SURFACE ENERGY BALANCE AND GROUND HEAT FLUX IN ORGANIC PERMAFROST TERRAIN UNDER VARIABLE MOISTURE CONDITIONS

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Abstract

Detailed measurements of the surface energy balance, soil thermal regime, and soil moisture were collected at a site near Churchill, Manitoba in July and August, 1987. The site consisted of a shallow layer of peat overlying mineral soil, with variations in surface relief of 10-20 cm. Sub-surface measurements were taken at dry, intermediate, and wet locations a few metres apart, to a depth of 15 cm. Surface moisture content and surface temperature showed wide variations over these horizontal distances.

The data were analyzed to examine the importance of horizontal variations in surface moisture. Although the reduction of thermal conductivity in the drying peat is important in reducing the ground heat flux, the dry layer is confined to the first few centimetres of the soil profile. The non-homogeneous surface moisture causes variations in surface temperature which make it difficult to model the surface energy balance as a one-dimensional process using a physically-based approach. It is questionable whether the sensitivity of a one-dimensional model will adequately represent the true sensitivity of permafrost to extreme changes in soil moisture in this environment.

Résumé

Des mesures détaillés du bilan énergétique, du régime thermique du sol, et de l'humidité du sol ont été effectuées sur un site près de Churchill, Manitoba, en Juillet et Août, 1987. Le site est caractérisé par une couche de tourbe mince sur un substrat minéral, avec des variations de relief de 10-20 cm. Des mesures de température ont été effectuées aux lieux secs, intermédiaires, et humides, jusqu'à une profondeur de 15 cm. L'humidité et la température de la surface ont varié beaucoup à cette échelle.

Les données ont été analysées pour examiner l'importance des variations horizontales de l'humidité du sol. La réduction de la conductivité thermique de la tourbe sèche est importante pour limiter le flux thermique du sol, mais la couche sèche se limite aux quelques premiers centimètres du sol. L'humidité non-homogène de la surface provoque des variations de température de la surface qui rendent difficile la tâche de modèliser le bilan énergétique en utilisant une approche physique à une dimension. On doute qu'un modèle à une dimension puisse reproduire exactement la sensibilité du pergélisol soumis aux changements extrêmes de l'humidité du sol.

Introduction

Studies dealing with variations in thermal conditions in permafrost frequently use the concept of the surface energy balance. Simply expressed, the surface separating the ground and atmosphere represents a boundary on which all energy fluxes must balance, such that:

$$Q^* - QH - QE - Qg = 0 \qquad (1)$$

where Q* is net radiation, QH and QE are sensible and latent heat fluxes, and Qg is the soil heat flux. Any factor which influences one or more of the components of the energy balance will also lead to changes in the others, and the effect on Qg and permafrost conditions can be evaluated. Examples of such factors are changes in surface albedo affecting Q*, surface roughness affecting QH and QE, surface moisture affecting QE, and soil thermal properties affecting Qg.

One factor which has often been considered important in permafrost studies is the presence or absence of peat (Brown and Péwé, 1973; Outcalt and Nelson, 1985). The thermal properties of peat tend to favour cooler ground temperatures: the low conductivity of dry peat in the summer inhibits warming, while the high conductivity of frozen, saturated peat in winter enhances cooling.

The surface energy balance concept has been applied to numerical models of ground temperatures by authors such as Outcalt *et al.* (1975), and Smith (1975, 1977). Each component is expressed mathematically, with surface temperature as the independent variable and any other factors (e.g. albedo, surface moisture, thermal properties) specified as parameters or observed values. In equation 1, at any point in time, the surface temperature (T_s) is the only unknown, and can be solved for using a root-finding procedure. This "equilibrium surface temperature" becomes the boundary condition for a numerical solution of Fourier's Law of heat conduction in the soil. Subsequent calculations through time provide a complete history of the soil thermal regime.

Smith and Riseborough (1983) use a model of this type to examine the sensitivity of permafrost to climatic change. They varied parameters in the model dealing with surface thermal properties, roughness, albedo, wetness, slope aspect, and snow cover. The most important parameter was surface wetness, which showed differences in mean annual surface temperature of 10 °C between dry and saturated conditions.

