Estimation of Hydraulic Properties in Permafrost-Affected Soils Using a Two-Directional Freeze-Thaw Algorithm

W. Robert Bolton
Alfred Wegener Institute for Polar and Marine Research
Julia Boike
Alfred Wegener Institute for Polar and Marine Research
Pier Paul Overduin
Alfred Wegener Institute for Polar and Marine Research

Abstract

In this study, a two-directional freeze-thaw (TDFT) algorithm is used to estimate the hydraulic conductivity and storage capacity of permafrost-affected soils. The physically based TDFT algorithm is based upon the Stefan equation. The algorithm is driven by both the surface temperature and temperature at a specified depth and uses physical properties of the soil (bulk density, porosity, soil moisture, organic and mineral fraction, and freezing temperature of water) as input variables. The TDFT is tested in a wet valley bottom and along a relatively dry hill slope in the Imnavait Creek Watershed, Alaska. Results indicate that the timing of thaw/freezeback periods, the maximum thaw depth, and latent heat effects are accurately simulated. Using the TDFT algorithm to derive hydraulic variables is an improvement from the previous methods used to represent the active layer in hydrological modeling studies.

Keywords: active layer; hydraulic conductivity; hydrologic modeling; permafrost; porosity.

Introduction

In (sub-) arctic environments, most biologic, geomorphic, hydrologic, and ecologic processes take place within the active layer —the thin soil layer above the permafrost that seasonally freezes and thaws (Kane et al. 1991, Hinzman et al. 1998, Woo 2000). The depth of the active layer, as well as the rate of seasonal freezing and thawing, is dependent upon a number of factors including soil properties (thermal conductivity, bulk density, moisture content, vegetation type/thickness), meteorologic conditions (air and surface temperature, presence of snow cover), and disturbance (either anthropogenic or natural disturbance such as wildfire) (Woo 1986). As each of these factors are spatially and temporally variable, the position of the freeze-thaw interface is also spatially and temporally variable (Woo & Steer 1983).

Hydrologically speaking, the (sub-) arctic environment is unique in that the thermal and hydrologic regimes of the soil (permafrost versus non-permafrost) can vary greatly over short horizontal spatial scales, with depth, and over time. It is well documented that the hydraulic conductivity of ice-rich permafrost soils can be several orders of magnitude lower than their unfrozen counterparts (e.g., Burt & Williams 1976, Kane & Stein 1983). Furthermore, ice-rich conditions at the freeze-thaw interface significantly reduce the permeability of the soil, effectively limiting the soil water capacity of the soil (Dingman 1975, Woo 1990).

However, the point and hill slope scale understanding of permafrost soils, specifically the effect of the freezing and thawing of the active layer on hydrologic proceses, have not been adequately or systematically incorporated into meso-scale hydrologic models (Vörösmarty et al. 1993). Representation of the active layer in previous modeling studies includes switching soil properties for the winter and summer periods (Sand & Kane 1986) and use of a simple square root of time function to estimate the active layer depth (Zhang et al. 1998, Schramm et al. 2007). Thermal models are absent in most hydrologic models because of computational time requirements and complexities in model coupling.

Fox (1992) introduced a physically-based one-directional freeze-thaw (ODFT) algorithm for estimating the position of the freeze-thaw interface. The ODFT is driven by surface temperature. Woo et al. (2004) modified the ODFT by inverting the equations and driving the algorithm in two-directions by using temperatures at the surface and at a specified depth in the soil column. This two-directional freeze-thaw (TDFT) algorithm improves the representation of the freeze-thaw interface during the freezeback period.

The objective of this paper is to present a method for estimating hydraulic conductivity and effective porosity (proxy for soil storage capacity) using the TDFT algorithm.

Methods

Overview of the TDFT algorithm

The main difference between frozen and thawed soils is the difference in hydraulic conductivity and storage capacity, both a function of amount of pore ice in the soils (Woo 1986). In our conceptualization of the arctic hydrologic regime, frozen soils are represented with a very low hydraulic conductivity, while thawed soils within the active layer are represented with a larger hydraulic conductivity. Effective porosity (P_{eff}) is used as a proxy for storage capacity. In frozen soils, the presence of pore ice reduces the effective porosity of the soil.

