Investigating the Effects of Lateral Water Flow on the Spatial Patterns of Thaw Depth

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Abstract

The effects of lateral water flow on the spatial distribution of the thaw depth in permafrost terrain have rarely been investigated with models. The GEOtop model, which solves the soil energy and water budgets in a coupled way and accounts for phase change, has been used to better understand how soil moisture spatial differences in the unfrozen upper part of the ground affect the thawing soil energy balance in idealized hillslope topography. Results show that, in terrains with thermal conductivity highly variable with soil moisture such as organic soils, wetter areas exhibit deeper thaw than drier areas. Conversely, if thermal conductivity depends less on soil moisture, as in mineral soils, the result is the opposite since the effect of the higher thermal capacity resulting from higher soil moisture prevails.

Keywords: thaw depth; GEOtop; permafrost modeling; spatial variability.

Introduction

In the thawed portion of seasonally frozen or permafrost terrain, lateral drainage flow of soil water often occurs and is a significant hydrological process in both high-latitude (e.g., Quinton & Marsh 1999) and high-altitude environments (e.g., Krainer & Mostler 2002). In the complex heterogeneity of soil properties, the spatial variability of soil moisture can also lead to different behaviors with respect to freeze and thaw processes. For example, Kane et al. (2001) observed that in the Alaskan arctic tundra, thawing is enhanced in water tracks and areas of high soil moisture. In a similar arctic environment, Wright et al. (2009) also observed that wetter areas are normally associated with deeper summer thaw.

Most permafrost models are one-dimensional (e.g., Hinzman et al. 1998, Lawrence & Slater 2005, Marchenko et al. 2008) and applied at large scales. As a result, the impact of soil moisture redistribution due to lateral drainage flow on permafrost evolution has rarely been investigated with models. Endrizzi et al. (2010) simulated with the GEOtop model (Rigon et al. 2006, Endrizzi 2009, Dall’Amico et al. 2011) the spatial variability of the thaw depth in a peat-covered small catchment in the Canadian arctic tundra, and found that the end-of-summer thaw depth is significantly greater where the water table is shallower.

The purpose of this work is to qualitatively assess how the spatial variability of soil moisture due to lateral water drainage redistribution in the unfrozen part of the soil affects the spatial variability of the thaw depth. This study will be theoretical and will be carried out in an idealized topography. There is no pretense of finding general results. The purpose is only to have an idea of the spatial variability of the end-of-summer thaw depth in different soil types and to understand controlling processes.

Methodology

This study will be conducted with a model using a very simple planar topography. As various soil types are thought to respond differently to hydraulic gradients and thermal forcing, three characteristic soil types are considered: 1) loam, representing a soil type with relatively low hydraulic conductivity and high thermal conductivity; 2) sand, a soil type with high hydraulic conductivity and thermal conductivity; and 3) peat, characterized by high hydraulic conductivity and low thermal conductivity.

We use the GEOtop distributed model (Rigon et al. 2006, Endrizzi 2009, Dall’Amico et al. 2011). This model is particularly appropriate for the present study because it couples the equations describing the three-dimensional water flow in the soil (Richards equation in variably saturated conditions for subsurface flow and De-Saint-Venant equation for the surface flow) with a one-dimensional form (normal to the surface) of heat equation accounting for soil freezing and thawing processes. Mass and energy exchanges with the atmosphere constitute the boundary conditions and include a dynamic multi-layer snow cover. Input data to the model are given by the digital elevation model and meteorological data consisting of distributed precipitation, air temperature, relative humidity, wind speed, and incoming longwave and shortwave radiation.

The model has been run for a 200-m-long planar hillslope, with a 20-degree slope angle and south aspect. The hillslope is described with a virtual DEM composed of 40 lines and 1 single column, where the column represents the slope direction and the rows the direction normal to the slope direction. Gradients are different from zero only in the slope direction. The model is forced with the meteorological data measured at the Corvatsch Meteoswiss station, located at 3305 m a.s.l. in the southeastern Swiss Alps (46.42°N, 9.82°E), at the summit of Piz Corvatsch mountain.

