Representing the effects of alpine grassland vegetation cover on the simulation of soil thermal dynamics by ecosystem models applied to the Qinghai-Tibetan Plateau

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1] Soil surface temperature is a critical boundary condition for the simulation of soil temperature by environmental models. It is influenced by atmospheric and soil conditions and by vegetation cover. In sophisticated land surface models, it is simulated iteratively by solving surface energy budget equations. In ecosystem, permafrost, and hydrology models, the consideration of soil surface temperature is generally simple. In this study, we developed a methodology for representing the effects of vegetation cover and atmospheric factors on the estimation of soil surface temperature for alpine grassland ecosystems on the Qinghai-Tibetan Plateau. Our approach integrated measurements from meteorological stations with simulations from a sophisticated land surface model to develop an equation set for estimating soil surface temperature. After implementing this equation set into an ecosystem model and evaluating the performance of the ecosystem model in simulating soil temperature at different depths in the soil profile, we applied the model to simulate interactions among vegetation cover, freeze-thaw cycles, and soil erosion to demonstrate potential applications made possible through the implementation of the methodology developed in this study. Results showed that (1) to properly estimate daily soil surface temperature, algorithms should use air temperature, downward solar radiation, and vegetation cover as independent variables; (2) the equation set developed in this study performed better than soil surface temperature algorithms used in other models; and (3) the ecosystem model performed well in simulating soil temperature throughout the soil profile using the equation set developed in this study. Our application of the model indicates that the representation in ecosystem models of the effects of vegetation cover on the simulation of soil thermal dynamics has the potential to substantially improve our understanding of the vulnerability of alpine grassland ecosystems to changes in climate and grazing regimes.


1. Introduction

2] Soil temperature is an important factor that controls many biogeophysical and biogeochemical processes. There are a variety of ways to estimate soil temperature in environmental models. For example, in the LPJ-DGVM and BEPS, soil temperature follows the air temperature cycle, which is assumed to be approximately sinusoidal, with a damped oscillation around a common mean and with a temporal lag [Stitch et al., 2003; Ju and Chen, 2005]. In Biome-BGC, soil temperature is calculated using a running average of air temperatures [Kimball et al., 1997; Bond-Lamberty et al., 2005]. Some other models (e.g., IBIS, DayCENT, TEM 5.0, and DOS-TEM) simulate soil temperatures by solving finite difference equations of soil heat conduction [Foley et al., 1996; Parton et al., 1998; Zhuang et al., 2001; Yi et al., 2009].

3] Surface temperature of soil is a critical boundary condition for solving finite difference heat equations; it is controlled not only by atmospheric and soil conditions, but also by vegetation cover. In sophisticated land surface models with approximately hourly time steps, soil surface temperature is calculated by iteratively solving the energy balance equations at the soil surface that take in account incoming and reflected solar radiation, incoming and outgoing longwave radiation, sensible and latent heat fluxes between the surface and atmosphere, and ground heat flux [Oleson et al., 2004]. In ecosystem, permafrost, and hydrology models driven with data at daily or monthly time steps, soil
surface temperature is alternatively assumed to be the same as atmosphere temperature [Zhuang et al., 2001; Jafarov et al., 2012], a linear function of air temperature [Zhang et al., 1997; Oelke et al., 2003], or derived by multiplying a factor (n-factor), which is the ratio of seasonal degree day sums at the ground surface to those in the air at standard screen height over a specific period [Carlson, 1952]. For example, n-factor of the thaw season can be expressed as follows:

\[ n_t = \frac{\int_0^{T_s} (T_s - T_f) \, dt}{\int_0^{T_a} (T_a - T_f) \, dt}, \]

where \( \theta_s \) and \( \theta_a \) are the number of days when surface and air temperatures are greater than 0°C, respectively. \( T_s \) and \( T_a \) are the ground surface and air temperatures, and \( T_f \) is the freezing point (0°C). n-factor was originally used in construction and engineering applications. Recently, it has been adopted for use in ecosystem, hydrology, and permafrost models [Hayashi et al., 2007; Peterson and Krantz, 2008; Juliussen and Humlum, 2007; Yi et al., 2009; Quinton et al., 2009]. However, a constant value of n-factor is generally used for the warm and/or cold seasons, but see Juliussen and Humlum [2007], which considered the effects of potential solar radiation on n-factor. Besides atmospheric environmental factors, vegetation cover also affects surface energy balance and thus soil surface temperature. The CENTURY and DayCENT models consider vegetation cover by calculating the maximum/minimum soil surface temperature using maximum/minimum air temperature and biomass [Parton, 1984; Parton et al., 1998]. In the dynamic organic soil version of the Terrestrial Ecosystem Model (DOS-TEM), a dynamic n-factor approach is used, which was derived based on the n-factor estimates of sites with different stand ages [Yi et al., 2010].

Alpine grassland ecosystems occupy more than half of the area of the Qinghai-Tibetan Plateau and are vulnerable to global climate change and anthropogenic disturbances [Sun and Zheng, 1998]. It is very important to project the responses of alpine grassland to the changes of climate and soil environment, such as the degradation of permafrost [Yi et al., 2011]. However, alpine grassland ecosystems also affect the atmosphere and soil by changing land surface-atmosphere interactions and soil thermal and hydrological processes. Both experimental studies [Blok et al., 2010] and modeling studies [Yi et al., 2007] have demonstrated that increases in vegetation cover would delay thawing of permafrost in the arctic region. Field studies also suggested that changes in alpine grassland cover affect soil thermal dynamics on the Qinghai-Tibetan Plateau [Wang et al., 2010]. These effects have seldom been considered in modeling studies on the Qinghai-Tibetan Plateau within ecosystem models [Zhang et al., 2007; Zhuang et al., 2010]. Therefore, the main objectives of this study are (1) to develop a methodology for the estimation of soil surface temperature that accounts for vegetation cover in alpine grassland ecosystems by integrating observations from meteorological stations on the Qinghai-Tibetan Plateau and simulated results from a land surface model, (2) to implement the methodology and validate the performance of an ecosystem model in simulating soil temperature dynamics, and (3) to apply the modified ecosystem model to analyze surface soil erosion as a demonstration of the types of applications made possible through the implementation of the methodology.

