SOIL LATENT HEAT AS A FILTER OF THE CLIMATE SIGNAL IN PERMAFROST

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Abstract

This paper explores the effects of latent heat on the thermal response of the upper 20 metres of the ground to warming at the surface. As a problem in geothermal analysis, the response of permafrost to a change in the surface temperature is a function of both the initial condition (ground temperature profile) and thermal properties. The response to warming of one soil under different surface thermal regimes is examined by numerical simulation in order to indicate the range of thermal responses which can be expected of earth materials due simply to differences in their initial thermal condition. The change at the surface is attenuated most in permafrost near the cold end of temperature range over which phase change is significant. Surface temperature change also alters the annual range at each depth, as seasonal extremes adjust to the new mean in different ways, depending on whether the surface temperature is shifting toward or away from the range in which latent heat has a significant effect. These results demonstrate that short term, near surface ground temperature trends cannot be an absolute measure of the magnitude and rate of change to the surface thermal regime when significant soil latent heat is present. Those sites most vulnerable to climate warming (warm, ice-rich permafrost) will provide the weakest ground temperature signal.

Résumé

Cet article explore les effets de la chaleur latente sur la réponse thermique des premiers 20 mètres de sol au réchauffement en surface. En tant que problème en analyse géothermique, la réponse du pergélisol à une variation de la température de la surface est une fonction autant de la condition initiale (profil de température du sol) que des propriétés thermiques. La réponse au réchauffement d'un sol selon différents régimes thermiques de surface est examinée par simulation numérique dans le but d'établir la gamme des réponses thermiques possibles des constituants du sol dues simplement aux différences de condition thermique initiale. Le changement en surface est surtout atténué dans le pergélisol au voisinage de la limite inférieure de la plage de température dans laquelle le changement de phase est important. La variation de température en surface modifie aussi la plage annuelle à chaque profondeur, à mesure que les extrêmes saisonniers s'ajustent à la nouvelle moyenne de différentes façons selon que la température de surface s'approche ou s'éloigne de la plage dans laquelle la chaleur latente a un effet important. Ces résultats montrent qu'à court terme, la tendance de la température du sol près de la surface ne peut être une mesure absolue de l'amplitude et du taux de variation du régime thermique de surface lorsque la chaleur latente est importante. La réponse aux sollicitations du réchauffement climatique sera plus faible dans les sols normalement les plus vulnérables, c'est-à-dire les sols gelés peu froids et riches en glace

Introduction

The usefulness of permafrost temperature monitoring as a means of detecting and quantifying climate change depends on our understanding of the relationships between (1) air temperature and ground surface temperature, and (2) surface temperature and the ground thermal regime. This paper explores aspects of the second of these problems, in particular the effects of latent heat on the thermal response of the upper 20 metres of the ground.

As a problem in geothermal analysis, the response of the ground to a change in the surface temperature regime is a function of both the initial condition (ground temperature profile) and thermal properties. In deep permafrost materials where latent heat effects are not dominant, it is possible to analyse ground temperature profiles and determine the timing and magnitude of past surface temperature shifts (e.g. Lachenbruch and Marshall 1986). More recent events influence temperatures at shallower depths.

In permafrost near the surface, however, the temperature dependent unfrozen water content causes latent heat to be absorbed or released over a range of temperatures, and also results in temperature dependent thermal conductivity and heat capacity. In consequence of this, any particular ground temperature profile implies a particular depth profile of thermal properties, which will determine the nature of the ground thermal response. The rate of change of properties is greatest just below the melting point of free water (Farouki 1981, Riseborough et al. 1983); the change in latent heat is the dominant temperature dependent property (Smith and Riseborough 1985).