In this paper, field data will be combined with the results of a numerical surface energy balance model to examine the effect of micro-scale variations in surface moisture on the ground thermal regime in organic terrain. The purpose is to evaluate the validity of using one-dimensional energy balance models to predict thermal changes in this type of environment.

Field Observations

The field site was located 20 km east of Churchill, Manitoba, 2 km inland from the Hudson Bay coast. The site is in a low-lying, poorly-drained area, with a sparse sedge cover 5-20 cm in height. The soil consisted of 10-15 cm of peat, overlying a cobble layer and a mineral soil. The area has numerous low hummocks, with local differences in relief of about 10-20 cm. The nature of the terrain results in a wide variation in surface moisture over very short distances.

Meteorological data were collected to allow determination of the surface energy balance. Net radiation, and profiles of air temperature, humidity, and wind speed were measured at 10 second intervals. Hourly averages were used to evaluate QH and QE using the combined Bowen Ratio/aerodynamic technique outlined in Halliwell and Rouse (1989).

Detailed subsurface data consisted of soil temperature profiles, soil heat flux, thermal conductivity and water contents from the surface to a depth of 15 cm. Three locations were instrumented at the site: a dry location at the top of a hummock, a wet location at the bottom of a depression, and an intermediate (mesic) location situated in a low area between the other two. The locations were separated horizontally by distances of 1-2 m, with a total elevation difference of about 20 cm. Although the elevation difference is small, the contrast in surface moisture regimes was large. The wet location was covered with 1-2 cm of water when instruments were installed in late June. The dry location was almost completely dry at the surface at this time.

Each location was instrumented with six soil temperature thermocouples at depths of 1 to 13 cm, a surface temperature thermocouple consisting of four fine-wire thermocouple junctions wired in parallel, a soil heat flux transducer at 1 cm, and a combined Time Domain Reflectometry/thermal conductivity (TDR/TC) probe at 5 cm. The TDR/TC probes were of the design given by Baker and Goodrich (1984). The wet and dry locations also had a second soil heat flux transducer at 10 cm, and a second TDR/TC probe at 10 cm.

To avoid interference between the various types of sensors, the instruments were arranged in three vertical planes: soil temperatures in one plane, soil heat flux transducers in a second, and TDR/TC probes in a third. The three planes were about 25 cm apart. The three sets of instruments were installed by cutting a section of peat out of the ground and inserting the sensors in the side of the exposed pit.

Additional water content readings were taken using vertically-installed TDR probes of lengths 2.5, 5.0, and 10.0 cm. Rather than being installed permanently, these probes were inserted each time a measurement was done. For each length of probe, readings were taken at five or six different spots at each location. This provided an indication of the variability in water content at each location. Although the use of very short probes (2.5 and 5.0 cm) with the TDR technique leads to reduced accuracy, the results provide a qualitative measure of variations in surface moisture.

Soil temperatures and soil heat flux were recorded along with the meteorological data. TDR and thermal conductivity readings were taken once a week on average. TDR readings were converted to volumetric water content using a calibration determined for two similar peat soils from the Churchill area. The measurement period extended from the beginning of July until mid-August, 1987.

Analysis

In this paper, analysis of the data focuses on horizontal and vertical variations in soil temperature, water content, thermal conductivity, and soil heat flux. The horizontal TDR/TC probes at 5 cm indicated that both the dry and wet locations remained very moist, with water contents ranging from saturated (85-90 % volumetric) at the start of the period to about 65 % by mid August. At 10 cm depth, both locations remained saturated throughout the measurement period. Variations in thermal conductivity followed this trend, with measured values ranging from 0.5-0.6 $Wm^{-1}°C^{-1}$ in the saturated peat, to a low of 0.2 $Wm^{-1}°C^{-1}$ at the lowest water content.

The only significant variations in moisture content occurred in the top few centimetres of the peat. At the dry location, the 0-2.5 cm TDR readings were 20-30 % on average. At the mesic location, the 0-2.5 cm readings averaged from 50 % to 90 %. The wet location averaged from 40 % to 100 %. Multiple readings at different points around each location showed a range of ±15 % at any point in time. The wet location remained saturated longer than the mesic location, but once it began to dry out the surface became drier than at the mesic location, due to differences in the moisture retention properties.