We chose the analysis of surface soil erosion as a demonstration application because soil erosion in alpine grassland ecosystems of the Qinghai-Tibetan Plateau depends on interactions of vegetation cover and freeze-thaw cycles, which themselves are influenced by soil erosion. An increase in vegetation cover decreases the daily maximum soil surface temperature, but increases the daily minimum soil surface temperature, and vice versa [Yang et al., 1999]. On the Qinghai-Tibetan Plateau, the freeze-thaw cycle is very common. Based on field measurements, the number of days with a surface soil freeze-thaw cycle can reach 180 days in the central part of the Qinghai-Tibetan Plateau [Yang et al., 2006a]. Freeze-thaw cycles cause changes in soil aggregate stability [Oztas and Fayetorbay, 2003], which affects soil erosion rates [Ferrick and Gatto, 2005]. Therefore, changes in vegetation cover on the Qinghai-Tibetan Plateau may have significant impacts on soil thermal dynamics and soil erosion, which will then affect vegetation growth. To be able to use an ecosystem model as a tool to simulate these feedback pathways requires that the model be able to properly consider the effects of vegetation cover on soil surface temperature.

2. Methods

In this study, we primarily used two models, a sophisticated land surface model (Community Land Model; CLM3) and an ecosystem model (DOS-TEM). Because soil surface temperatures on the Qinghai-Tibetan Plateau were only measured for bare soil conditions at meteorological stations, we use CLM3 as a tool to simulate the effect of vegetation cover on soil surface temperature. Therefore, we first validated the simulation of 2 cm and 5 cm soil temperatures of CLM3 against measurements from a vegetated flux tower on the Qinghai-Tibetan Plateau. We then used CLM3 to simulate soil surface temperature for a range of environmental conditions and vegetation covers with the purpose of developing empirical equations for estimating soil surface temperature for incorporation into DOS-TEM. We compared these empirical equations to those developed for bare-ground measurements of soil surface temperature at a network of meteorological stations on the Qinghai-Tibetan Plateau. We then incorporated one of the equations into DOS-TEM and validated DOS-TEM with respect to its simulation of soil temperatures at different depths in the soil profile across the network of meteorological stations on the Qinghai-Tibetan Plateau. Sensitivity tests were also performed to investigate the effects of vegetation cover change on soil thermal dynamics. Finally, we applied DOS-TEM to investigate the potential effects of surface erosion in alpine grassland ecosystems of the Qinghai-Tibetan Plateau associated with changes in freeze-thaw cycles caused by changes in vegetation cover.

2.1. Model Descriptions

CLM3 is designed to simulate the exchanges of water, energy, and momentum between the land surface and atmosphere [Oleson, 2004]. CLM3 is driven by atmospheric variables at half-hour time steps. The atmospheric variables include air temperature, downward longwave radiation, downward shortwave radiation, precipitation, humidity, and wind speed. The surface temperatures of soil and snow are
calculated iteratively in CLM3 by solving energy balance equations at soil/snow surfaces. The two-stream approximation method is used for radiation transfer within vegetative canopies. The performance of CLM3 in cold regions has been demonstrated in several studies [Yi et al., 2006; Nicolsky et al., 2007; Lawrence et al., 2008]. In this study, CLM3 was used to simulate soil surface temperature under different vegetation covers for the purpose of developing empirical equations of soil surface temperature for incorporation into DOS-TEM.

The TEM family of models is designed to simulate the carbon and nitrogen pools of vegetation and soil, and the carbon and nitrogen fluxes among vegetation, soil, and the atmosphere [Raich et al., 1991; McGuire et al., 1992]. In general, the various versions of TEM are driven by monthly atmospheric variables, i.e., air temperature, radiation, precipitation, and vapor pressure. In the arctic and boreal regions, thawing of permafrost is delayed by organic soil layers, which are affected by wildfire disturbances and ecological successions. A recent version of TEM (i.e., DOS-TEM) can simulate the dynamics of organic soil layers [Yi et al., 2010]. The DOS-TEM consists of four modules including environmental, ecological, fire disturbance, and dynamic organic soil modules. The environmental module operates at a daily time step using daily air temperature, surface solar radiation, precipitation, and vapor pressure data sets, which are downscaled from monthly input data. It considers the radiation and water fluxes among the atmosphere, canopy, snowpack and soil. Soil moisture and temperature are updated daily. A two-directional Stefan algorithm is used to predict the positions of freezing/thawing fronts in the soil [Woo et al., 2004]. The temperatures of soil layers above the first freezing/thawing front and below the last freezing/thawing front are updated separately by solving finite difference equations. Temperatures of the soil layers between the first and last freezing/thawing fronts are assumed to be at the freezing point. Soil moisture is only updated for unfrozen layers by solving Richard equations. Both the thermal and hydraulic properties of soil layers are affected by its water content [Yi et al., 2009]. In each soil layer, relatively reactive and nonreactive soil carbon pools were simulated [Yuan et al., 2012].

The soil thermal conductivity in DOS-TEM is calculated following Farouki [1986]. However, Luo et al. [2009] found that this algorithm overestimates soil thermal conductivity on the Qinghai-Tibetan Plateau. In this study, we applied a new soil thermal conductivity algorithm proposed by Luo et al. [2009].

\[
k = \begin{cases} 
K_e k_{sat} + (1 - K_e) k_{dry} & S_r > 1 \times 10^{-5} \\
K_e k_{dry} & S_r \leq 1 \times 10^{-5}
\end{cases}
\]

\[
k_{sat} = \begin{cases} 
k_{sat}^{1-\theta_{sat}} k_{sat}^\theta_{sat} & T \geq T_f \\
k_{sat}^{1-\theta_{sat}} k_{sat}^\theta_{sat} & T < T_f
\end{cases}
\]

\[
k_f = k_d^\eta k_a^\kappa
\]

\[
k_{sat} = \theta_{sat} \times 10^{-\theta_{sat}}
\]

\[
K_e = \frac{k_s}{S_r + (k - 1)S_r}
\]

where \(k\), \(k_{sat}\), \(k_{dry}\), \(k_d\), \(k_a\), \(k_{sat}\), \(k_{dry}\), \(k_p\), and \(k_s\) are thermal conductivities (W m\(^{-1}\) K\(^{-1}\)) of soil, saturated soil, dry soil, soil solid, liquid water, ice, quartz sand, and other components, respectively. \(\theta_{sat}\) and \(\theta_{dry}\) are porosity and liquid water content of soil (%), respectively. \(K_e\) is the Kersten number. \(S_r\) is the soil saturation. \(\kappa\), \(\eta\), and \(k\) are three parameters, whose values for different soil types can be found in Luo et al. [2009].
2.2. Validation of CLM3 at the Vegetated Flux Tower

