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The effect of snow: How to better model ground surface temperatures

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ABSTRACT

We present an inverse modeling approach for reconstructing the effective thermal conductivity of snow on a daily basis using air temperature, ground temperature and snow depth measurements. The method is applied to four sites in Alaska. To validate the method we used measured snow densities and snow water equivalents. The modeled thermal conductivities of snow for the two interior Alaska sites have relatively low values and reach their maximum near the end of the snow season, while the conductivities at the two sites on the Alaskan North Slope are higher and reach their maximum earlier in the snow season. We show that the reconstructed daily thermal conductivities allow for more accurate modeling of ground surface temperatures when compared to applying a constant thermal conductivity for the snow layer.

1. Introduction

Through its role as an insulator, snow cover plays a key role in the Arctic climate system by controlling heat exchanges between the atmosphere and the ground surface. This strongly influences the dynamics of the active layer and underlying permafrost (e.g., Goodrich, 1982; Ling and Zhang, 2004; Sazonova and Romanovsky, 2003; Shiklomanov and Nelson, 1999; Zhang, 2005). Warming and thawing of near surface permafrost have already been observed in parts of the Arctic, and permafrost degradation is expected to continue through the 21st century, with significant impacts on infrastructure and ecosystems (Callaghan et al., 2011; Instanes and Anisimov, 2008; Oberman, 2008). A number of studies point to the potential for a significant positive climate feedback related to carbon release from thawing permafrost (Grosse et al., 2011; Schaefer et al., 2011). Improving the veracity of projected changes in the active layer and permafrost conditions requires better parameterization of snow thermal properties, in particular, snow thermal conductivity. While often assumed to be constant throughout the entire snow season, in reality snow thermal conductivity depends on many factors, such as air and snow temperatures, snow density, and grain structures, and hence varies with time and position within the snow layer (Sturm et al., 1997).

Past studies of snow thermal conductivity have made use of field observations, laboratory experiments, and theoretical frameworks (e.g., Abel, 1893; Fukusako, 1990; Mellor, 1977; Pitman and Zuckerman, 1967; Sturm et al., 1997, 2002; Yen, 1962). Brun et al. (2013) and Domine et al. (2013) recently employed physically-based approaches making use of outputs from atmospheric reanalyses to simulate snowpack properties and soil temperatures.

Here, we present an approach to simulate snow "effective" thermal conductivities on a daily basis using the Geophysical Institute Permafrost Laboratory (GIPL) numerical transient model (Jafarov et al., 2012; Nicolsky et al., 2007; Sergueev et al., 2003). Effective conductivity (hereafter simply referred to as conductivity) includes the combined effect of conduction through the ice grains, conduction through the air in the void spaces, and radiative exchange across the void spaces (Anderson, 1976; Marks and Dozier, 1992; Morin et al., 2010). We use an inverse modeling approach originally introduced by Tipenko and Romanovsky (2002), and similar to that used by Sergienko et al. (2008).

The GIPL model simulates ground temperatures and the seasonal freeze/thaw layer dynamics, and has been successfully validated against ground temperature measurements in shallow boreholes across Alaska (Jafarov et al., 2013; Nicolsky et al., 2009; Romanovsky and Osterkamp, 2000). The model incorporates the effects of air temperature, snow, soil moisture and multi-layered soil thermal properties (Nicolsky et al., 2007; Sergueev et al., 2003). We use as model input measurements collected from four permafrost monitoring stations in Alaska, including snow depth, air temperatures and ground temperatures and obtain time series of snow thermal conductivity over the entire snow season. We show that the obtained thermal conductivity values improve the simulation of the ground surface temperature dynamics for the entire snow season.

Throughout the manuscript, we use the terms "estimated" and "reconstructed" interchangeably. The term "estimated snow conductivity" refers to the obtained snow conductivity as a result of using the inverse

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method, whereas the term "reconstructed" refers to the estimated snow conductivity after a moving-average filter is applied. The estimated and reconstructed values are marked by "ˆ" and "˘" signs, correspondingly.

2. Method

2.1. Physical model

The GIPL model solves the 1-D heat equation with phase changes (Carslaw and Jaeger, 1959):

$$C\frac{\partial T(x,t)}{\partial t}+L\frac{\partial \theta(x,T)}{\partial T}\frac{\partial T(x,t)}{\partial t}=\frac{\partial}{\partial x}\left(k\frac{\partial T(x,t)}{\partial x}\right), \quad (1)$$

where $T(x,t)$ is the temperature and L [Jm^{-3}] is the volumetric latent heat fusion of water. Here, t stands for time and $x\in(x_u,x_l)$ is the spatial variable with the ground surface at $x=0$. The upper boundary $x_u=x_u(t)$ depends on time in order to track the evolution of snow cover. The quantity $x_u(t)$ is equal to the snow cover depth when snow is present, or is zero otherwise. The lower boundary x_l is fixed and represents a certain depth below the active layer. Eq. (1) is complemented with boundary conditions $T(x_u,t)=T_{air}$ and $T(x_l,t)=T_l$, where T_{air} and T_l are observed temperatures at the ground (snow) surface $x=x_u(t)$ and at the depth $x=x_l$, respectively. In permafrost models driven with data at daily time step, air temperature at the snow surface is assumed to be the same as the near-surface air temperature (Ling and Zhang, 2004; Westermann et al., 2013). Additionally, we supply the initial conditions $T(x,0)=T_0(x)$, where $T_0(x)$ is the temperature at $x\in[0,x_l)$ at time $t=0$. The volumetric water content $\theta(x,T)$ for ground material $0\le x<x_l$ in Eq. (1) is defined as:

$$\theta(x,T)=\eta(x)\phi(T,x), \quad \phi(T,x)=\begin{cases}1, & T\ge T_*\\ |T_*|^b|T|^{-b}, & T<T_*\end{cases}, \quad (2)$$

where $\phi(T,x)$ represents the pore liquid water fraction, $\eta(x)$ is the soil porosity, T_* is the so-called freezing point depression, and b is a dimensionless parameter obtained from unfrozen water curve fitting (Romanovsky and Osterkamp, 2000). The volumetric water content $\theta(x,T)$ in Eq. (2) for a snow layer is equal to zero, so $\theta(x,T)=0$ for $x_u\le x<0$. The quantities $k=k(x,T)$ [Wm^{-1} K^{-1}] and $C=C(x,T)$ [Jm^{-3} K^{-1}] are thermal conductivity and the volumetric heat capacity, respectively, and are defined as follows:

$$C=C_s, \quad k=k_s, \quad x_u<x<0, \quad (3)$$

$$C=C_t\phi+C_f(1-\phi), \quad k=k_t^{\phi}k_f^{1-\phi}, \quad 0\le x<x_l \quad (4)$$

where $C_s=C_s(t)$ and $k_s=k_s(t)$ are the volumetric heat capacity and bulk thermal conductivity of snow, respectively. The quantities marked with the subscripts "t" and "f" represent effective thermal properties of the ground material for the thawed and frozen states, respectively. The modeled soil column consists of several layers, each having its own thermal properties C_t, C_f, k_t, and k_f. Unlike the thermal properties of snow, C_t, C_f, k_t, and k_f are assumed to be time-independent and to vary only with depth.

The snow layer is represented as a homogeneous substance with changing thickness during the snow season. To obtain the temperature distribution within the snow layer, the GIPL model solves the heat Eq. (1), where $k_s=k_s(t)$ and $C_s=C_s(t)$ are snow thermal parameters that may change in time but do not change spatially. To simulate the measured ground surface temperatures, it is therefore important to assign proper daily snow thermal properties.

It is common to define snow heat capacity and thermal conductivity with a function that depends on snow density, (e.g. Abel, 1893; Anderson, 1976; Ostin and Andersson, 1991; Yen, 1981). Goodrich (1982) and Douville et al. (1995) represented snow heat capacity as a linear function of snow density in order to calculate ground temperatures. Given that the differences between the snow heat capacity equations used by Goodrich (1982) and Douville et al. (1995) are not significant, we used Douville et al. (1995):

$$C_s=C_i\cdot\rho_s/\rho_i, \quad (5)$$

where C_i is the heat capacity of ice, and $\rho_s=\rho_s(t)$ and ρ_i [g·cm^{-3}] are densities of snow and ice respectively.

There are a variety of empirical methods for estimating the thermal conductivity of snow k_s as a function of density, summarized by Yen (1969), and Sturm et al. (1997), and more recently by Calonne et al. (2011), and Riche and Schneebeli (2013). Sturm et al. (1997) derived an empirical relationship between snow density and conductivity based on an overview of prior work and his own field observations. More recent studies (e.g. Calonne et al., 2011; Riche and Schneebeli, 2013) introduce empirical relationships that better correspond with the observed data. Since the difference between the formulas derived from these two recent works is not significant we chose the empirical relationship by Calonne et al. (2011):

$$k_s=2.5\rho_s^2-0.123\rho_s+0.024. \quad (6)$$

In the current model formulation both the thermal conductivity and the heat capacity of snow depend on snow density. We run the GIPL model by using a vertical domain starting from the ground/snow surface $x_u(t)$ to a depth of $x_l=1$ m, with a 0.01-m grid resolution.

2.2. Data assimilation technique

The intent of data assimilation is to minimize the difference between the simulated and measured ground surface temperature. Note that the computations of ground surface temperature rely on modeling the heat exchange below the ground surface. Nicolsky et al. (2007, 2009) showed that the thermal properties of ground material can be estimated by using only the surface and sub-surface ground temperatures. Therefore, we assume that the thermal properties k_t, k_f, C_t, C_f, the soil porosity η, and the parameterization of the unfrozen water content T_*,b are known and can be utilized to simulate temperature dynamics below the ground surface. Once the sub-surface thermal properties are found, the simulated ground surface temperature effectively depends only on the thermal properties of the snow cover. The latter can be parameterized by snow density ρ_s according to Eqs. ((5)–(6)).

To find the density ρ_s, we use an inverse modeling approach outlined by Tarantola (2005). In such an approach, an objective function or cost function J_{BT} can be defined as

$$J_{BT}(\rho_s)=J_1(\rho_s)+J_2(\rho_s), \quad (7)$$

where J_{BT} is the cost function J defined by Beck and Arnold (1977) ("B") and Tikhonov and Leonov (1996) ("T"), and

$$J_1(\rho_s)=\frac{1}{\delta T^2}\|T_m-T(\rho_s)\|^2=\frac{1}{\delta T^2}\frac{1}{t_f}\int_0^{t_f}(T_m(\tau)-T(0,\tau;\rho_s(\tau)))^2d\tau. \quad (8)$$

Here, J_1 is the discrepancy between the observed $T_m=T_m(t)$ and simulated $T(\rho_s)=T(0,t;\rho_s(t))$ ground surface temperatures. The latter is computed according to Eq. (1) with a time-varying snow density $\rho_s(t)$, when snow covers the ground surface over the period $[0,t_f]$. The second term in Eq. (7) is the regularization term

$$J_2(\rho_s)=\frac{1}{\delta\rho^2}\|\rho_s-\tilde{\rho}_s\|^2=\frac{1}{\delta\rho^2}\frac{1}{t_f}\int_0^{t_f}\left(\rho_s(\tau)-\tilde{\rho}_s(\tau)\right)^2d\tau, \quad (9)$$