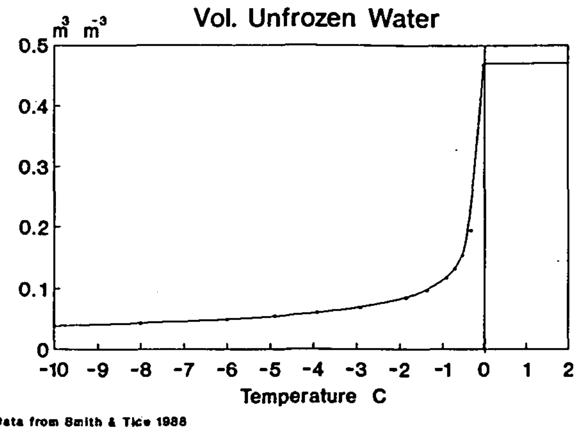


Figure 1. Volumetric unfrozen water content curve for Goodrich clay, used in simulations. Data from Smith and Tice 1988.

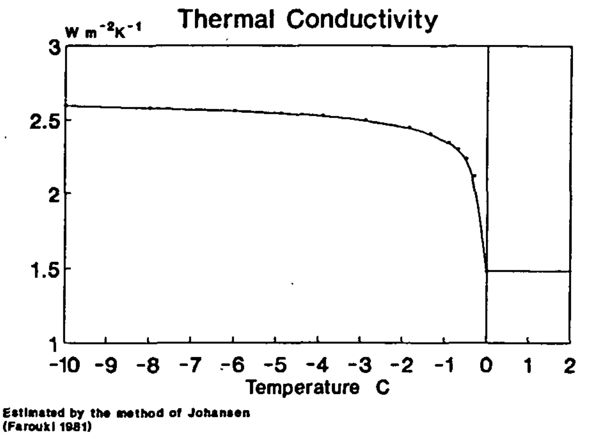


Figure 2. Thermal conductivity for Goodrich clay, used in simulations. Estimated using the method of Johansen (Farouki 1981).

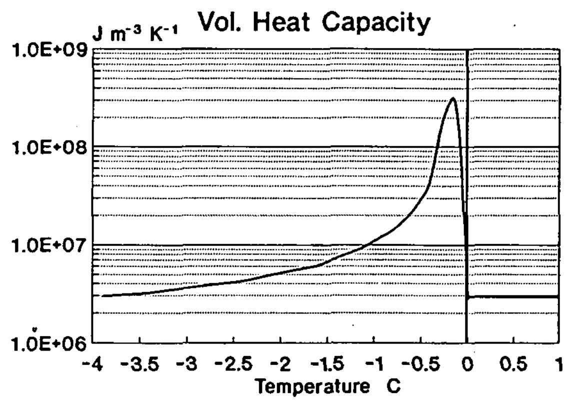


Figure 3. Volumetric heat capacity for Goodrich clay used in simulations.

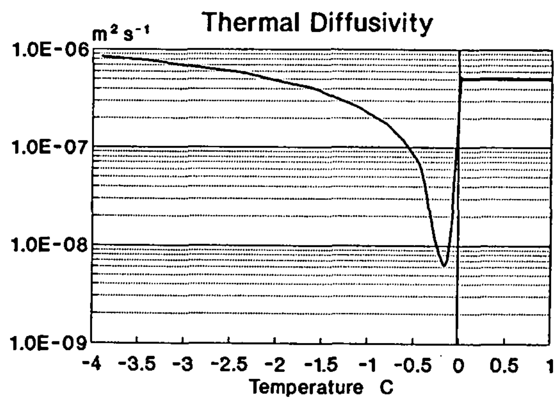


Figure 4. Thermal diffusivity for Goodrich clay, used in simulations.

These effects may have significant implications for the usefulness of permafrost ground temperatures as a means of monitoring changes to the surface thermal regime. The aim of the geothermal simulations presented here is to indicate the range of short term thermal responses which can be expected of earth materials as a result of differences in initial thermal conditions (temperature profile).

METHODOLOGY

The response of ground temperatures to warming at the surface was examined by numerical simulation for a catena of ground thermal regimes, having mean annual surface temperatures (M.A.S.T.) ranging from -8° to +1 °C (M.A.S.T. of -8, -3, -2, -1, -0.5, -0.25, and +1). The cases differed only in the mean annual surface temperature: the temperature dependent soil thermal properties specified for each case were identical (Figures 1 and 2): volumetric heat capacity and latent heat were estimated for Goodrich Clay from the temperature-unfrozen water relationship presented in Smith and Tice (1988); thermal conductivity was estimated using Johansen's method (Farouki 1981).

