Hydro-thermal regimes of dry active layer
– first two years observations at a grassland site –

ISHIKAWA Mamoru[1], ZHANG Yinsheng[1], KADOTA Tsutomu[1], YABUKI Hironori[1] and OHATA Tetsuo[1]

[1] Institute of Observational Research for Global Change, Yokohama, Japan

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I Introduction

The widely discussed climate models predict largest increases of surface temperature over Northern Hemisphere land areas from 40-70º, where permafrost is widely distributed. Increases of ground surface temperature will likely result in deeper migration of seasonal thaw depth and degradation of permafrost, potentially leading to substantial changes in conditions for plant growth, gas fluxes, groundwater flow regimes and evaporation in regional, hemispheric and global scales.

The active layer, *i.e.*, the annually freezing and thawing upper ground in permafrost region, is the most sensitive component. Understanding active layer thermal regimes is, therefore, crucial for predicting near-future hydro-biological changes. A number of literatures have been discussed active layer thermal regimes, focusing on the processes of heat and mass transfer dynamics. Although conduction is widely accepted to be dominant heat-transfer mechanism, the importance of non-conductive mechanisms associated with water migration in either liquid and vapor phase, and with water phase changes have been stressed by numerous observational evidences.

This paper addresses the seasonal changes of heat-transfer regimes of active layer at the southern boundaries of Eurasian discontinuous permafrost region, Mongolia, where significant increases in surface temperatures have been observed and thus rapid degradation of permafrost is anticipated.

The observations have started from summer 2002 and are still undertaken. The data currently available and analysis provided us some new aspects of hydro-thermal features of dry active layer of this region.

II Site description and instruments

Observational site is on the wide flat pasture plain of Nalaikh depression (N47º 45′ 07.7″, E107º 20′ 05.2″, 1,441 m a.m.s.l.), about 60 km southwest of Ulaanbaatar. Through 30-m borehole observation, permafrost condition at this site was found below the depth of 5.3 m (Ishikawa *et al*., 2003). Drying core samples taken at every 20 cm depth showed considerable difference in gravimetric water contents between the active layer (less than 10%) and upper permafrost (ranging from 15 to 27%). The permafrost temperatures are extremely warm and nearly thawing point, ranging from 0 to -0.5 °C.

Instruments for hydro-meteorological observations were summarized in Zhang *et al*., (2002). Among them, the most uncertain outputs are derived from soil moisture probe, ECHO-10 (DECAGON Inc.) that outputs voltage value proportional to volumetric water contents (VWCs). Laboratory correlation found good linear correlation between the voltage outputs and VWC (r^2 = 0.98). All observations were initiated at the mid July 2002 and still continue until now. The data until July 2004 will be presented and used for analysis presented later.

III Observational results

1. Atmospheric conditions

The two years observations experienced contrasting weather conditions (Fig. 1). It was dry (63.2 mm in total rainfalls) and low net radiation (39.4 W/m^2 in mean) for the first summer (JAS, 2002), while wetter

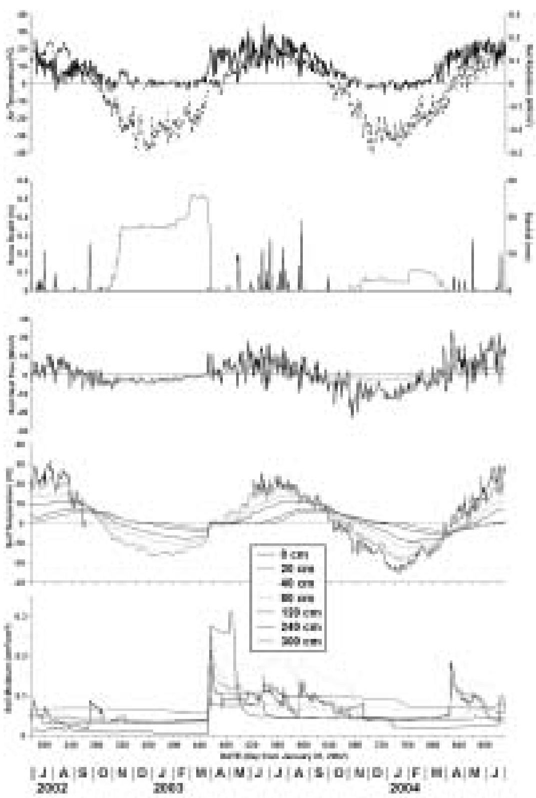

Fig. 1 Data records.

(126.7 mm) and greater radiation (126.7 W/m^2) for the second summer (JAS, 2003). Winter snowfalls also differed between the first (52 cm in maximum depth) and second winter (11 cm).