Differences in surface moisture led to differences in evaporation, soil heat flux, and surface temperature between

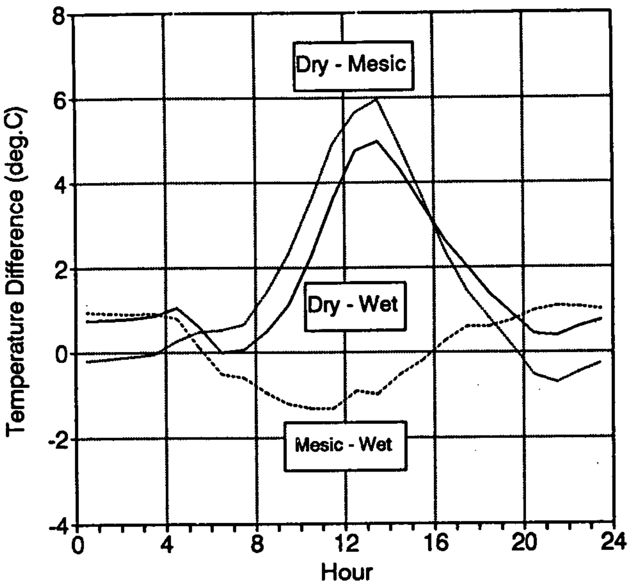

Figure 1. Mean differences in surface temperature between the dry, mesic, and wet locations, versus time of day. Averaged over the period from July 1 to August 15, 1987.

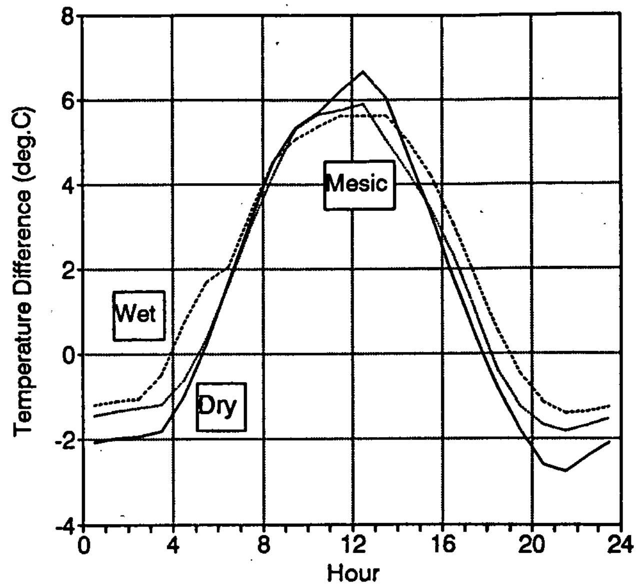

Figure 2. Mean vertical temperature differences between the surface and 2 cm depth at the dry, mesic, and wet locations. Averaged over the period from July 1 to August 15, 1987.

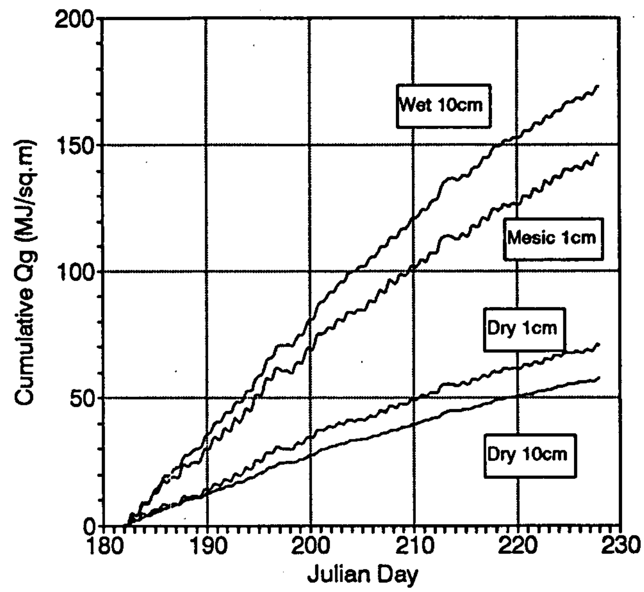

Figure 3. Cumulative heat flux values. July 1, 1987 (Julian Day 182) to August 15, 1987 (Julian Day 227).

