (15), \( T_{s,1} \) can be resolved.

2.2.3. Frost/thaw depth and soil temperature profile

Li and Cheng (1995) used Green’s equation to solve the problem of time-cumulative heat transfer with phase transition and obtained the following equations to calculate the frost depth \( (\zeta_f) \) and thaw depth \( (\zeta_t) \) over time in a soil column.

\[
\zeta_f^2(t) = \frac{2}{L_c\rho_1}\left\{-\kappa_f \int_0^t [T_{s,D_1}(t) - T_f] dt + \int_0^{\zeta_f} zC_{s,f}[T_s(z,t) - T_f] dz \right\} \tag{18}
\]

\[
\zeta_t^2(t) = \frac{2}{L_c\rho_1}\left\{-\kappa_t \int_0^t [T_{s,D_1}(t) - T_f] dt + \int_0^{\zeta_t} zC_{s,l}[T_s(z,t) - T_f] dz \right\} \tag{19}
\]

where \( \kappa_f \) and \( \kappa_t \) are the thermal conductivity of freezing and thawing soils \( (\text{W m}^{-1} \text{K}^{-1}) \), respectively; \( C_{s,f} \) and \( C_{s,l} \) are the volumetric heat capacity of freezing and thawing soil \( (\text{J m}^{-3} \text{K}^{-1}) \), respectively; \( z \) is the soil thickness; \( T_i(z,t) \) is the soil temperature profile within \( z \).

The second terms of the above equations are far less than the first terms (Li and Cheng, 1995), therefore, neglecting them and writing the integral of the surface temperature with time in the form of cumulative freezing/thawing indices; the frost and thaw depth can be expressed in equations of the approximation Stefan solution (The Institute of Geocryology, 1974).

\[
\zeta_f = \sqrt{\frac{2\kappa_f\tau_h}{L_c\rho_1} \sum_{i=1}^{t} (T_f - T_{s,D_i})} \tag{20}
\]

\[
\zeta_t = \sqrt{\frac{2\kappa_t\tau_h}{L_c\rho_1} \sum_{i=1}^{t} (T_{s,D_i} - T_f)} \tag{21}
\]

where \( \tau_h \) is the time length (s) and \( \tau_h = 3600 \text{ s} \) in this study.
In the above equations, thermal conductivity of frozen and thawed soils are functions of soil type, density, and moisture. However, it is difficult to express them in simple relationship. This model uses a lookup table to obtain the thermal conductivity of freezing and thawing soils. The thermal conductivity of typical unsaturated freezing and thawing soils are given in the General Geocryology, which was published by The Institute of Geocryology (1974) of former USSR.

After determining the position of freezing front, the soil temperature in the root zone and deep soil are estimated by a simple function of frost depth. If \( z \leq \zeta_f \), the soil temperature at a given depth \( z \) is

\[
T_{s,z} = T_t + (T_{s,D} - T_t) \left( 1 - \frac{z}{\zeta_f} \right)
\]

where \( D \) is the total soil depth (m).

2.3. Linking the frozen soil parameterization with SiB2

The freeze/thaw cycle is manipulated based on the assumptions that (1) the freezing/thawing process only occurs from the top of a soil column, and the freezing/thawing from the bottom is very small and can be neglected (The Institute of Geocryology, 1974); (2) at any time, there are no more than two freezing/thawing front(s) in a soil column, one accounting for the seasonal penetration of freezing/thawing front, another accounting for the short-term freeze/thaw cycle. Accordingly, once \( T_{s,1} \) is below the freezing point, the freezing index and then the first penetration depth are calculated. When \( T_{s,1} > \zeta_f \), the thawing index is calculated and the energy for thawing \( (T_{s,1} - T_t) \) is subtracted from the freezing index, but the first penetration depth will be kept unchanged until the thawing index can be as great as the freezing index. Otherwise, the second penetration depth that represents short-term freeze/thaw cycle will be calculated. It is set to zero when \( T_{s,1} \leq T_t \) again.