[10] Soil/snow surface temperatures of bare ground are regularly measured by a network of meteorological stations of the Chinese Meteorological Administration (Figure 1). Only two of the meteorological stations in this network are located in regions with permafrost; the data quality of these two stations is not as good as other stations in the network, and the active layer thickness is not measured at these stations. In addition, there are no measurements of soil moisture measurement at meteorological stations, and there are few direct measurements of soil surface temperatures under vegetation cover on the Qinghai-Tibetan Plateau. Thus, in this study, CLM3 was used to simulate soil moisture and soil surface temperature under different vegetation cover conditions. Half-hour atmospheric temperature, incoming solar radiation, precipitation, vapor pressure, and wind speed from a flux tower site over the period from January 2005 to December 2006 were used to drive CLM3. The site is located in Xidatan (94°08′E, 35°43′N, 4538 m above sea level (asl)) on the Qinghai-Tibetan Plateau (Figure 1). Over the study period, the mean annual air temperature, precipitation, and downward solar radiation were −3.4°C, 440 mm, and 218 W/m², respectively (Figure 2). The vegetation of the observation field is alpine Kobresia pygmaea meadow [Yao et al., 2011]. Leaf area index (LAI) is an important factor, which controls a number of biogeophysical and biogeochemical processes, including evapotranspiration and photosynthesis. It is defined as one half of the total green leaf area per unit ground surface area [Chen et al., 1997]. In this study, we specified LAI to be 0.2 in January and 2.5 in August, and between 0.2 and 2.5 in other months. The maximum LAI is 2.5 at this site, and the soil texture is sandy loam (R. Li, Cryosphere Research Station on Qinghai-Xizang Plateau, Cold and Arid Region Environmental and Engineering Research Institute, personal communication, 2012).

[11] CLM3 was first validated using the measured near-surface soil temperature (2 cm) and moisture (5 cm) during the period of January 2005 to December 2006 at Xidatan. It was then run to generate a large number of samples with different combinations of changed atmospheric forcings on the basis of site measurements. Air temperature was changed from −2°C to +12°C at 2°C intervals based on the temperature ranges of Xidatan and those of the meteorological stations on the Qinghai-Tibetan Plateau. Precipitation was changed from −50% to +150% at 20% intervals. Incoming shortwave radiation was changed from −50% to +150% at 20% intervals (if the incoming shortwave radiation is greater than 95% of the solar constant, then the sample was excluded). The leaf area index was also changed from 0 to 2.5 at intervals of 0.5.

2.3. Development and Validation of Empirical Equations Based on CLM3 Simulations

[12] We used the results of the CLM3 simulations across the range of environmental conditions to develop equations for daily mean soil surface temperature (ts, K). Our approach was to estimate daily soil surface temperature based on equations that estimate daily maximum soil surface temperature (tsmax, K) and daily minimum soil surface temperature (tsmin, K), which were each evaluated for their dependence on simulated daily maximum air temperature (tamax, K); daily minimum air temperature (tamin, K), other atmospheric variables (downward solar radiation, wind, and precipitation), and LAI. We fitted equations for both bare (LAI = 0) and vegetated soil.

[13] We also developed similar equations for bare soil based on measurements across the network of meteorological stations on the Qinghai-Tibetan Plateau. The standard sampling protocol at the meteorological stations is that soil temperatures at 0, 5, 10, 15, and 20 cm are measured at 2, 8, 14, and 20 o’clock every day. Some stations have measurements of deeper soil temperature (down to 3.2 m). The observation field is vegetation free (Figure 1). Thermometers for 0 cm soil maximum, average, and minimum temperatures are positioned horizontally on the soil surface. Half of these thermometers are buried in soil, and half are in air. If fully buried by snow, the thermometers were brought to snow surface, with half below the snow surface and half above the snow surface. We also used these measurements to develop empirical equations for daily mean soil surface temperature to identify which set of atmospheric variables should be used in the empirical equation sets (see Appendix A).

[14] This is not the first modeling effort to consider the effects of vegetation on soil temperature. It is well known that the aboveground vegetation affects soil surface temperature by intercepting downward shortwave radiation, shading the
soil, and trapping outgoing longwave radiation. Zheng et al. [1993] used $\exp(-0.5\text{LAI})$ to represent the effects of vegetation on 10 cm soil temperature, while Parton et al. [1998] used biomass in the equations of maximum/minimum soil surface temperature. In this study, we fitted equations, one with LAI and the other with $\exp(-0.5\text{LAI})$ as an independent variable using the 2005 part of the simulated data set. The fitted equations were tested using the 2006 part of the simulated data set. We also compared the performance of the soil surface temperature equation sets developed in this study to the use of air temperature as an estimate of soil surface temperature and the use of the algorithm used to estimate soil surface temperature in DayCENT [Parton et al., 1998], which is as follows:

$$
\begin{align*}
\text{tsmax} & = \frac{25.4}{1 + 18 \times e^{-0.2 \times \text{tsmax}}} \times (e^{0.0048 \times \text{B}} - 0.13) + \text{tsmax} \\
\text{tsmin} & = \text{tamin} + 0.006 \times \text{B} - 1.82 \\
\text{ts} & = K_{\text{mn}} \times \text{tsmin} + K_{\text{mx}} \times \text{tsmax},
\end{align*}
$$

where $\text{tsmax}$, $\text{tsmin}$, $\text{tamax}$, and $\text{tamin}$ are the daily maximum and minimum soil surface temperature and daily maximum and minimum air temperatures, respectively, but with units of °C; $K_{\text{mn}}$ and $K_{\text{mx}}$ are related to day length. When day length is less than 12 h, $K_{\text{mn}} > 0.5 > K_{\text{mx}}$ and vice versa. B is biomass (g/cm²). We used $B = 80\text{LAI}$, which was fitted using the field data from the Shule River Basin on the northeast edge of the Qinghai-Tibetan Plateau [Qin et al., 2013].