the three locations. In addition, the drying of the surface layers led to large vertical gradients in both temperature and moisture content. Fig. 1 shows the mean horizontal differences in surface temperature as a function of time of day, comparing the locations in pairs. The dry location is warmest during the day, averaging 5-6 °C warmer than the mesic and wet locations. Differences on individual days can reach 8-12 °C. At night, the dry location is cooler, but the differences average less than 1 °C. The differences between the mesic and wet locations are small. The mesic location is cooler on average during the day, because the surface at the wet location dried to a greater extent in the latter part of the measurement period. In the beginning, these two locations were very similar. Differences between all sites are smallest on days with cloud cover, and greatest on clear days with strong solar heating.

Figure 2 illustrates the mean vertical temperature gradient at each location. The temperature differences between the surface and 2 cm depth average about 6 °C at mid-day for all three locations. On sunny days, the differences reach 8-12 °C. At night, the gradients are reversed as the surfaces cool, and the differences average 1-3 °C. The largest temperature gradients occur at the dry location. Note that these variations are similar in magnitude to the horizontal variations (fig. 1).

Although the dry location exhibits the highest surface temperatures, the thermal properties of the peat serve to reduce the soil heat flux. The cumulative soil heat flux at each location is shown in figure 3. This graph represents the heat flux plate readings, but the values have been adjusted to account for errors in the readings. Halliwell and Rouse (1987) have previously examined the sources of errors in heat flux plates in peat soils, and found a significant degree of underestimation. For this study, the plate readings were compared to the flux calculated using the measured temperature gradients and thermal conductivities. This comparison indicated that the plate readings needed to be multiplied by a factor of 1.7-2.5 to give correct values. Figure 3 has included this correction.

The heat flux plate installed at 1 cm at the wet location exhibited variations which are believed to be the result of electrical malfunction, and are not included in fig. 3. However, the 10 cm plate readings indicate that Qg at the wet location is 2-2.5 times greater than at the dry location. Qg at the mesic location (1cm) is nearly as high as at the wet location. The 20 % difference between the 1 cm and 10 cm plates at the dry location indicates the magnitude of errors associated with the heat flux plates and correction factors.

(Divergence of the soil heat flux would lead to the 10 cm readings exceeding the 1 cm reading at the dry location, if any real difference exists.) The difference between the wet and dry locations is much greater than this error, indicating that the measured difference is real.

Model Results

The surface energy balance and ground thermal regime were modelled and the results were compared with the field data. A four day period from July 9 to 12 was selected, and simulated on an hourly basis. This period followed a rainfall on July 8, and was relatively clear, with no precipitation. Most of the surface around the site was wet throughout this period.

The model used is described in detail in Halliwell (1989). Soil temperatures are calculated using a three-time-level numerical scheme outlined by Goodrich (1980). The energy balance calculations are similar to the models described by Outcalt *et al.* (1975) and Smith (1975, 1977), except in the evaporation calculations. Outcalt *et al.* (1975) and Smith (1975, 1977) use relatively simple evaporation models. In reality, the evaporation rate is proportional to the water vapour gradient between the surface and the atmosphere, so knowledge of the surface vapour pressure. However, the use of the surface energy balance concept requires linking the value of QE to T_s. Outcalt *et al.* (1975) do this by describing a surface relative humidity function, so that surface vapour pressure is related to saturation vapour pressure at T_s. Smith (1975, 1977) uses the Priestley-Taylor model (Priestley and Taylor, 1972; Davies and Allen, 1973), which relates QE to Q*, Qg, air temperature (through the slope of the saturation vapour pressure curve) and an empirical coefficient describing surface moisture availability. Smith, therefore, does not explicitly use T_s or surface vapour pressure.