The frozen soil submodel is called in SiB2 just after the surface temperature is updated with snowmelt-refreeze calculation. The procedures of linking the new frozen soil parameterization with SiB2 are as follows:

1. \( T_{s,1} \) is calculated and then \( T_{s,1} \) and the water content in the soil layer where frost/thaw depth are used to determine the position of freezing/thawing front. (2) Calculate the soil temperature of each layer. (3) The liquid water content and its changing rate are calculated, and then the volumetric ice and liquid water contents in each layer are updated. (4) Solve the water balance equation, the hydraulic conductivity is updated by considering the effect of soil freezing/thawing. (5) Calculate the thermal conductivity and heat capacity of soil for the next time step from the updated ice and liquid water contents.

3. Data and model calibration

3.1. Data

3.1.1. GAME-Tibet observations and experiment site

As part of Global Energy and Water Cycle Experiment (GEWEX), the GEWEX Asian Monsoon Experiment (GAME) has been implemented since 1996. GAME selects four representative regions as observation fields. The Tibetan Plateau was chosen as one of the regions because this very high plateau is considered as a heat source of atmosphere in summer and to have an important impact on the Asian monsoon system (Ye and Gao, 1979). The experiment in the Tibetan Plateau is named GAME-Tibet and its overall goal is to clarify the interactions between the land surface and the atmosphere over the plateau in the context of the Asian monsoon system. To achieve this goal, a plateau-scale experiment and a mesoscale experiment were carried out by the prephase observation period (POP) field work in August–September 1997 and the intensive observation period (IOP) field work in May–September 1998. A much denser automatic observation network of meteorology and land surface hydrology has been established for long-period observation (Koike et al., 1999). Numerous high quality data is being obtained and a database of GAME-Tibet IOP and POP has been developed (http://monsoon.t.u-tokyo.ac.jp/tibet/data/index.html).
Fig. 2 shows the observation system of the GAME-Tibet experiment.

The Tibetan Plateau is a region with wide area of frozen soil. Permafrost and seasonal frozen soil underlay about 50% and 44% of the plateau’s area, respectively (Li and Cheng, 1999). Therefore, the GAME-Tibet observations of soil moisture and temperature provide important data sets for validating frozen soil parameterization. Because the seasonal freezing and thawing occurs from September to the next May in most area of the Tibetan Plateau, and because the model should be calibrated and validated independently, we use the data obtained in summer to calibrate the parameter set of SiB2 and use the data obtained in winter to validate the new frozen soil parameterization. Until now, all the data during the POP and IOP have been processed; the data after the IOP will be obtained and processed in near future. Consequently, summer data are available for model calibration at all Automatic Weather Station (AWS) sites, but a whole year hourly simultaneous observation of both AWS and Soil Moisture and Temperature Measurement System (SMTMS) is only available at D66 and MS3608. In addition, hourly precipitation was also measured at these two sites during IOP, and supplementary daily precipitation data of 1997 was obtained near MS3608 at Nagqu station. Accordingly, we select the MS3608 site for both calibration and validation for consistency. MS3608 is located at 31°13.6’ N latitude, 91°47.0’ E longitude, with an elevation of 4610 m. The landscape at this site is short-grazed grassland, which usually lasts 3–4 months in summer.

3.1.2. Forcing data

One AWS was installed at MS3608. The station consists of a 6-m panza mast, a solar panel, a rechargeable battery, data logger, and the sensors. The data logger is a programmable one, TEAC DR101M. The sensors are listed in the Table 1.

Among the six forcing variables required to run SiB2, the shortwave downward radiation, air temperature, and wind speed are from AWS’s measurement directly. The longwave downward radiation is calculated by Brunt’s equation (Monteith, 1973), which is a standard method in SiB2. The vapor pressure is transformed from humidity. The precipitation in summer is measured during the IOP using a rain gauge with digital recorder, Ogasawara weight. In winter, the hourly precipitation is interpolated by averaging the daily precipitation in Nagqu station among 24 h. This
interpolation is reasonable because the precipitation is very small in winter in the Tibetan Plateau. It also smooths the precipitation.

The meteorological data at MS3608 were intermittently missed from February 3 to May 2, 1998 because of some malfunction. To remove the unreliable data and to keep a time consistency with soil temperature and moisture observations, the one-dimensional offline SiB2 with frozen soil parameterization is run with the forcing data from July 6 to August 10, 1998 for calibration, and from September 1, 1997 to January 31, 1998 for validation. Unfortunately, the validation in the thawing period is not capable of being implemented in this paper due to data missing, but this will be done in the future with the forthcoming one-dimensional data set and the high resolution GAME reanalysis data set.