The freeze-thaw interface in the active layer is determined using the TDFT algorithm. The physically-based TDFT algorithm is based upon the Stefan's equation of heat conduction (Jumikis 1977). One of the assumptions of the Stefan equation is that sensible heat effects are negligible. This assumption typically holds in soils with high moisture contents (Lunardini 1981). The algorithm requires inputs for each soil layer: specified depth, bulk density, porosity, fraction of organic and mineral soils, soil moisture content, the threshold for freezing the soil moisture, and minimum unfrozen water content in frozen soils. For each soil layer, the thermal conductivity and the energy required to freeze or thaw that soil layer is calculated. The algorithm is driven by the surface temperature and temperature at the bottom of the soil column. At each time step, the energy available to freeze/thaw the soil is determined using a simple 'degree-day' formulation. The amount of energy available is compared to the amount of energy required to freeze/thaw a soil layer. The total amount of freezing or thawing of the soil layers is determined by the total amount of energy available from the surface and at the bottom of the soil column. The procedures for determining the thermal conductivity, the energy available for freezing/thawing the soils, and determination of the total freezing and thawing of the soil layers are described in detail in Woo et al. (2004). In our formulation, the TDFT is slightly modified such that soil thawing from the bottom of the soil column is not allowed.

Estimation of hydraulic properties

The next steps are to estimate the freeze-thaw interface, the hydraulic conductivity, and the effective porosity. After each time step, the total thawed thickness and frozen thickness (D_t and D_f, respectively) of each soil layer are determined. The position of the freeze-thaw interface is estimated for each soil layer, from the surface downward, by evaluating the thermal condition of the neighboring soil layers. For example, if soil layer X_{+1} is completely frozen ($D_t = 0.0$), while soil layer X_{-1} is completely thawed (D_t = soil layer thickness), then we assume that the active layer is developing and the position of the freeze-thaw boundary is located at D_t depth from the top of the soil layer, X. By evaluating each soil layer after each time step, multiple freeze-thaw interfaces within the soil column are possible.

Once the D_t and D_f are determined for each soil layer, the hydraulic conductivity for each soil layer is estimated using a simple weighting function. For each soil layer, the frozen and thawed hydraulic conductivity are specified. If the entire soil layer is either frozen or thawed, the appropriate hydraulic conductivity is assigned to that soil layer. If the freeze-thaw interface is located within a soil layer, the horizontal (K_H) and vertical hydraulic (K_V) conductivities are determined using simple weighting functions:

$$K_H = \frac{K_f D_f + K_t D_t}{D} \tag{1}$$

$$K_V = \frac{D}{\left[\dfrac{D_f}{K_f} + \dfrac{D_t}{K_t}\right]} \tag{2}$$

where $K_{f,d}$ are the frozen and thawed hydraulic conductivities, and D is the total thickness of the soil layer.

The effective porosity for each soil layer is simply estimated from the fractional components of the soil layer:

$$P_{eff} = 1.0 - F_m - F_o - F_{ice} \tag{3}$$

where $F_{m,o,ice}$ are the fractional components of mineral, organic, and ice.

Results

Evaluation of the TDFT

The TDFT algorithm is tested in the Imnavait Creek Watershed, Alaska (68°30′N, 149°15′W). Two sites, one in a valley bottom and one along a hill slope, are selected for testing. Compared with the hill slope site, the valley bottom site has a thick surface organic layer (0.3 m versus 0.2 m) and has a high soil moisture content. Table 1 shows the soil properties used to test the TDFT. The soil moisture content of each soil layer is the linear interpolation of field measurements to the center of each soil layer.

The surface temperature sensors used in the valley and hill slope sites are located at 2 and 7 cm below the land surface, respectively. Ground temperature measurements, located at 98 and 69 cm below the ground surface, were used for valley bottom and hill slope sites. In our simulations, the ground temperature can only be used in the soil freezing process (soils are not allowed to thaw from depth).

Initial simulations resulted in an underestimation of the maximum thaw depths at each site, probably as a result of underestimation of the amplitude of fluctuation of the ground surface temperature. The timing of the beginning of thaw/freezeback process was also slightly offset from measured field data. To better estimate ground surface temperatures, the near-surface soil temperatures were estimated via extrapolation of amplitude to the surface (Fig. 1). This resulted in an amplification of annual surface temperature fluctuation by 5% at the valley bottom and by 38/27% for temperatures above/below 0°C at the hill slope site. We explain the difference in amplitude by differences in near-surface moisture availability. Compared to the dry hill slope site, the wet valley bottom site has a high thermal diffusivity in the near surface soils. The combination of the high thermal diffusivity as well as the near proximity to the ground surface result in a much smaller surface temperature adjustment at the valley bottom site. Figure 2 displays the simulated freeze-thaw interface compared with measured soil temperatures and soil moisture content at each site.