### 2.4. Validation of DOS-TEM Simulations of Soil Temperatures

[15] The equations from section 2.3 were then used in DOS-TEM to calculate soil surface temperatures of bare soil, which was used as a boundary condition for the subsequent soil temperature simulations. We ran the modified DOS-TEM at 21 meteorological stations on the Qinghai-Tibetan Plateau that have more than 20 years of atmospheric and soil measurements for bare soil. Among these 21 stations, 19 have silty clay soil, and 2 have clay soil [FAO/IAEA/ISRIC/ISSCAS/JRC, 2009]. The total depth of soil column

![Figure 3](image-url)  
**Figure 3.** Comparison of soil surface temperatures ($t_s$) for bare soil at the Xidatan flux tower site simulated by CLM3 for bare soil (line) with those estimated by an equation developed from measurements across the network of meteorological stations (with air temperature and radiation as independent variables; see Appendix A) (dots).

### Table 1. The Empirical Equations for Calculating Daily Mean Soil Surface Temperature ($t_s$) From Empirical Equations for Daily Maximum ($t_{\text{tsmax}}$, K) and Minimum ($t_{\text{tsmin}}$, K) Soil Surface Temperature based on CLM3 Simulations and Based on Meteorological Station Measurements for Bare Soil at the Meteorological Stations

<table>
<thead>
<tr>
<th>Equation</th>
<th>$t_{\text{tsmax}}$</th>
<th>$t_{\text{tsmin}}$</th>
<th>$t_s$</th>
<th>$r^2$</th>
<th>rmse</th>
</tr>
</thead>
<tbody>
<tr>
<td>Based on CLM3 without soil moisture</td>
<td>$0.65t_{\text{tsmax}} + 0.077r + 97.73$</td>
<td>$0.90t_{\text{tsmin}} + 31.16$</td>
<td>$0.36t_{\text{tsmax}} + 0.65t_{\text{tsmin}} - 4.88$</td>
<td>0.81</td>
<td>57.23</td>
</tr>
<tr>
<td>$t_{\text{ts}}$</td>
<td>$0.89t_{\text{tsmax}} + 0.062r - 76.53s + 51.31$</td>
<td>$0.91t_{\text{tsmin}} - 9.58s + 26.51$</td>
<td>$0.41t_{\text{tsmax}} + 0.63t_{\text{tsmin}} - 11.66$</td>
<td>0.97</td>
<td>8.04</td>
</tr>
<tr>
<td>$t_{\text{ts}}$</td>
<td>$0.86t_{\text{tsmax}} + 0.062r - 76.53s + 51.31$</td>
<td>$0.91t_{\text{tsmin}} - 9.58s + 26.51$</td>
<td>$0.41t_{\text{tsmax}} + 0.63t_{\text{tsmin}} - 11.66$</td>
<td>0.97</td>
<td>8.04</td>
</tr>
<tr>
<td>Based on CLM3 with soil moisture</td>
<td>$0.87t_{\text{tsmax}} + 0.021r + 46.79$</td>
<td>$1.06t_{\text{tsmin}} - 17.29$</td>
<td>$0.15t_{\text{tsmax}} + 0.82t_{\text{tsmin}} + 11.56$</td>
<td>0.67</td>
<td>97.16</td>
</tr>
<tr>
<td>Based on meteorological station measurements</td>
<td>$0.87t_{\text{tsmax}} + 0.021r + 46.79$</td>
<td>$1.06t_{\text{tsmin}} - 17.29$</td>
<td>$0.15t_{\text{tsmax}} + 0.82t_{\text{tsmin}} + 11.56$</td>
<td>0.92</td>
<td>22.26</td>
</tr>
</tbody>
</table>

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Two sets of equations are provided for the CLM3 simulations: those that do not consider soil moisture and those that consider soil moisture. The equations for $t_{\text{tsmax}}$ and $t_{\text{tsmax}}$ consider the following atmospheric variables: maximum ($t_{\text{tsmax}}$, K)/minimum ($t_{\text{tsmin}}$, K) atmospheric temperature (K), surface radiation ($r$, W/m²) from meteorological stations on the Qinghai-Tibetan Plateau, and volumetric soil water content ($s$, %) from CLM3 simulations. The equation sets for the CLM3 simulations were developed based on 5116 simulated data points in 2005, and the equation set for the meteorological station measurements was based on 346,834 pre-1990 measured data points. We evaluated the fits of the CLM3 equation sets with 5002 simulated data points in 2006 and evaluated the equation set for the meteorological station measurements with 254,778 post-1990 measured data points. $a$ and $b$ are slope and intercept of the fitted line between simulated/empirically estimated and measured soil temperature; $r^2$ is the square of the coefficient of determination, and rmse is the root mean squared error.
is about 50 m with the above mentioned soil texture distributed homogeneously in the top 1 m (seven layers) and underlain by ~4 m gravel (five layers) and ~45 m bedrock (nine layers).

[16] The monthly maximum, minimum, and mean air temperature; radiation; precipitation; vapor pressure; and relative humidity were calculated based on corresponding daily values. If the number of valid daily values of a variable was less than 20 in a month, then the monthly value of this variable was considered as missing, which was then filled with the multiyear mean of this variable of the same meteorological station.

[17] DOS-TEM was first run to equilibrium using the first-year data from available measurements as driving variables and then run with monthly climate over the period with measurements. The simulated soil temperatures at different depths were compared with the measurements.

2.5. Sensitivity Analysis of the Effects of Vegetation Cover Change on Soil Temperatures

[18] To investigate the effects of vegetation cover change on soil temperatures, we ran DOS-TEM with LAI from 0 to 3 with an interval of 1, using the same atmospheric driving variables from meteorological stations on the Qinghai-Tibetan Plateau. The temperature of the near-surface soil (usually within 20 cm of the surface) is important for many biogeochemical processes and is usually used in modeling studies to calculate soil respiration, e.g., in C-CLASS [Arain et al., 2002]. In this study, the simulated 10 cm soil temperatures at different depths were compared with the measurements.