The geothermal model uses supplied thermal conductivity and unfrozen water content data (and additional physical data) to estimate heat capacity and thermal diffusivity (Figures 3 and 4), with latent heat being treated as an apparent heat capacity. At any temperature, thermal conductivity and unfrozen water content are estimated by cubic spline interpolation, with the contribution of latent heat to the apparent heat capacity estimated from the slope of the spline curve. To speed up operation of the model, a table of thermal properties is first established (in increments of 0.1 °C), so that calculations are not required at each time step. Comparative tests between tabular values and values calculated at each time step show excellent agreement between results.

A sinusoidal temperature wave with a period of one year and an amplitude of 10 °C was imposed at the ground surface (Figure 5). Ground temperatures were initially established with surface temperature set at the mean annual surface temperatures specified above, and with a linear geothermal gradient of 0.02 °C per metre prior to the imposition of the surface temperature wave: simulations were first run for 20 to 100 years to establish initial "equilibrium" ground

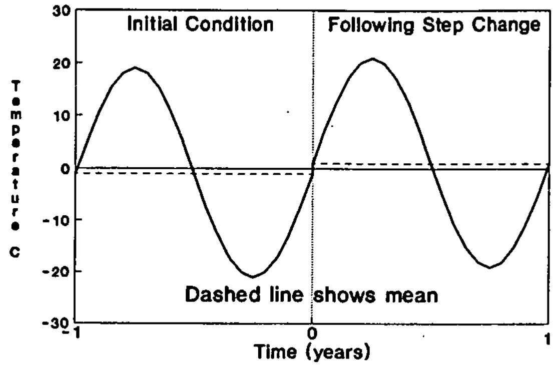


Figure 5. Typical surface boundary condition used in geothermal simulations.

temperature profiles. The profiles exhibit "thermal offset" in the mean annual temperature profile caused by the temperature dependence of the thermal conductivity (Goodrich 1978, Burn and Smith 1988). This effect makes it difficult to establish whether a true periodic steady state ground temperature profile has been established. In the simulations presented here, profiles differed by less than 0.001 °C between successive years at the end of equilibration.

Following equilibration, the mean of the surface temperature wave was increased by 2°C, and the new surface condition sustained for a further 10 years. Profiles are still far from a new equilibrium at the end of the simulations.

The simulations were performed using a one-dimensional implicit finite difference model, as described in Smith and Riseborough (1985). In these simulations, the time step was 6 hours, the node spacing was 250 mm, and the lowest node was at a depth of 125 metres, with temperature held constant at the lower boundary. This depth and boundary condition are adequate for the current simulations: results using the analytical solution for a step change of surface temperature without phase change (Birch, 1948) show that the temperature change at 100 m depth ten years after a step change of 2 degrees at the surface is about 0.0001 °C with a diffusivity of 10^{-6} m^2s^{-1} (slightly above the highest diffusivity used in the simulations).

Results and discussion

Annual ranges

The annual temperature envelopes under both the initial equilibrium conditions and after 10 years of warming are shown in Figures 6a to 6g. Under equilibrium conditions, the depth of zero annual amplitude (arbitrarily defined here as the depth at which the amplitude is 1 percent of the surface amplitude) is greatest under the coldest surface conditions (at about 13 m), an becomes progressively shallower in the warmer permafrost simulations: in the two warmest permafrost simulations, the depth of zero annual amplitude is 3 m or less. For the case with no permafrost present, the depth of zero annual annual amplitude is about 9 m.

After ten years under the new surface temperature condition, annual ranges for each case change as the profiles shift to warmer temperatures. At an initial M.A.S.T. of -8, the depth of zero annual amplitude changes little under the new surface regime. At (initial) M.A.S.T. of -3 and -2, the depth of zero annual amplitude decreases under the new regime, as the profiles moves toward the temperature range where phase change dominates. At still warmer temperatures the zero annual amplitude depth becomes progressively deeper again, as the profile moves away from the range of significant phase change.