The model in Halliwell (1989) incorporates the difference in temperature between the surface, T_s, and the source of evaporation within the soil, T_i. The importance of this distinction for evaporation from bare soils was recognized by Fuchs and Tanner (1967) and Tanner and Fuchs (1968). Halliwell (1989) expands on their work to define a surface thermal resistance to evaporation, which exists in conjunction with a surface water vapour resistance. A similar resistance is employed by Choudhury and Monteith (1988). Within the evaporation model, surface drying leads to the development of a surface layer which impedes the movement of water vapour from the sub-surface to the atmosphere. In addition, this dry layer impedes the movement of thermal energy from the surface to the source of evaporation. The vertical temperature gradients indicated in figure 2 suggest that this distinction between T_s and T_i is important. This method of incorporating combined thermal and vapour transfer in the surface layer is much simpler than performing vapour transfer calculations in the entire soil column, as is done in Outcalt and Nelson (1985).

Halliwell (1989) assumes that the vapour and thermal resistances always exist in the same ratio, k_r. Therefore, the resistances are r_s (vapour) and $k_r r_s$ (thermal). The case of k_r=0.0 corresponds to no thermal resistance (equivalent to Outcalt *et al.*, 1975, and implied in Smith, 1975, 1977). Typical thermal and vapour properties of peat suggest a value of k_r=0.3. If the soil matrix is non-conductive and all thermal transfer is via diffusion in the pore air, then k_r should be about 1.1. This limits the possible values for k_r. Within the model, for a given value of QE, the difference in temperature between T_s and T_i will increase as k_r increases. T_i is the important temperature as far as the ground thermal regime is concerned.

Figure 4 shows the results of the model, using four values of k_r from 0.0 to 1.1. In the model, r_s is allowed to vary in response to surface evaporation and precipitation. In the absence of surface fluxes of moisture, r_s tends to return to a value consistent with surface soil moisture. The result is a low value of r_s at night, and a high value during the day. The coefficients used to cycle r_s are determined empirically from observed QE values.

Figure 4 demonstrates that QH and QE can be modelled equally well using any value of k_r. However, the details of the evaporation model lead to variations in subsurface temperature for the various values of k_r (fig. 5). The surface temperatures (fig. 5b) are similar for all k_r, consistent with the similar model values for QH. The temperature at the evaporative source (T_i, fig. 5c) shows differences of up to 3-4 °C during peak evaporation periods. At night, the differences are negligible. In comparison with the measured soil temperatures (fig. 5a), the modelled surface temperatures are slightly cooler than the wet surface. The T_i values from the simulations for higher k_r values appear to agree well with the soil temperatures measured at 2 cm in the wet location.

Although the results suggest that the model with high k_r is closely emulating the environment from the wet location (which represents the greatest proportion of the surface at this time), reason argues against acceptance. The wet location was saturated at the surface over this period, so that water should be evaporating from a source at a temperature equal to the surface temperature rather than at some depth (2 cm). The reduction in QE from potential rates (complete surface saturation, with r_s=0) is the result of a reduction in area which can evaporate freely, rather than the presence of a dry layer overlying the evaporation source. In the model, the source of evaporation is displaced vertically from the surface where transfers of Q* and QH occur. In reality, the transition from wet to dry areas corresponds to a shift in available energy from QE to QH, and any difference in effective surface temperature for QH and QE has more to do with horizontal variations. Although the behaviour of the model can be described in physical terms, the agreement between the model and observations is empirical. However, the model used here, which allows a difference in temperature between T_s and T_i, is more realistic than previous models which assume T_s=T_i.

Discussion

The data presented in this paper on surface temperature and soil heat flux are consistent with previous interpretations of the effect of the thermal properties of peat. Soil heat flux at the dry location was less than half the value at the wet