2. Ground conditions

The contrasting atmospheric conditions significantly affected on the seasonal ground hydro-thermal regimes (Fig. 1). Due to thermal insulating effect by snow cover, soil heat fluxes at 4 cm depths fluctuated negatively for the second winter, while nearly constant for the first winter. Direction of soil heat flux between 4 and 20 cm tended to be upward and downward throughout the second and first winter, respectively. Noticeably mean values of 20 cm were higher in second winter (-1.6 in mean for OND and -1.0 W/m^2 for JFM) than those of the first (-3.5 and -2.3 W/m^2). This trend disagreed with that at 4 cm depth where soil heat fluxes for the second year were higher than those for the first.

Probably due to lower amount of snowfalls, soil temperatures for the second winter were lower than those for the first. Soil temperatures through summer (JAS) were higher for the first year (20.0 and 15.2 °C in mean at 0 cm and between 20 and 80 cm depths, respectively) than those of the second (14.6 and 13.4 °C). These trends also disagreed with those of net radiation amounts.

Soil moisture regimes were strongly controlled by rainfall and snowfall conditions. VWCs were generally low during the first summer, ranging less than 7% through all soil layers. At the end of the first winter, rapid snow melting provided considerable amount of liquid water even into deeper soil layers. Accordingly greater VWCs were observed throughout the second summer (7 to 20%). Detail sequence of rapid snowmelt infiltration is described by Ishikawa *et al.*, (2003).

IV Analysis

Atmospheric heat flux components include net radiation (Q_n), sensible (Q_h) and latent heats (Q_e), are defined here to be positive toward the surface of ground. Summer heat fluxes were estimated by Bowen ratio approach, while winter by Bulk method approach. This is because most energy exchange occurs at the snow surface and that Bowen ratio method needs for snow surface flux that is unsupervised by the heat flux plate sensor installed in the ground layer.

Soil heat flux components can be divided into conductive (ΔH_g), non-conductive (ΔH_l) heats and heat flux supplied by rainfalls (Q_r). ΔH_l includes heat production due to water phase changes and water migration per unit square, and was estimated by considering conservation of thermal energy in a one-dimensional system as formulated by

$$\Delta H_l = c_h \frac{\partial}{\partial t} T - k_h \frac{\partial^2}{\partial z^2} T \qquad (1)$$

where c_h is heat capacity, T is soil temperature, z is depth. For purely conductive transport of heat, the flux is described by Fourier's law

$$\Delta H_g = -k_h (T_{t,Zi+1} - T_{t,Zi}) / (Z_{i+1} - Z_i) \qquad (2)$$

where k_h is the bulk thermal conductivity. $T_{t,\ zi}$ expresses soil temperature at the *i*-th depth layer at the time step *t*. Z_i is the depth of *i*-th layer. The material thermal properties, k_h and c_h depend on volume fractions of the constituting phases as well as on their geometries. The heat capacity is obtained by summing the contribution of the phases,

$$c_h = \sum_m \theta_m \rho_m c_{hm} \qquad (3)$$

where θ, ρ and c are volume fraction, mass density, and heat capacity per unit mass, respectively, of phase *m*, *i.e.* *m* = (*s*; soil particle, *i*; ice, *w*; water). The geometric mean equation is applied to calculate thermal conductivity, forming

$$k_h = k_s^{1-n} k_i^{n-\theta} k_w^{\theta} \qquad (4)$$

where θ is the volume fraction of the unfrozen water per unit soil volume, k_s, k_i and k_w is the thermal conductivity of the soil particle, ice and water, respectively, and n is the porosity.

For unfrozen soils, positive ΔH_l indicates the occurrence of phase change from vapor to liquid (*i.e.* condensation), while negative indicates liquid to vapor (evaporation). For frozen soils, positive ΔH_l indicates from vapor to solid (condensation) and/or liquid to solid (freezing), while negative ΔH_l indicates from solid to vapor (sublimation) and/or solid to liquid (thawing).

Assuming that rain temperature is equal to air temperature (T_a), Q_r was estimated by;

$$Q_r = C_{hw} (T_a - T_s) P_r \qquad (5)$$

where T_s and P_r are ground surface temperature and precipitation, respectively.

V Analytical results and interpretations

1. Summer period

Q_r were small (from -2 to 2 W/m^2) and neglectful, comparing to atmospheric heat fluxes (Fig. 2). Among atmospheric heat fluxes, Q_h dominantly consumed heat during the early summer of dry season. Until July, Q_e started to increase and its general trend was similar to that of Q_h, probably due to evaporative loss of water provided by rainfalls.

Seasonal evolution of Q_h was inversely corresponded to that of ΔH_l at depths up to 60 cm. From early to late summer 2003, originally negative ΔH_l (less than -20 W/m^2) turned to be positive (20 W/m^2), indicating soil internal energy loss by evaporation during early summer and energy gain by condensation during late summer. Similar trends were found during late summer 2002 and early summer 2004. Soil internal evaporation occur during the period of Q_h increasing, while condensation during Q_h decreasing.