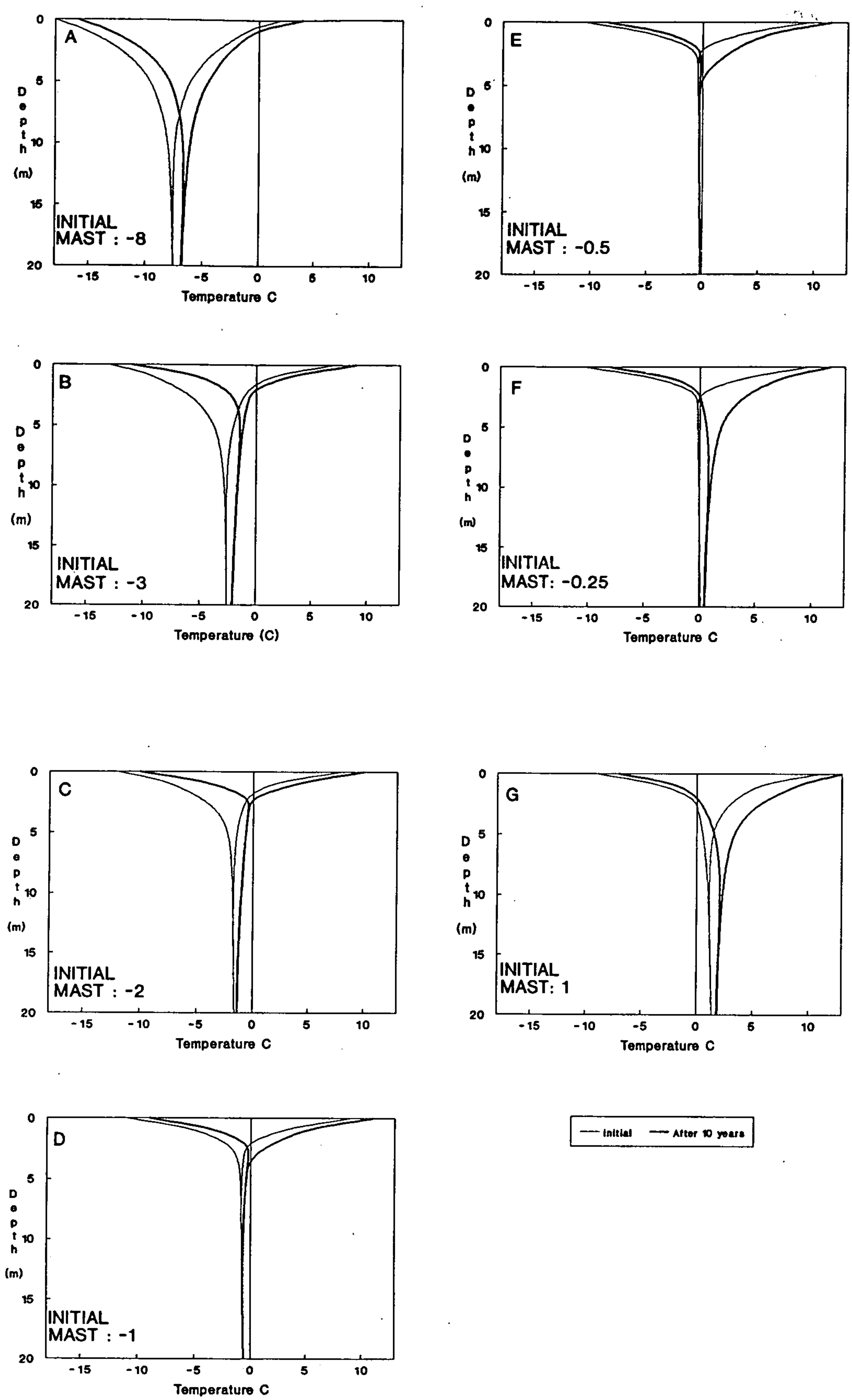
The maximum and minimum of the annual ranges shift in different ways under the new regimes. With the initial M.A.S.T. of -8, the maximum and minimum shift by roughly the same amounts; with the initial M.A.S.T. of -3 and -2, minimum temperatures increase to a much greater extent than maximum temperatures do. At an initial M.A.S.T of -1 and above, maximum temperatures increase more than minimum temperatures do.