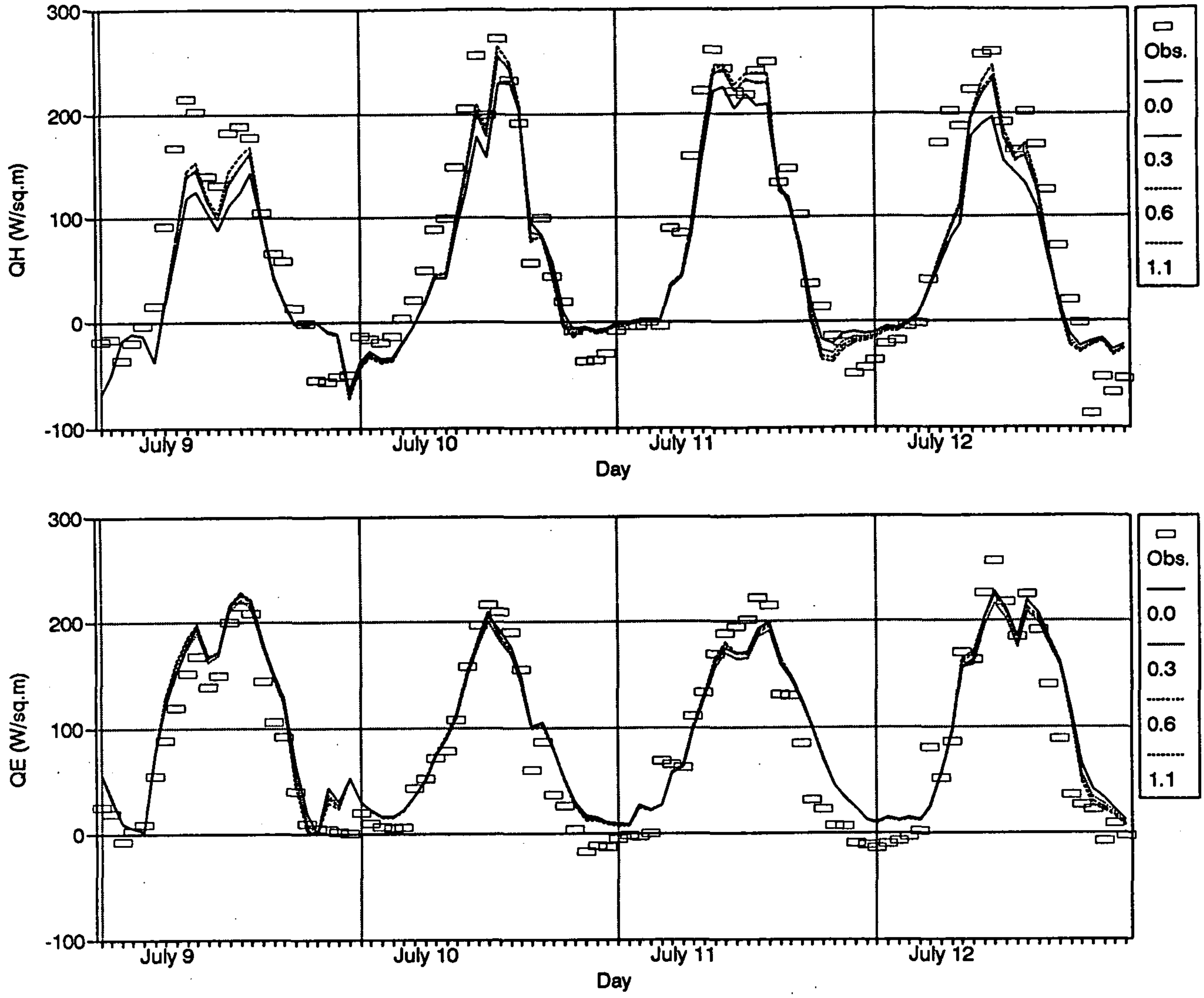

Figure 4. Observed and modelled QH and QE. Modelled values for 4 values of kr.

location, in spite of warmer surface temperatures. The interesting aspect of the data is that the surface drying is confined to a layer of only a few centimetres: both locations have similar water contents at a depth of 5 cm and are saturated at 10 cm. This layer is thin enough that surface evaporation processes will also be affecting total energy fluxes, and an explanation using simple conduction may be inadequate. The presence of a large surface thermal resistance in the evaporation process, as employed in the model, would accentuate the cooling effect of a dry surface layer. This is consistent with Outcalt and Nelson (1985), who also stress the importance of water vapour diffusion in thermal transfer in peat soils.

Differences in surface temperatures between wet and dry locations are large, and of similar magnitude to vertical temperature differences between the surface and 2cm. When these variations are compared to the results from a one-dimensional surface energy balance model, the empirical nature of the model is clear. Although the model can duplicate the energy balance values and provides realistic temperatures, it does not duplicate the two-dimensional physical system. As a result, the sensitivity of a one-dimensional model (especially one which does not account for a surface thermal resistance) may not be consistent with the true sensitivity of permafrost to more extreme changes in surface moisture conditions. Predictions of long-term changes to permafrost based on these simple models should be interpreted with this in mind.