2.6. Application of DOS-TEM to Investigate Dependence of Soil Erosion on Changes in Freeze-Thaws Cycles and Vegetation Cover

[19] The successful ability to consider the effects of vegetation cover on soil thermal dynamics in DOS-TEM makes it possible to use the model as a tool to investigate how interactions between vegetation cover and soil thermal dynamics influence other processes, such as the loss of carbon from the system through soil erosion. To use DOS-TEM in this demonstration application, we first calibrated the model for an alpine steppe site near Anduo (AD, 91°48′E, 32°28′N, 4871 m asl, Figure 1), using the equation set to estimate soil surface temperature for vegetated soil. The target values used

Table 2. The Empirical Equations for Calculating Daily Mean Soil Surface Temperature ($t_s$) From Empirical Equations for Daily Maximum ($t_{max}$, K) and Minimum ($t_{min}$, K) Soil Surface Temperature Based on CLM3 Simulations for a Range of LAI Values at the Meteorological Stations

<table>
<thead>
<tr>
<th>Equation</th>
<th>$a$</th>
<th>$b$</th>
<th>$r^2$</th>
<th>rmse</th>
</tr>
</thead>
<tbody>
<tr>
<td>Using LAI $t_{max}$</td>
<td>0.57$t_{max}$ + 0.03LAI + 0.09$r$ + 16.56</td>
<td>0.71</td>
<td>82.36</td>
<td>0.69</td>
</tr>
<tr>
<td>$t_{min}$</td>
<td>0.86$t_{min}$ + 0.65LAI + 38.37</td>
<td>0.96</td>
<td>11.37</td>
<td>0.96</td>
</tr>
<tr>
<td>$t_s$</td>
<td>0.41$t_{max}$ + 0.64$t_{min}$ − 15.45</td>
<td>0.97</td>
<td>9.86</td>
<td>0.94</td>
</tr>
<tr>
<td>Using exp(−0.5LAI)</td>
<td>$t_{max}$</td>
<td>0.78$t_{max}$ + 0.05 × exp(−0.5LAI)$r$ + 68.74</td>
<td>0.90</td>
<td>115.46</td>
</tr>
<tr>
<td>$t_{min}$</td>
<td>0.86$t_{min}$ + 0.65LAI + 38.37</td>
<td>0.96</td>
<td>10.78</td>
<td>0.96</td>
</tr>
<tr>
<td>$t_s$</td>
<td>0.47$t_{max}$ + 0.56$t_{min}$</td>
<td>0.92</td>
<td>20.70</td>
<td>0.91</td>
</tr>
</tbody>
</table>

Two sets of equations are provided: those that use LAI as an independent variable and those that use exp(−0.5LAI) and an independent variable. The equations for $t_{max}$ and $t_{min}$ consider the following atmospheric variables: maximum ($t_{max}$, K)/minimum ($t_{min}$, K) atmospheric temperature (K) and surface radiation ($r$, W/m²) from meteorological stations on the Qinghai-Tibetan Plateau. Simulated data in 2005 (30,870 data points) were used for deriving the empirical equations, and those in 2006 (30,136 data points) were used for evaluating the fit of the empirical equations. $a$ and $b$ are slope and intercept of the fitted line between calculated and simulated temperatures, respectively; $r^2$ is the square of the coefficient of determination, and rmse is the root mean squared error.

Figure 4. Comparisons between the evaluation data set of daily mean soil surface temperature ($t_s$-eval) and (a) the calculated values ($t_s$-calc) using the recommended equation set (the second equation set in Table 2), (b) using air temperature as the estimate for soil surface temperature ($t_a$), and (c) calculated using algorithm of DayCENT ($t_s$-cent).
are the same as those of Zhuang et al. [2010], with the exception that the C:N ratio of new product was set to 17 [He et al., 2006], and thus, the nitrogen uptake rate ($N_{\text{uptake}}$) is 5.6 g N m$^{-2}$ yr$^{-1}$. The model was then run at Wudaokou (WDL, 93°04'E, 35°12'N, 4626 m asl, Figure 1) and Tuotuohe (TTH, 92°33'E, 34°19'N, 4582 m asl) for verification. Finally, sensitivity tests were performed to investigate the effects of freeze-thaw-related erosion. The CRU TS3.0 data set (available from http://badc.nerc.ac.uk/view/badc.nerc.ac.uk__ATOM_dataent_1256223773328276) was used to drive DOS-TEM, with the mean monthly climate of 1901–1930 for the calibration run and equilibration run, the monthly climate of 1901–1930 for the spin-up run (120 years), and the monthly climate of 1901–2009 for the transient run. The TEM family of models uses GIRR $(0.251 + 0.509(1-\text{CLD}))$ to calculate downward solar radiation at the top of the vegetation; GIRR is the solar radiation at the top of atmosphere, and CLD is cloud cover (%). The calculated solar radiation is usually about one third to two thirds of the observations in summer. We assumed that only the soil organic C and N of the first soil layer (2 cm) can be eroded, and the erosion rate is the product of the maximum erosion rate and the function of the number of freeze-thaw cycles (NUMFTC). The maximum erosion rate was set to be 0 (no erosion), 1/100, 1/10, and 1/very strong erosion. The function for the number of freeze-thaw cycles was assumed to be a cumulative function of a normal distribution with $(\text{NUMFTC}/365 - 0.5)/(0.5 \times 0.5)$ as the independent variable, so when NUMFTC is 182 days, f(NUMFTC) = 0.5; f (NUMFTC) increases/decreases rapidly as NUMFTC days increase/decrease around 182 days. The daily positive/negative cumulative degree hour is needed for calculating the position of the thawing/freezing front. However, in DOS-TEM, only the daily maximum, mean, and minimum soil surface temperatures are available. We used part of the half-hour soil surface temperature data sets of CLM3 simulation from section 2.2 to determine the frequency distribution of soil surface temperatures in one day:

- 1. We calculate hourly soil surface temperature from half-hour data sets.
- 2. For each day, we evenly assign 100 intervals between daily minimum and mean soil surface temperatures.
- 3. We loop through hourly soil surface temperatures; if a hourly soil surface temperature is located in one of the 100 intervals, then the number of that interval increases by one.
- 4. We use the counts in the 100 intervals to determine the frequency distribution of soil surface temperature.

The same procedure was used to determine the frequency distribution of soil surface temperature between daily mean and maximum soil surface temperatures. The positive and negative cumulative degree hour is then calculated by using the frequency distribution and daily maximum, mean, and minimum soil surface temperatures. The remaining half-hour soil surface temperature data sets were used for validation. With the negative/positive cumulative degree hour, the location of the freezing/thawing front can be calculated. If the freezing/thawing front penetrates through 2 cm unfrozen/frozen ground, then it is considered a freeze-thaw cycle.

3. Results

3.1. Performance of CLM3 at the Xidatan Flux Tower

The performance of CLM3 to simulate near-surface soil temperature and moisture at the flux tower was good (Figure 2). The coefficient of determination ($r^2$) and root mean squared error (rmse) between simulated and measured 2 cm daily mean soil temperatures were 0.98 and 1.57, respectively, and those for 5 cm daily-averaged soil moistures were 0.84 and 0.05, respectively.