This demonstrates that a change in the mean surface temperature results in more than just a change in mean annual ground temperatures. The thermal regime is changed, altering the annual range: seasonal extremes adjust to the new mean in different ways, depending on whether the surface temperature is shifting toward or away from the range in which latent heat has a significant effect. Increasing temperatures through the "warm frozen" range requires the exchange of far more energy per degree than the warming of unfrozen material (or of very cold permafrost). As a result, any seasonal bias in changes to the annual range at the ground surface will likely be obliterated in the sub-surface temperature record.

Trends in mean annual temperature

Mean annual temperatures were calculated based on results generated at 5 day intervals. These were used to generate an approximately continuous mean annual temperature curve for selected depths for each case, by calculating a running mean of 73 values (essentially a mean annual temperature updated every 5 days).

Figure 7 shows the trends of mean annual temperature at the 3 metre depth for all cases, while Figure 8a compares these changes to one another, by subtracting the initial temperature from each of the trends. Figures 8b and 8c show the rise in mean annual temperature at the 5 m and 10 m depths respectively. For all cases, the surface change lags significantly with depth. For the coldest case, most of the temperature rise at the surface has penetrated to 3 metres within two years following the change, and more than half of the surface temperature increase has reached 10 m depth by the end of the simulation. In contrast, the case initially at -0.5 °C M.A.S.T. has warmed by less than 1 °C at 3m after 10 years, and at 10 m depth is effectively unchanged even



Figures 6a - 6g. Annual temperature envelopes for simulation cases: Initial annual envelop and envelop 10 years following change in M.A.S.T.