Table 1: Soil properties of the valley bottom and hill slope sites.

Layer	Soil Depth (cm)	Valley Bottom Site				Hill Slope Site			
		Bulk Density (kg/m^3)	Porosity (%)	Organic / Mineral Fractions	SMC / SMC min	Bulk Density (kg/m^3)	Porosity (%)	Organic / Mineral Fractions	SMC / SMC min
1	10	150	80	0.20 / 0.0	0.45 / 0.05	150	80	0.20 / 0.0	0.40 / 0.06
2	10	260	63	0.37 / 0.0	0.60 / 0.15	560	63	0.37 / 0.0	0.40/ 0.04
3	10	855	55	0.25 / 0.2	0.55 / 0.14	1530	40	0.0 / 0.6	0.38 / 0.10
4	10	1530	48	0.0 / 0.48	0.48 / 0.14	1530	40	0.0 / 0.6	0.40 / 0.12
5	10	1530	40	0.0 / 0.6	0.40 / 0.06	1530	40	0.0 / 0.6	0.40 / 0.12
6	10	1530	40	0.0 / 0.6	0.40 / 0.06	1530	40	0.0 / 0.6	0.40 / 0.12
7	10	1530	40	0.0 / 0.6	0.40 / 0.05	1530	40	0.0 / 0.6	0.40 / 0.12
8	15	1530	40	0.0 / 0.6	0.40 / 0.05	1530	40	0.0 / 0.6	0.40 / 0.12

SMC / SMC min: Volumetric soil moisture content / Volumetric unfrozen water content of frozen soil.

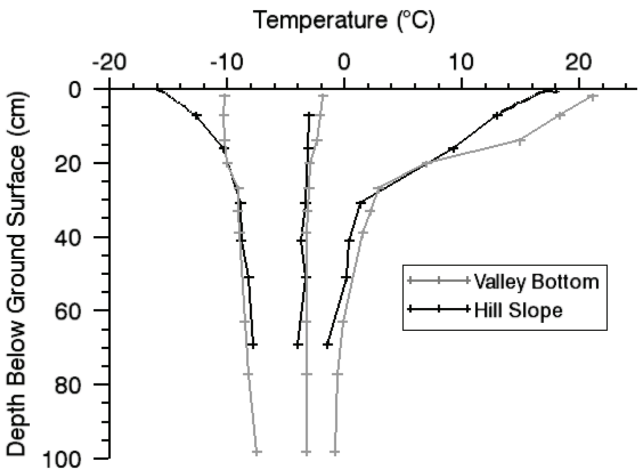

Figure 1. Trumpet curves showing the minimum, maximum, and average temperatures at the valley bottom site (light line) and hill slope site (dark line) for 2003. The "+" indicates location of measurements.

For these simulations and comparisons, daily temperature and soil moisture data measured at 12:00 are used. Results indicate that for most years, the TDFT algorithm is able to simulate the beginning of thaw, the maximum thaw depth, and freezeback period. The latent heat effect during the freezeback period is also represented. Estimated porosities derived from these simulations are shown in Figure 3.