The soil surface temperatures simulated by CLM3 for bare ground at the Xidatan flux tower were comparable to those estimated for the flux tower by an empirical equation developed from measurements at the meteorological stations (Figure 3); the equation was driven by the same air temperature and radiation fields used to drive CLM3 (see equation $t_e = f(t_r)$ in Table A2 of Appendix A). The $r^2$ and rmse between simulated and empirically estimated soil surface temperatures were 0.90 and 2.75, and the slope $a$ and intercept $b$ were 1.11 and $-31.17$, respectively. The empirically estimated soil surface temperatures were less than those simulated by CLM3 in winter because the empirically estimated soil surface temperatures were for bare ground without snow cover, while the simulated soil surface temperatures were

Table 3. Comparisons Between Monthly Soil Temperatures Simulated by DOS-TEM at Different Depths With Measured Values at Bare Sites of Meteorological Stations on the Qinghai-Tibetan Plateaua

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>num</th>
<th>Equations</th>
<th>$a$</th>
<th>$b$</th>
<th>$r^2$</th>
<th>rmse</th>
</tr>
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<tr>
<td>5</td>
<td>3721</td>
<td>1</td>
<td>1.02</td>
<td>0.68</td>
<td>0.98</td>
<td>1.72</td>
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<tr>
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<td>2</td>
<td>0.86</td>
<td>0.32</td>
<td>0.97</td>
<td>2.00</td>
<td></td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>0.77</td>
<td>−4.30</td>
<td>0.69</td>
<td>8.23</td>
<td></td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>0.83</td>
<td>−2.45</td>
<td>0.95</td>
<td>4.62</td>
<td></td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>0.88</td>
<td>1.26</td>
<td>0.97</td>
<td>1.78</td>
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<tr>
<td>20</td>
<td>3708</td>
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<td>0.97</td>
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<tr>
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<td>0.97</td>
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<td>−0.25</td>
<td>0.88</td>
<td>2.31</td>
<td></td>
</tr>
</tbody>
</table>

*a* num is the number of pairs of monthly soil temperature data used for comparison; $a$ and $b$ are slope and intercept of the fitted line between simulated and measured; $r^2$ is coefficient of determination; and rmse is root mean squared error. Equation 1 is $t_e = f(t_r)$ all in Table A2, 2 and 3 are the first two sets of equations in Table 1 without and with soil moisture as independent variable, 4 is using air temperature directly, and 5 is the proposed equation set (second in Table 2).
below the snow cover and considered the effects of snow cover insulation on soil surface temperature.

3.2. Soil Surface Temperature Equations Developed from CLM3 Simulations

[28] Based on the CLM3 simulations for bare soil at the meteorological stations, we developed empirical equation sets for estimating daily mean soil surface temperature that had did and did not consider simulated soil moisture as an independent variable (Table 1). These equations used daily maximum air temperature, daily minimum air temperature, and radiation as independent variables. The choice of developing these equations based on atmospheric temperature and radiation was based on analyses that considered whether wind speed and precipitation would add more information to empirical estimates of soil surface temperature (see Appendix A). The comparison of the performance of these two data sets was similar in that they explained 96% of the variation in the measurements, but the root mean squared error was slightly less for the equation set that used soil moisture as an independent predictor variable. The performance of these equation sets was also similar to that of analogous equation sets developed from measurements at the meteorological stations (Table 1). Therefore, we felt confident that the CLM3 simulations could be used to derive equation sets that consider vegetation cover as an independent variable.

[29] Based on the CLM3 simulations we conducted at the meteorological stations for a range of LAI values, we developed two empirical equation sets for estimating daily mean soil surface temperature, one of which considered LAI as an independent variable and the other considered exp(−0.5LAI) as an independent variable (Table 2). The performance of the equation set using LAI as an independent variable was slightly better than that of the equation set that used exp(−0.5LAI) as an independent variable (94% versus 91% of variance explained and a 2.19 versus 2.66 root mean squared error). However, in the equation set that uses LAI as an independent variable, the coefficient of LAI was positive in the calculation of daily maximum soil surface temperature (i.e., increase of daily maximum soil surface temperature with increase of vegetation cover), which is not realistic. Therefore, in subsequent analyses in this study we employed the equation set that used exp(−0.5LAI) as an independent variable. Although this equation set slight underestimates soil surface temperature in summer (Figure 4a), it performs substantially better than just using air temperature as a surrogate for estimating soil surface temperature, which underestimates soil surface temperature in all seasons (Figure 4b). The equation set also performs substantially better than the algorithm used by DayCENT for estimating soil surface temperature, which overestimates soil surface temperature in summer and underestimates it in the winter (Figure 4c).

3.3. Performance of DOS-TEM Using the Recommended Equation Set for Estimating Soil Surface Temperature for Bare Soil (LAI = 0)

[30] For the simulation of soil temperature at different depths for bare soil across the network of meteorological stations, the incorporation of the second equation set from Table 2 into DOS-TEM generally performed better than the
use of four other equations for soil surface temperature (see results for equation (5) in Table 3 versus those for other equations, see also Figure 5). The performance of DOS-TEM using any of the equations considered in Table 3 was better near the soil surface than at deeper depths. For example, the use of the second equation set of Table 2 resulted in 97% variance explained and a root mean squared error (rmse) of 1.78 at 5 cm, but at 80 cm explained 88% of the variance and a rmse of 2.31. The performance of the DOS-TEM simulation using air temperature as an estimate of soil surface temperature was much worse than that using the second equation set in Table 2, with rmse ~2.5 times bigger at all depths (results not shown). The incorporation of the equation set that depended on soil moisture simulated by CLM3 (i.e., the second equation set in Table 1) into DOS-TEM resulted in particularly poor simulations of soil temperature at different depths (see results for equation (3) in Table 3).

3.4. Effects of Vegetation Cover on Soil Temperatures Simulated by DOS-TEM

[31] The monthly mean of the daily maximum soil surface temperature simulated by the recommended equation set used by DOS-TEM across the network of meteorological stations on the Qinghai-Tibetan Plateau decreased with increased vegetation cover (Figure 6). The monthly mean of the daily maximum soil surface temperature for bare soil (LAI = 0) was about 5°C higher than those for LAI = 1 during the growing season from May to September, and was 2°C–3°C higher during the nongrowing season. The monthly mean of the minimum soil surface temperature increased with increased vegetation cover, but the differences were usually smaller than 1°C across the range of LAI considered (Figure 6). At 10 cm depth, the difference in soil temperature between bare soil (LAI = 0) and vegetated soil with LAI = 3 was about 2°C during growing seasons, but less in nongrowing seasons.