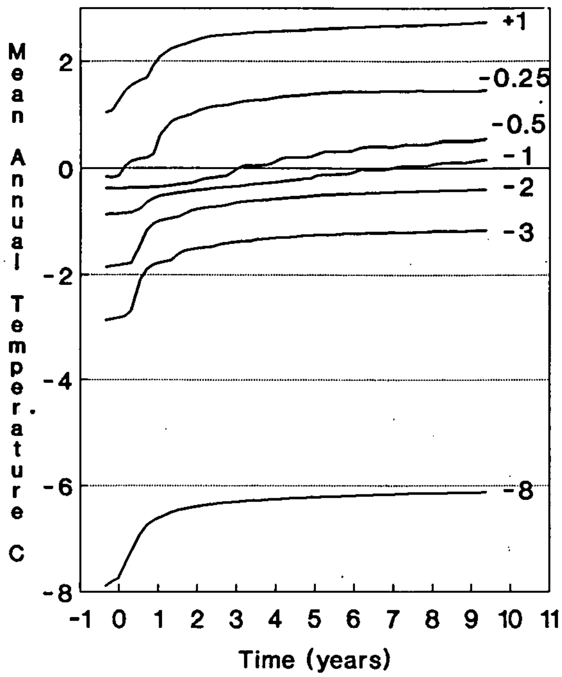
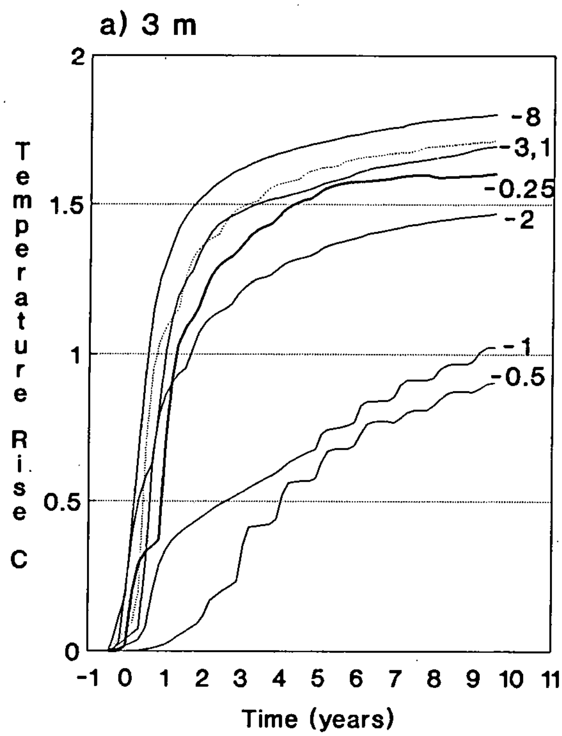
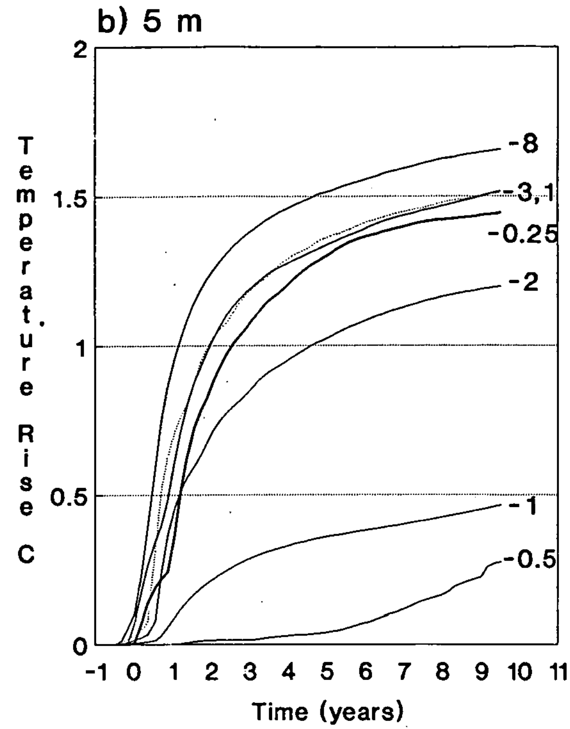
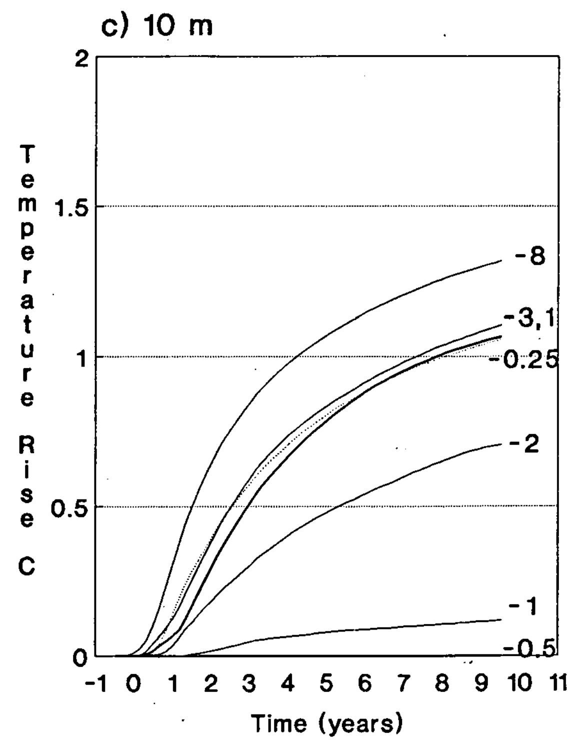


Figure 7. Changes in mean annual temperature at 3 metre depth, following increase in M.A.S.T (at year = 0). Labels refer to initial M.A.S.T.







Figures 8a - 8c. Rise in mean annual temperature from initial value, for 3, 5, and 10 m depths. Labels refer to initial M.A.S.T.

after this interval. The other cases have responses which fall between these extremes: at 3 m depth the range of responses is uniformly distributed between the extremes, while at 10 m responses tend to group between those that experience a small latent heat effect and those where the effect is dominant. The difference in the rate of propagation of the surface disturbance into the soil for the coldest case and the non-permafrost case can be attributed to the higher thermal diffusivity of the cold permafrost (i.e. of frozen relative to unfrozen soil).

The attenuation of the annual surface range (i.e. the depth of zero annual amplitude) is not a good predictor of the rate of change in response to surface change, (compare Figures 6d and f, and Figure 8a-c for the -0.25 and -1 M.A.S.T. cases).