Modeling Assumptions

The modeled domain is given by the first 10 m of soil, discretized with 25 layers of variable thicknesses that increase with depth from the surface, since temperature gradients normally decline with depth. The lower boundary of the deepest layer is considered impermeable to water flux, which is reasonable since at this depth the soil generally remains
The heat fluxes at the upper and lower boundaries of the modeled 10-m-deep soil domain are given, respectively, by the ground heat flux and geothermal heat flux. If the ground surface is snow free, the former is given by the surface heat flux (algebraic sum of net radiation and turbulent fluxes), which, in turn, depends on the temperature of the surface, an unknown in the system. If a snow cover is present, the ground heat flux is given by the soil-snow conductive heat flux, which constitutes the exchange coupling term between the heat equations for soil and snow. The geothermal heat flux is the heat flux that maintains an approximately constant temperature geothermal gradient below the zero-amplitude annual soil temperature depth, normally at 10–15 m in the Alps (Gruber et al. 2004), which is comparable with the depth of the modeled soil. A common value of the geothermal heat flux is 0.08 W/m² (Medici & Rybach 1995).

**Model Initialization**

The model results should ideally be completely independent of the initial conditions, which always have a certain degree of arbitrariness. In order to accommodate this, we proceeded in the following way. The initial conditions of soil water content and temperature over the domain have simply been guessed and, for simplicity, considered uniform over the hillslope. They are given by the absence of snow cover, soil temperature equal to -5°C at all depths, and soil saturated below a depth of 30 cm. We selected a particular one-year set of meteorological data (from July 2002 to July 2003), and we repeated it several times. Therefore, if the model is forced with periodical meteorological data, it is expected that the state variables also eventually become periodical. When it is determined that the temporal evolution of both soil temperature and moisture has become a periodical function within a reasonable approximation, the results can be considered acceptably independent from the initial condition.

Figure 1 shows the time evolution of the soil temperature at the 10-m depth. While at the surface the effect of the initial condition is immediately lost, at the lowest depth, where there is greater thermal inertia, the effect is felt for approximately 7–8 years, after which the temperature evolution reaches a periodic state. Therefore, if we examine the results after 10 years, we can state that the results are reasonably independent from the initial condition in the whole domain. This is true for the three soils types studied here.

### Spatial Variability of the End-of-Summer Thaw Depth

The thaw depth (Fig. 2 upper part) has been defined here as the lowest interface separating a frozen (below) and unfrozen (above) soil layer. This variable reaches the maximum value near the end of September for each soil type. Maximum values are around 1.35 m for sand, 0.9 m for loam, and 0.3 m for peat. In the sandy soil, the thaw depth is the largest since the porosity is relatively low and, consequently, there is less ice to melt. Conversely, the peat yields the shallowest thaw as a result of its high porosity (and therefore there is more ice to melt) and very low thermal conductivity.

Another variable related to the thaw depth is the depth of the saturation front (Fig. 2 lower part), which is defined as the lowest interface between saturated (below) and unsaturated (above) soil layers, no matter if the soil is saturated with frozen or unfrozen water. In this application, the depth of the saturation front is normally just a little deeper than the correspondent maximum thaw depth, although it rises at times during summer. This demonstrates that, due to the relatively steep slope, the thawed part of the hillslope is in general well drained.
After initializing the model as previously described, the longitudinal profiles of the thaw depth and saturation front at the end of the summer are well correlated in the sandy and loamy soils (Fig. 3). Drainage flow determines the longitudinal profile of the saturation front, which is relatively deep in the upper part of the hillslope since water balance is here negative and downward drainage is not compensated by drainage from upward regions. As we move downslope, the saturation front rises and approaches a state of almost zero longitudinal derivatives. Then it sharply sinks in close proximity to the lateral lower boundary of the hillslope as a consequence of the boundary condition. Therefore, the longitudinal profile of the hillslope can be roughly partitioned in three regions: a relatively dry portion at the top (hereafter referred to as top portion), a relatively wet region in the center (mid portion) where infinite slope conditions are approached, and a small dry region at the bottom (bottom portion) which directly feels the effect of the boundary condition. The depth of the saturation front at the end of the summer ranges from 0.73 m in the mid portion to 0.86 m in the top portion for loam, and from 1.1 m to 1.4 m for sand. The larger values for sand are explained by the higher hydraulic conductivity and enhanced drainage. The end-of-summer thaw depth is always larger where the saturation front is deeper. This means that relatively dry (wet) regions always experience relatively deep (shallow) thaw. The three portions in which the hillslope can be subdivided according to the profile of the saturation depth can be clearly recognized also in the profile of the end-of-summer thaw depth, which varies from 0.87 m in the mid portion to 0.91 m in the top portion in loam, and from 1.30 m
Patterns of Soil Temperature and Soil Moisture