3.1.3. Validation data

The validation of frozen soil parameterization needs the observation of soil moisture and temperature profiles. One set of SMTMS as shown in Fig. 3 was installed at MS3608; the system had been working continuously for 1 year except in early May when some troubles occurred due to static electricity under the very dry condition. The time domain reflectometry (TDR) was employed to measure the soil moisture. Volumetric water content at depths of 4, 20, 60, 100, 160, and 196 cm was measured by TDR (TRIME-MUX6 cable tester). The probes consist of a pair of rods and were installed horizontally into the profiles. The data were recorded by the DATAMark-LS3000 data logger. The measurement interval was 1 h. The soil temperature at depths of 4, 20, 60, 80, 100, 130, 160, 200, and 242 cm was measured by the infrared radiometer, TASCO THI500. A few missing data of soil temperature in September 10 and 11, 1997 were linearly interpolated from the neighboring data in time.

In addition, flux data of solar radiation, sensible heat, latent heat, and ground thermal heat were measured and used to calibrate SiB2 in the Amdo Planetary Boundary Layer (PBL) site (Fig. 2) (Gao and Kim, personal communication). However, the heat flux imbalance is still an unresolved problem in the GAME-Tibet observation (Wang et al., 1999). Validation of heat fluxes needs a reexamination of the imbalance.

3.2. Calibration

3.2.1. Parameter set in the Tibetan Plateau

3.2.1.1. Parameters associated with land cover type.

The land cover type at MS3608 belongs to the short vegetation/C4 grassland, which is the sixth class in Sellers et al.’s (1996b) definition of global land cover classification. Most of the static parameters associated with land cover type are derived from Sellers et al. (1996b) directly, while canopy height,

<table>
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<th>Channel</th>
<th>Variables</th>
<th>Type of sensor</th>
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<td>aerobane</td>
</tr>
<tr>
<td>2</td>
<td>wind speed</td>
<td>aerobane</td>
</tr>
<tr>
<td>3</td>
<td>air temperature</td>
<td>Pt-100</td>
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<td>4</td>
<td>air humidity</td>
<td>capacitance sensor</td>
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<td>infrared thermometer</td>
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<td>soil temperature</td>
<td>Pt-100</td>
</tr>
<tr>
<td>8</td>
<td>atmospheric pressure</td>
<td>semiconductor sensor</td>
</tr>
</tbody>
</table>

Table 1

Sensors of AWS (after Ishikawa et al., 1999)

Fig. 3. Soil moisture and temperature measurement system (SMTMS).
vegetation cover fraction, and root depth are measured in the GAME-Tibet experiment. The leaf reflectance is derived from the spectral reflectance measurement during the POP and the IOP with the spectralscope, FieldSpec FR. The bandwidth is from 380 to 2500 nm with a fine spectral resolution of 1 nm in the visible and infrared bands. Table 2a lists out the parameter values used in this study.

3.2.1.2. Parameters associated with soil type. The soil type at MS3608 is sandy loam, which is the second class in Sellers et al. (1996b). The values of the parameters defined by Clapp and Hornberger (1978) \((B, \Psi_s, K_s, \theta_s)\) are optimally calibrated with the summer data.

3.2.1.3. Time-space varying vegetation parameters. NDVI is calculated by using the measurement of spectral reflectance directly. Reflectance of the visible and the infrared bands is integrated from 580 to 680 nm and from 725 to 1100 nm, respectively. And then, FPAR, \(N\), and \(L_T\) are derived from NDVI by the methods of Sellers et al. (1996b) and with the reference to Wang (1996).

3.2.1.4. Aerodynamic parameters. Various methods have been used to calculate the value of aerodynamic roughness in the Tibetan Plateau (Wang et al., 1999; Gao et al., 2000b; Zhou et al., 2000). A plausible value is 1–3 mm in the near neutral state, and it was likely that the growth of grass during summer monsoon has a little influence on it. In this paper, we use 0.002 m in summer and 0.001 m in winter. Other aerodynamic parameter values are obtained with the reference to Gao and Kim (personal communication).