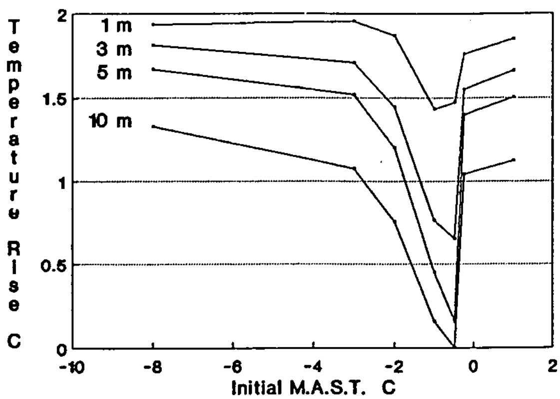


Figure 9. Rise in mean annual temperature after 10 years, as a function of depth and initial M.A.S.T. Labels refer to depth.

The curves for shallower depths (eg. Figure 8a) at most sites exhibit annual jumps in the mean, rather than a smooth increase over the year. This effect is due to the changes in temperature range as the soil warms: where the range changes more in summer than winter (warmer sites), the mean jumps as the summer temperatures are added to the running mean. The steps in the mean for cooler sites are therefore out of phase with those for the warmer sites.

The relative responses of the three warmest permafrost cases demonstrates the role of soil latent heat in determining the thermal response of the ground. For the warmest permafrost case, temperatures change almost as quickly as for the non-permafrost case: this can be attributed to the relatively small ice content remaining in the soil at this temperature, and also to the fact that permafrost thickness was initially only 12.5 m for this case. In contrast, the response is considerably slower for the two cooler warm-permafrost cases, since most of the ground ice must first thaw in order to raise the ground temperature through 0 °C. At 3 m depth the response of the -0.5 ° M.A.S.T. case is consistently less than for -1 ° M.A.S.T. case, while at 5 m the relationship appears likely to reverse within the next decade: the rate of change can increase quickly when the annual maximum temperature gets beyond the phase change temperature range (i.e. when active layer reaches this depth).

The effect of depth and initial condition on the thermal response at a given point in time (in this case, after 10 years of warming) can be examined by plotting the rise in mean annual temperature as a function of the initial temperature (Figure 9). The lowest response is not for the case at the warmest freezing temperature, but rather for the case which undergoes greatest change of phase: the case for which this is true varies with depth (compare depths 1 and 3 m, Figure 9).

The behavior illustrated in Figure 9 roughly follows the temperature dependent thermal diffusivity function (Figure 4), emphasizing the fundamental control of this property in thermal problems. The shapes of the curves in Figure 9 are slightly different for each depth because of the combined effects of the annual temperature range at each depth and the time lag with the surface temperature change. For the depths examined here, the greatest spread in responses (highest versus lowest temperature rise) is at the 5 m depth.

Conclusions

These simulations represent a simplification of what might be expected in nature: the influence of snow cover on the surface temperature regime, the distribution of ice in the soil profile, and the presence of a surface organic layer will all alter the thermal property profile of the ground. The nature (step, ramp, or non-linear function) and magnitude of the change in mean ground surface temperature, as well as any change in the annual range will also influence the response of the ground thermal regime. The "temperature of lowest thermal response" will depend on the magnitude and type of surface change, which suggests that inter-annual variability may also be important. Notwithstanding these limitations, behavior of the type presented in Figure 9 has been documented in the field (Burgess and Riseborough, this volume).

The temperature range over which latent heat is released or absorbed for a particular soil establishes the temperature range over which it behaves as "warm" permafrost: coarse grained soils containing no fine grained material will thaw over an extremely narrow temperature range, perhaps less than one tenth of one degree, while in fine grained soils latent heat effects can be observed over several degrees. As well, the amount of ice present in the frozen soil determines the penetration rate of a surface warming trend into warm permafrost, with ice rich sediments taking longer to thaw. Differences in thermal properties (due to differences in soil mineralogy, density, etc.) will also alter the thermal response of the soil.

The results presented here suggest that differences in ground temperature trends cannot be an absolute measure of differences in the magnitude changes to the surface thermal regime. Those sites most vulnerable to climate warming (warm, ice-rich permafrost) will provide the weakest ground temperature signal.

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