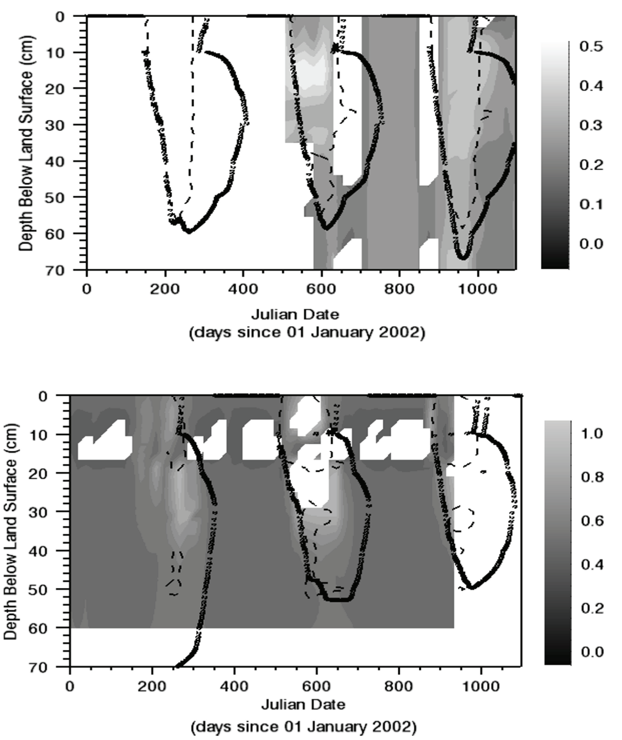

Figure 2. Comparison of the simulated freeze-thaw interface (dark dots) in the valley bottom (top panel) and hill slope (bottom panel) sites with the -0.1°C isotherm (dark dashes) and unfrozen soil moisture content (background). Color bar indicates volumetric soil moisture content. Solid white regions indicate missing data.

Discussion and Conclusions

The seasonal freezing and thawing of the active layer and the associated changes in hydraulic properties are defining features of arctic hydrologic systems. Despite the known importance of the active layer in hydrologic systems, it has been poorly represented in meso-scale hydrologic models. The TDFT algorithm provides a foundation for estimating the freezing and thawing of the soils, hydraulic conductivity, and storage capacity of the soils.

The TDFT, based upon the Stefan equation, assumes that sensible heat effects are negligible. Care must be taken when applying the TDFT in drier areas, where sensible heat fluxes may be an important component of the energy balance. A basic understanding of the physical system being modeled is critical before applying the TDFT.

Using the TDFT to estimate hydraulic properties is an improvement over previously used methods in that 1) changes in the thermal and hydrologic regimes are continuously and adequately captured over time; 2) the ability to simulate the freezeback period allows for longer (year-to-year) simulations, whereas the other methods such as using the

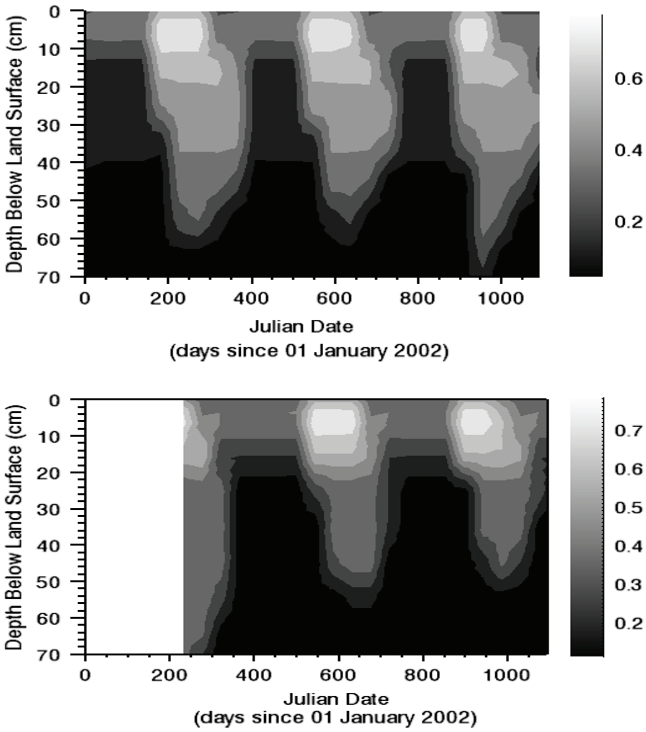

Figure 3. Simulated soil porosity at the valley site (top panel) and hill slope (bottom panel). Solid white regions indicate missing data.

square root of time function is only able to simulate the thawing process; 3) the TDFT is physically based, requiring no prior calibration; and 4) the TDFT is computationally cheap.

The depth and rate of thawing/freezing of the active layer are both spatially and temporally variable. As a result the hydraulic properties of the soils are spatially (x-, y-, and z-directions) and temporally (both short and long term) variable. Accurate freeze-thaw boundary simulations using the TDFT algorithm are highly dependent upon accurate surface temperature data. In order to take full advantage of the TDFT derived variables in (spatially-distributed) meso-scale models, a method needs to be developed to obtain accurate spatial and temporal surface temperatures from within the modeling area.

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