3.2.2. Results with summer data

The original SiB2 without frozen soil was ran with the forcing data in summer from July 6 to August 10, 1998. The primary object was to validate and when necessary, to optimally calibrate the parameter set in the Tibet Plateau, and then this optimized parameter set can be used to validate the modified SiB2 with frozen soil parameterization. Sellers et al. (1989) presented an optimization procedure by minimizing the difference between model calculations and micrometeorological observations, but in this study, since most parameters can be directly retrieved from the parameter sets defined for global biomo types (Sellers et al., 1996b), and other parameters are believed to fall within a reasonable range, we employ a simple method of interactively comparing the simulation results with observations. Table 2 shows the parameter values. Moreover, Fig. 4 shows the comparison of the simulated surface temperature with the observed one. The calculated result is realistic, with a mean absolute error of 1.32 K. Fig. 5 compares the simulated soil moisture in the surface layer with observations at 4 cm, simulated soil moisture in the root zone with both observations at 4 and 20 cm, and simulated soil moisture in deep soil with observations at 100 cm. The simulation results of soil moisture are reasonable, though the simulated moisture in the surface layer is more sensitive to precipitation.

Additionally, a parameter set with similar values to Table 2a–d was obtained by calibrating with flux data (Gao and Kim, personal communication). But as discussed before, the flux data need to be checked again because of the imbalance problem of heat flux in the Tibetan Plateau.

3.3. Validation results and analyses

The modified SiB2 with frozen soil parameterization was tested using the forcing data in winter from September 1, 1997 to January 31, 1998, together with the optimized parameter set given in Table 2. Simulation results of frost depth, soil temperature, and soil moisture are compared with the validation data measured by SMTMS.

3.3.1. Frost depth

The frost/thaw depth was not measured directly in the GAME-Tibet experiment. It was determined by linearly interpolating soil temperature profiles to find the freezing point. The deepest position at which temperature turns from below freezing to above freezing is considered to be the frost depth. Interpolated and simulated values of frost depth are compared in Fig. 6. This model estimates the frost depth precisely, with a mean absolute error of 0.15 m, which is about 9% of the total frost depth. The errors are mainly in the deep soil. Obviously, they are closely related with the errors of soil temperature and moisture estimations.
### Table 2

(a) Static parameters associated with land cover type

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Value</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>(z_2)</td>
<td>canopy top height</td>
<td>m</td>
<td>0.04</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(z_1)</td>
<td>canopy base height</td>
<td>m</td>
<td>0.01</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(V)</td>
<td>vegetation cover fraction</td>
<td>–</td>
<td>0.9</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(\lambda_k)</td>
<td>leaf angle distribution factor</td>
<td>–</td>
<td>–0.3</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(D_r)</td>
<td>root depth</td>
<td>m</td>
<td>0.15</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(\Psi_e)</td>
<td>1/2 critical leaf water potential limit</td>
<td>m</td>
<td>–200.0</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\delta_{0,0})</td>
<td>leaf transmittance, iv: radiation</td>
<td>–</td>
<td>0.07</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\delta_{1,0})</td>
<td>wavelength, 0 = visible, 1 = near infrared;</td>
<td>–</td>
<td>0.25</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\delta_{0,1})</td>
<td>il: vegetation state, 0 = live (green),</td>
<td>–</td>
<td>0.22</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\delta_{1,1})</td>
<td>and 1 = dead (stems and trunk)</td>
<td>–</td>
<td>0.38</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\alpha_{0,0})</td>
<td>leaf reflectance</td>
<td>–</td>
<td>0.21</td>
<td>GAME-Tibet and</td>
</tr>
<tr>
<td>(\alpha_{1,0})</td>
<td>–</td>
<td>–</td>
<td>0.58</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\alpha_{0,1})</td>
<td>–</td>
<td>–</td>
<td>0.30</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\alpha_{1,1})</td>
<td>–</td>
<td>–</td>
<td>0.58</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\varepsilon)</td>
<td>intrinsic quantum efficiency</td>
<td>mol mol(^{-1})</td>
<td>0.05</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(m)</td>
<td>conductance–photosynthesis slope parameter</td>
<td>–</td>
<td>4.0</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(b)</td>
<td>conductance–photosynthesis intercept</td>
<td>mol m(^{-2}) s(^{-1})</td>
<td>0.04</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(f_b)</td>
<td>leaf respiration factor</td>
<td>–</td>
<td>0.025</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\beta_{wc})</td>
<td>wc and we coupling parameter</td>
<td>–</td>
<td>0.80</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(\beta_{ws})</td>
<td>wc, we, and ws coupling parameter</td>
<td>–</td>
<td>0.95</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_5)</td>
<td>high temperature stress factor, respiration</td>
<td>K(^{-1})</td>
<td>1.30</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_6)</td>
<td>high temperature stress factor, respiration</td>
<td>K</td>
<td>328.16</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_3)</td>
<td>slope of low temperature inhibition function</td>
<td>K(^{-1})</td>
<td>0.2</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_4)</td>
<td>1/2 point of low temperature inhibition function</td>
<td>K</td>
<td>288.16</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_1)</td>
<td>slope of high temperature inhibition function</td>
<td>K(^{-1})</td>
<td>0.3</td>
<td>Sellers et al. (1996b)</td>
</tr>
<tr>
<td>(s_2)</td>
<td>1/2 point of high temperature inhibition function</td>
<td>K</td>
<td>313.16</td>
<td>Sellers et al. (1996b)</td>
</tr>
</tbody>
</table>