3.5. Effects of Surface Erosion Associated With Freeze-Thaw Cycles

[32] The simulated C and N pools and net primary production were comparable to measurements at Wudaoliang and Tuotuohe on the Qinghai-Tibetan Plateau (Figure 7) [Luo et al., 2004; Luo et al., 2005]. On the Qinghai-Tibetan Plateau, freezing and thawing are frequent, e.g., at Tuotuohe, the number of freeze-thaw cycles ranges 60–150 days/yr. Freezing/thawing has the potential to have significant impacts on alpine grassland ecosystems (Figure 8). In regions with high erosion potential (i.e., large wind and runoff), the C and N of the top soil layers are lost quickly, which affects vegetation growth and reduces vegetation cover. This causes the frequency of freeze-thaw to be enhanced (i.e., number of freezing/thawing cycles increases) and makes soil more prone to erosion. However, in regions with low erosion potential, the effect of freeze-thaw cycles is minor.
4. Discussion

4.1. Evaluation of the Methodology Developed in This Study

[31] With adequate half-hour atmospheric forcing, the land surface model CLM3 performed well in simulating near-surface soil temperature through solving surface energy balance equations. Our approach therefore was to use CLM3 simulations to derive empirical relationships for daily soil surface temperature that could be incorporated into environmental models that require daily or pseudo-daily inputs.

[34] Our analysis of a suite of atmospheric driving variables for the development of empirical relationships led us to consider only air temperature and radiation and to reject the use of wind speed and precipitation (see Appendix A). Daily mean air temperature has often been used as a direct estimate for daily soil surface temperature in ecosystem models, but our analysis indicates that its use on the Qinghai-Tibetan Plateau would underestimate daily mean soil surface temperature. Radiation is another important factor to consider because surface soil temperature can vary greater between sunny and cloudy days. For example, in a measurement plot of the Shule River Basin, which is located on the northeast edge of the Qinghai-Tibetan Plateau, the difference of soil surface temperature of bare soil between a sunny day (9 August 2011) and a cloudy day (10 August 2011) was more than 20°C (Figure 9). Equations with radiation can also be used in landscape-scale ecosystem modeling studies to account for the effects of topography on soil surface temperature that may have consequences for ecosystem processes [Kane et al., 2007]. Although wind speed affects evapotranspiration and reduces soil surface temperature, the inclusion of wind speed in equations that also considered air temperature and radiation as independent variables did not improve the overall performance of those equations in estimating daily soil surface temperature. Furthermore, wind speed is usually not provided in some global data sets, e.g., the CRU TS3.0 data set. We do not recommend that wind speed be considered in calculating daily soil surface temperature in ecosystem models applied to the Qinghai-Tibetan Plateau.

[35] Soil surface temperature is sensitive to the change of volumetric soil water content. For the equation set we derived that includes CLM3-simulated soil moisture as an independent variable, an increase of volumetric water content of the surface soil layer by 10% will decrease daily maximum soil surface temperature by 7.7 K and daily minimum soil surface temperature by ~1 K. Generally, in cold regions, ecosystem models are able to more accurately simulate soil temperature than soil moisture [Yi et al., 2009]. Our use of the equation set with soil moisture as an independent variable performed poorly compared to other equations for soil surface temperature. Future studies need to make improvements in reducing the uncertainties in soil moisture simulation in cold regions. In the meantime, we suggest that soil moisture not be included as an independent variable in calculating daily soil surface temperature.

[36] Based on the CLM3 simulations, we were successful at developing a strongly performing empirical relationship for daily soil surface temperature that used air temperature, radiation, and leaf area index as independent variables. Our

Figure 8. The comparisons of sensitivity runs with different maximum erosion rates, i.e., control (no erosion, black), 1/100 (red), 1/10 (green), and 1 (very strong erosion, blue) for (1) vegetation carbon (VEGC, gC m\(^{-2}\)), (2) net primary production (NPP, gC yr\(^{-1}\) m\(^{-2}\)), (3) soil carbon (SOLC, gC m\(^{-2}\)), and (4) number of daily freezing/thawing cycles (NUMFTC) at Tuotuohe (TTH).
incorporation of this relationship for daily soil surface temperature into DOS-TEM indicated that when soil surface temperature is properly simulated, soil temperatures at different depths can also be well simulated by DOS-TEM.

4.2. Evaluation and Comparison With Other Methodologies

There have been several ecosystem modeling studies on the Qinghai-Tibetan Plateau, e.g., by TEM 5.0, which used air temperature as an approximation for soil surface temperature, and CENTURY 4.5. Some studies did not compare simulated soil temperatures with measurements [Zhang et al., 2007], some only compared 20 cm soil temperatures [Tan et al., 2010], and some only compared soil temperatures of a few stations [Zhuang et al., 2010]. Our analyses made use of more data, and we achieved improved results in simulating soil temperatures at different depths across a large network of meteorological stations compared with the above mentioned studies.

Our analysis showed the use of air temperature for soil surface temperature causes an underestimation of soil surface temperature in both summer and winter on Qinghai-Tibetan Plateau. For the algorithm used by DayCENT, our analysis showed that the algorithm overestimated soil surface temperature in summer and underestimated it in winter. The DayCENT algorithm has some features in common with our algorithm in that daily soil surface temperature is a function of daily maximum and minimum soil surface temperature, which depend on air temperature and biomass. In our algorithms for both bare and vegetated soil, the daily minimum soil surface temperature played a more important role than maximum soil surface temperature in determining daily mean soil surface temperature. Figure 10 presents a simple example with the 10 day mean diurnal soil surface temperatures for both winter (January) and summer (August) of Xidatan, and corresponding sinusoidal temperatures with the same temperature ranges. It is obvious that soil surface temperatures increased more quickly than the sinusoidal curve in the morning and decreased more quickly in the afternoon. This phenomenon is due to the thin atmosphere over the Qinghai-Tibetan Plateau, as shortwave radiation can pass through the thin atmosphere efficiently to heat the land surface, and the temperature of land surface increases quickly in the morning; the longwave radiation can also pass through the thin atmosphere quickly, and the temperature of land surface decreases quickly in the afternoon. In the DayCENT

Figure 9. (a, b) Conventional pictures taken around 3 P.M. Beijing time on 9 and 10 August, which were sunny and cloudy, respectively, on Shule River Basin on the northeast edge of the Qinghai-Tibetan Plateau. (c, d) Corresponding thermal infrared pictures taken with Testo Thermal Imager 882 (Testo Inc., Germany).