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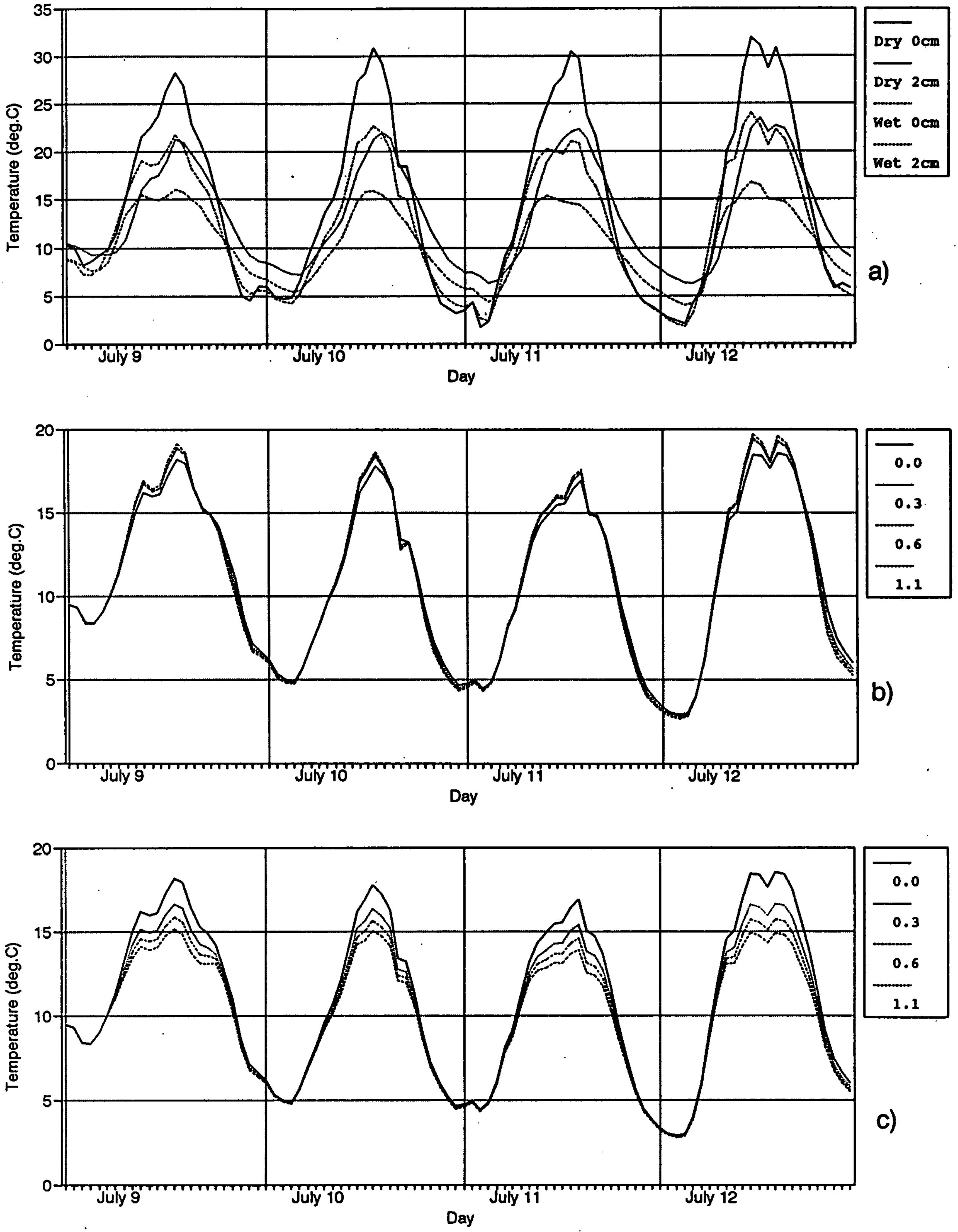

Figure 5. a) Observed surface and 2 cm temperatures ; b) modelled surface temperatures (T_s); c) modelled evaporative temperatures (T_i); for July 9-12, 1987.

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