(b) Static parameters associated with soil type

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Value</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>(D_T)</td>
<td>total depth of three soil moisture layers</td>
<td>m</td>
<td>2.0</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(\alpha_s, \alpha_i)</td>
<td>soil reflectance</td>
<td>–, –</td>
<td>0.12, 0.20</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>(B)</td>
<td>soil wetness exponent</td>
<td>–</td>
<td>4.20</td>
<td>optimization</td>
</tr>
<tr>
<td>(\Psi_s)</td>
<td>soil tension at saturation</td>
<td>m</td>
<td>–0.05</td>
<td>optimization</td>
</tr>
<tr>
<td>(K_s)</td>
<td>hydraulic conductivity at saturation</td>
<td>m s(^{-1})</td>
<td>1.2 \times 10^4</td>
<td>optimization</td>
</tr>
<tr>
<td>(\theta_s)</td>
<td>soil porosity</td>
<td>%</td>
<td>0.42</td>
<td>optimization</td>
</tr>
<tr>
<td>(\Phi_s)</td>
<td>mean topographic slope</td>
<td>radians</td>
<td>0.02</td>
<td>GAME-Tibet</td>
</tr>
</tbody>
</table>

(c) Time–space varying vegetation parameters

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Value</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>FPAR</td>
<td>canopy absorbed fraction of photosynthetically active radiation (par)</td>
<td>–</td>
<td>0.30</td>
<td>0.20</td>
</tr>
<tr>
<td>(V_{max,0})</td>
<td>Rubisco velocity of sun-leaf</td>
<td>mol m(^{-2}) s(^{-1})</td>
<td>3.0e – 5</td>
<td>3.0e – 5</td>
</tr>
<tr>
<td>(G(u)/u)</td>
<td>Time–mean leaf projection (g(mu)/mu)</td>
<td>–</td>
<td>1.0</td>
<td>1.0</td>
</tr>
</tbody>
</table>
3.3.2. Soil temperature

Comparison between the observations and model calculations of soil temperatures at 4 cm (Fig. 7a) indicates that (1) as a whole, the simulated soil temperature is close to the observed one except some days in December, with a mean absolute error of 2.84 K and a correlation coefficient of 0.89. (2) Simulation result displays a much greater variation in the phase transition period, while the observations show that soil temperature keeps as a quasi-constant around the freezing point during this period. This can be explained using Eq. (17), where the liquid water changing rate is a monotonic decreasing function of soil temperature. When soil temperature is just below the freezing point, the phase changing rate reaches its maximum value, making heat diffusivity a very small value and therefore hampering the heat transfer. Accordingly, when soil temperature varies around the freezing point, heat is transferred very slowly. This process can be modeled with very fine resolution of soil layers; a surface layer of 4 cm even amplifies the process.