Figure 10. Comparisons between 10 day mean diurnal soil surface temperature (solid line) and corresponding sinusoidal temperature (dashed line) with the same daily range for (a) January and (b) August at Xidatan.
algorithm, the parameters $K_{mn}$ and $K_{mx}$ defined in section 2.3 are related to day length, which results in $K_{mn}$ being greater than 0.5 in winter and less than 0.5 in summer. Taking Xidatan as an example, the day length ranges from 10 to 14 h, which results in $K_{mn}$ ranges from 0.65 to 0.35 and $K_{mx}$ ranges from 0.35 to 0.65. In contrast, the parameters in our algorithm are analogous to $K_{mn}$ and $K_{mx}$ values in which $K_{mn}$ is always greater than $K_{mx}$. This is the primary reason why applying the DayCENT algorithm on the Qinghai-Tibetan Plateau causes an overestimation of soil surface temperature in summer.

4.3. Effects of Aboveground Vegetation on Soil Surface Temperature

Vegetation cover has a significant impact on soil surface temperature on the Qinghai-Tibetan Plateau. For example, the difference between surface temperatures of bare soil and the temperature at the top of the vegetation canopy can be about 20°C in a sunny day (Figure 9). Since the temperature underneath the canopy is usually less than the temperature of canopy under strong solar radiation conditions, the difference in soil surface temperature between bare soil and vegetated soil can be even greater.

Overgrazing is one of the most important disturbances to alpine grassland ecosystems on the Qinghai-Tibetan Plateau [Cui and Graf, 2009]. Reduction of vegetation cover, in addition to directly exposing surface soil to erosion, would increase freeze and thaw intensity, reduce the soil stability, and intensify soil erosion, which is currently a big problem of the Qinghai-Tibetan Plateau [Zhang et al., 2005]. Loss of soil carbon and nitrogen would further reduce soil fertility and subsequently reduce alpine grassland productivity and vegetation cover. Modeling studies should take this disturbance and associated feedbacks into account when projecting the response of alpine grassland ecosystems to climate change. In this study, our development of a methodology allowed us to conduct a preliminary first-order analysis to investigate this issue. The results of this analysis suggest that (1) soil erosion rates can depend substantially on freeze-thaw cycles and vegetation cover; (2) there may be strong feedbacks among soil erosion rates, freeze-thaw cycles, and vegetation cover; and (3) overgrazing has the potential to push the system beyond a tipping point from which it cannot recover. This analysis indicates that there is a need to better evaluate the strength of these feedbacks and what levels of overgrazing may drive the system beyond a tipping point. Therefore, additional field and laboratory research need to be conducted to develop more robust algorithms that relate soil erodibility with freeze-thaw cycles and that relate soil erosion with erodibility and climate conditions.

5. Conclusions

This study evaluated the effects of atmospheric factors and vegetation cover on soil surface temperature by integrating the measurements from meteorological stations on the Qinghai-Tibetan Plateau and simulations from a sophisticated land surface model. Results showed that air temperature, radiation, and vegetation cover (LAI) should be considered when calculating soil surface temperature. The equation set we developed for use in ecosystem models performed better at estimating soil surface temperature on the Qinghai-Tibetan Plateau than those currently used in ecosystem models. DOS-TEM performed well when soil surface temperatures were calculated with the equation set we developed in this study. Sensitivity tests using DOS-TEM with different vegetation covers showed that a decrease of vegetation cover would cause an increase in the daily maximum soil surface temperature and a decrease in the daily minimum soil surface temperature, and thus increase the daily temperature range. The daily mean soil surface temperature would also increase if vegetation cover decreases.

The use of the equation set we developed by DOS-TEM allowed us to conduct a preliminary analysis of factors that control soil erosion rates on the Qinghai-Tibetan Plateau. This analysis indicated that changes in vegetation cover would have significant effects on soil thermal regimes and on surface erosion associated with freeze-thaw cycles, which would further adversely affect alpine grassland ecosystems. Thus, the effects of vegetation cover change should be considered in modeling studies that assess how climate change and changes in grazing regimes affect permafrost and the function and structure of alpine grassland ecosystems on the Qinghai-Tibetan Plateau.
Appendix A: An Analysis of Alternative Soil Surface Temperature Equations Based on Measurements Across a Network of Meteorological Stations

[43] We conducted an analysis of which atmospheric variables needed to be included in the equations we developed to estimate soil surface temperature from CLM3 simulations. Our approach in this analysis was to develop equations based on measurement of soil surface temperature across the network of meteorological stations on the Qinghai-Tibetan Plateau. Equations for soil surface temperature were developed using measured daily mean soil surface temperature (t_s, K), atmospheric temperature (t_a, K), solar radiation (r, W/m²), and wind speed (w, m/s) before 1990 and evaluated using data after 1990 (including 1990). Only a few stations measured downward solar radiation; thus, we calculated it using data after 1990 (including 1990). Only a few stations measured downward solar radiation; thus, we calculated it following Yang et al. [2006b]. These equations were divided into two categories. One category used t_a as a response variable (hereafter the n-factor category), whereas the other used t_s (hereafter t_s category). Altogether, we developed 12 sets of equations for soil surface temperature (i.e., t_s/t_a = f(w), t_s/t_a = f(r), t_s/t_a = f(w, c), t_s/t_a = f(r, w, c)).

[44] Tables A1 and A2 report the comparisons between measured and calculated daily mean soil surface temperatures using fitted equations of different categories. The coefficients of determination (r²) and root mean squared errors (rmse), ranged 0.94–0.96 and 1.98–2.48 (K); the slopes (a) and intercepts (b) of fitted lines ranged 0.81–0.95 and 13.42–53.86 (K), respectively. Generally speaking, the t_s category equations performed better than n-factor category equations (i.e., a were closer to 1, b were closer to 0, r² were greater, and rmse were smaller). Equations that used air temperature and radiation performed better than those that used air temperature alone. Furthermore, the addition of wind speed and precipitation did not improve the ability to estimate daily soil surface temperature. The differences were small between equations fitted with warm and cold season measurements separately and with warm and cold season measurements together.

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