In loam and sand, the longitudinal profile of the annual averaged soil temperature (Fig. 5) at all depths is well correlated with the longitudinal profile of the thaw depth at the end of the summer, which means that the regions with deeper thaw are also warmer. The annual averaged soil temperature varies by about 0.10–0.25°C, with higher variability at the surface. Conversely, the opposite pattern is observed in the peat; regions with deeper thaw are colder.

The longitudinal profiles of the annual averaged total soil moisture (Fig. 6) are also well correlated with the longitudinal profile of the depth of the saturation front for sandy and loamy soils. In peat, the longitudinal profile of annual averaged soil moisture is not so clearly correlated with the depth of the saturation front, but shows instead a clear correlation with the thaw depth in the sense that where soil moisture is higher, thaw is deeper. Comparing the spatial variability of the annual averaged soil moisture and temperature at the surface, we observe that higher soil moisture is always correlated with colder temperatures.
Discussion

The study has shown that higher soil moisture leads to shallower thaw in certain types of soils, and to deeper thaw in others. There are two important processes that affect the spatial variability of the thaw depth in the presence of significant soil moisture spatial differences:

1) Where soil moisture is higher, the soil thermal conductivity is also higher, and this is particularly true for peat. This contributes to the positive heat flux exchanged with the atmosphere during the summer that penetrates more deeply into the soil.

2) Soil thermal capacity is also higher where soil moisture is higher. Thermal capacity has a “real” component, given by the fact that wetter soils require more heat than drier soils to increase their temperature, and also an “apparent” component, since wetter soils normally have larger ice content to thaw. In addition, wetter soils lose more energy by evaporation than drier soils. This second process obviously causes deeper thaw in drier soils.

In peat, the first process prevails, even if the thermal capacity of the peat soil also varies significantly with water content. Conversely, in sand and loam the second process prevails since the thermal conductivity does not depend so strongly on soil moisture as does peat.

The quantitative results presented here are not to be generalized to other more complex cases. They are relative only to the simulated case. Other numerical experiments in more complex topographies must be carried out in order to extend these results.

Conclusion

In this study, the spatial variability of the thaw depth in a simple 20-degree sloped ideal planar topography has been investigated using the GEOtop model. The purpose is to have a qualitative idea of how soil moisture redistribution due to lateral drainage can affect the end-of-summer thaw depth. This effect has been investigated in three different soil types: loam, sand, and peat.

The spatial variability of the thaw depth has been found relatively limited, reaching approximately 20 cm when the thaw depth is around 1.5 m. However, this cannot be generalized to other cases. It is remarkable to note that in
loam and sand, deeper thaw has been observed where soil moisture is lower, but in peat, deeper thaw occurs where soil moisture is higher. Wetter soils actually have higher thermal conductivity but also higher thermal capacity. The former allows deeper penetration of the summer thaw, while the latter reduces the thaw rate as more energy is required to increase soil temperature. In peat, as a result of the very high difference between thermal conductivity in drier and wetter portions, the former process prevails. On the other hand, in sand and loam, where the thermal conductivity is only moderately dependent on soil moisture, the effect of thermal capacity is prevalent.

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