The model calculation of soil temperature at $D_1 + D_2/2$ (9.5 cm) is used to represent the temperature in the root zone. It is compared with the observation at 20 cm in Fig. 7b. The results indicate

![Graph comparing observation and simulation results of surface temperature](image-url)

Fig. 4. Comparison of the observations and model simulation results of surface temperature at MS3608 site, Tibetan Plateau, from July 6 to August 10, 1998.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Summer</th>
<th>Winter</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$N$</td>
<td>green leaf fraction</td>
<td>–</td>
<td>0.90</td>
<td>0.70</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>$L_T$</td>
<td>leaf area index</td>
<td>–</td>
<td>0.80</td>
<td>0.30</td>
<td>GAME-Tibet</td>
</tr>
</tbody>
</table>

(c) Time–space varying vegetation parameters

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Value</th>
<th>Value</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>z$_0$</td>
<td>roughness length</td>
<td>m</td>
<td>0.002</td>
<td>0.001</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>$d$</td>
<td>zero plane displacement</td>
<td>m</td>
<td>0.006</td>
<td>0.003</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>$C_1$</td>
<td>rb (bulk canopy boundary layer resistance)</td>
<td>–</td>
<td>4.0</td>
<td>4.0</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>$C_2$</td>
<td>rc (bulk stomatal resistance of upper-story vegetation) coefficient</td>
<td>–</td>
<td>70.0</td>
<td>70.0</td>
<td>GAME-Tibet</td>
</tr>
</tbody>
</table>

(d) Aerodynamic parameters

Table 2 (continued)

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition</th>
<th>Unit</th>
<th>Value</th>
<th>Value</th>
<th>Data source/reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$N$</td>
<td>green leaf fraction</td>
<td>–</td>
<td>0.90</td>
<td>0.70</td>
<td>GAME-Tibet</td>
</tr>
<tr>
<td>$L_T$</td>
<td>leaf area index</td>
<td>–</td>
<td>0.80</td>
<td>0.30</td>
<td>GAME-Tibet</td>
</tr>
</tbody>
</table>
that (1) the simulated soil temperature during phase transition is realistic. (2) The soil temperature after phase transition is just a resemblance of $T_{s,D}$ displayed in Fig. 7a and shows much higher variation than observation. This can be explained by Eq. (22). However, the objective of predicting soil moisture is to calculate the unfrozen water content, which has a very small changing rate when soil is completely frozen. Therefore, more consideration is put on the averaged trend of soil temperature, for which a simple relationship of Eq. (22) can give good results.
The model calculation of soil temperature in the deep soil is compared with the observation at 100 cm. As shown in Fig. 7c, the results are very close to the actual situation, with a mean absolute error of 1.25 K and a correlation coefficient of 0.95.

3.3.3. Soil moisture

The simulation results of soil moisture using the modified SiB2 with frozen soil parameterization are compared with the observations as well as the model calculations using the original SiB2 without frozen soil parameterization in Fig. 8.

Fig. 8a shows the comparison in the surface layer. The results indicate that the new frozen soil parameterization improves the soil moisture estimation significantly. The simulation results of liquid water content with and without frozen soil parameterization are quite different. The mean absolute errors of 2 (simulation result in the surface layer with frozen soil parameterization) and 3 (simulation result in the surface layer without frozen soil parameterization) are 0.020 and 0.066 (m³ m⁻³), respectively. The correlation coefficient between 1 (observation of soil moisture at 4 cm) and 2 is 0.89, and between 1 and 3 is 0.82. For the phase transition time, the main transition period from mid-October to mid-November is predicted well in the model, though the calculated phase transition is earlier and the moisture variation is greater than the actual situation. This error is due to the amplified simulation result of the soil temperature at 4 cm during the phase transition period.

Fig. 8b shows the comparison in the root zone. The central depth of the root zone is 9.5 cm, which falls within the observations of 4 and 20 cm. Since the observation at 20 cm has high noise (Fig. 8b), the observation at 4 cm is used as the comparison baseline. The results also indicate that the estimation of liquid water content is much improved with frozen soil parameterization. The mean absolute errors of 3 (simulation result in the root zone with frozen soil parameterization) and 4 (simulation result in the root zone without frozen soil parameterization) are 0.013 and 0.070 (m³ m⁻³), respectively. The correlation coefficient between 1 (observation of soil moisture at 4 cm) and 3 is 0.97, and between 1 and 4 is 0.87. The phase transition time in the root zone is predicted realistically, but the amplitude is higher than the observation.

Fig. 8c shows the comparison in the deep soil. The results indicate that (1) the phase transition in the deep soil is predicted with frozen soil parameterization, but the transition is quicker than the observation. (2) The model without frozen soil parameterization manifests a higher liquid water content, and conversely the
model with frozen soil parameterization manifests lower liquid water content. The mean absolute errors of 2 (simulation result in the deep soil with frozen soil parameterization) and 3 (simulation result in the deep soil without frozen soil parameterization) are 0.051 and 0.038 \( \text{m}^3 \text{m}^{-3} \), respectively. The error of 2 is difficult to explain. The observation shows that the unfrozen water content keeps a high value when soil temperature is a few Kelvin below the freezing point. A high osmotic potential in the deep soil can account for the phenomenon, but this explanation needs to be proved.

Fig. 7. Comparisons of the observations and model simulation results of soil temperature in (a) the surface layer, (b) the root zone, and (c) the deep soil at MS3608 site, Tibetan Plateau, from September 1, 1997 to January 31, 1998 (the freezing point is illustrated in each figure for visualized analysis).
Fig. 8. Comparisons of the observations and model simulation results of soil moisture in (a) the surface layer, (b) the root zone, and (c) the deep soil at MS3608 site, Tibetan Plateau, from September 1, 1997 to January 31, 1998 (precipitation observation is illustrated together with (a) for direct visualization of its relationship with the soil moisture in the surface layer).
4. Conclusions

This paper describes a new frozen soil parameterization in the land surface scheme. A modified approximate Stefan solution was incorporated in the framework of the land surface model-SiB2 to calculate the frost/thaw depth over time and to estimate the soil moisture and temperature profiles during the freeze/thaw cycle. The new frozen soil parameterization included the following improvements to the original SiB2. (1) The governing equations of water balance were modified to involve the soil freezing/thawing process. The unfrozen water content was expressed by a power function derived from experiments. The interlayer water flow and the hydraulic conductivity were calculated according to soil freezing. (2) The soil temperature at 4 cm was estimated using an analytic method. And then, the approximate Stefan solution was used to predict the frost/thaw depth over time and to calculate soil temperature using a simple relationship with frost depth. The freezing/thawing indices were calculated from 4 cm below the surface because this manipulation can remove the great variation in the surface temperature, making the Stefan solution’s assumption of linear soil temperature profile in the frozen soil column more reasonable. (3) Heat capacity and thermal conductivity were updated with the unfrozen water content and its changing rate.

One-dimensional validation of the model was implemented with the GAME-Tibet observations at MS3608, Tibetan Plateau. First, the original SiB2 without frozen soil parameterization was calibrated using the observations in summer from July 6 to August 10, 1998; an optimized parameter set for the short grassland in the Tibetan Plateau was obtained. Then, the modified SiB2 with frozen soil parameterization was validated using the soil temperature and moisture observations in winter from September 1, 1997 to January 31, 1998. The results showed that (1) the new frozen soil parameterization estimates the frost depth with a mean absolute error of 0.11 m or 6% of the total frost depth. (2) The model calculations of the soil temperature are reasonable, but the diurnal variation of temperature is amplified. (3) The model improves the soil moisture estimation in the surface layer and the root zone significantly. The errors of the liquid water content calculations with frozen soil parameterization are much smaller than those without frozen soil parameterization. The phase transition time is predicted realistically. (4) The model calculations in the deep soil manifest the phase transition process but underestimate the liquid water content.

Analyses showed that the errors in the model are closely correlated with each other. An amplified estimation of soil temperature variation results in the fluctuation of soil moisture during phase transition period. Some of the errors are due to the approximate Stefan solution, which behaves better in accounting for daily but not diurnal penetration of the frost/thaw depth. An improvement with a numeric Stefan solution can avoid the weakness, but needs much finer resolution of soil layers. For this, the simple structure of the soil model in land surface schemes has to be changed, and more detailed information of soil is required.

At last, since the water content in the surface layer and the root zone are directly associated with evaporation and canopy transpiration, the improved performance of the model in estimating liquid water content during soil freezing is believable to achieve a more reasonable simulation of sensible and latent heat fluxes in winter.

Acknowledgements

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