Above findings indicate that soil internal evaporation and condensation processes are determined by soil-atmospheric temperature gradient that is proportional to the amount of Q_h. When soil temperatures are relatively lower than atmospheric temperatures, downward heat fluxes promote soil warming, increase relative humidity of voids between soil particles, and enable soil internal evaporation. When soil temperatures are relatively high, on the other hand, upward heat fluxes promote soil cooling, decreases relative humidity of voids, and enable soil internal condensation.

Unlike to shallower depths, soil moisture condition inversely controlled by summer ΔH_l at the deeper soil layers. ΔH_l during JAS 2002 tended to be negative (from -4 to 0 W/m^2), while positive during JAS 2003 (0 to 1 W/m^2). Accordingly, evaporative energy losses were dominated for the first summer and that condensation energy gain for the second summer, although VWCs in the first summer were higher than those in the second summer. These facts indicate that evaporation is more dominant for dry soils than for wet soils.

2. Freezing/Thawing Periods

Differences of VWCs between two years were reflected in those of ΔH_l during freezing period. More ΔH_l were produced at the beginning of second winter after soil temperatures turned to be positive. This is due to high amount of soil water before the onset of freezing.

It is difficult to compare quantitatively ΔH_l between the first and second year because of rapid snowmelt infiltration only in the first year. Temporary positive ΔH_l at some depths near the thawing front of first year suggest refreezing of snowmelt-derived water infiltrated.

3. Winter periods

Although winter soils hydro-thermal states are more complex, including freezing and thawing processes, ΔH_l evolved in a similar manner with summer. Shallower ΔH_l during winter 2003/2004 turned from positive (more than 20 W/m^2) to nearly zero at the end of January when snow cover thickness has sharply increased. Snow thermal insulating effect changed soil-snow-atmospheric temperature gradient to be smaller, resulting in upward soil heat fluxes. Accordingly amount of energy gain due to condensation has decreased.

ΔH_l at deeper soil layers evolved inversely in accordance with soil water (ice) contents. ΔH_l for the first winter were generally higher than those for the second winter, although ice contents for the second winter were higher than those for the first.

Identification of phase changes considered, comparing temporal changes of ΔH_l and sublimation/condensation-derived heat production (ΔH_{lv}) calculated by

$$\Delta H_{lv} = \Delta H_l - L_{sl}\rho_w \frac{\partial}{\partial t}\theta_{ECHO} \tag{6}$$

where L_{sl} is latent heat for phase change from liquid to solid, ρ_w is water density and θ_{ECHO} is the liquid water contents derived from ECHO probe outputs. Discrepancies in the amount between ΔH_l and ΔH_{lv} correspond to heats produced by freezing and thawing processes.

It was found that freezing and thawing were minor contribution for ΔH_l throughout the first winter (Fig. 3). For the second winter, these contributions were limited at the beginning and end of the periods. Most of ΔH_l were resulted from solid water vaporization such as sublimation and condensation.

4. Contribution of non-conductive heat

Fig. 4 shows the dimensionless ratio between contributions to thermal dynamics originating from production of non-conductive and from conductive heats ($\ln|\Delta H_l/\Delta H_g|$). Comparing to conductive heat, non-conductive heats were predominant for the shallower depth throughout the period and for entire

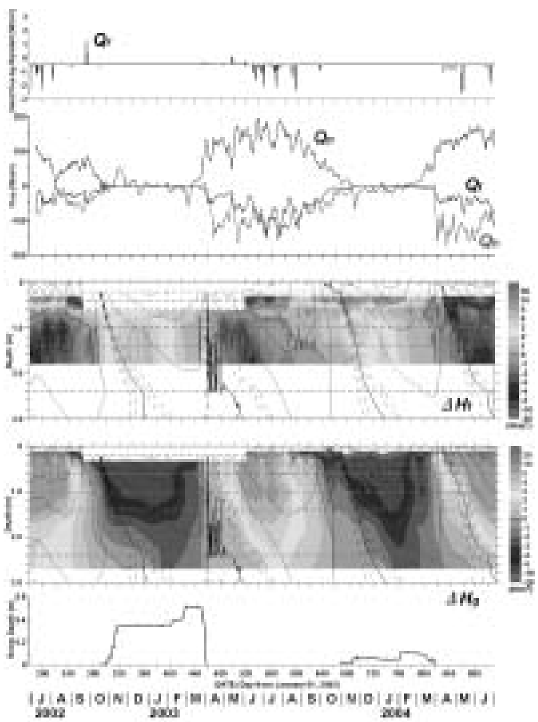

Fig. 2 Evolution of atmospheric and soil heat fluxes.

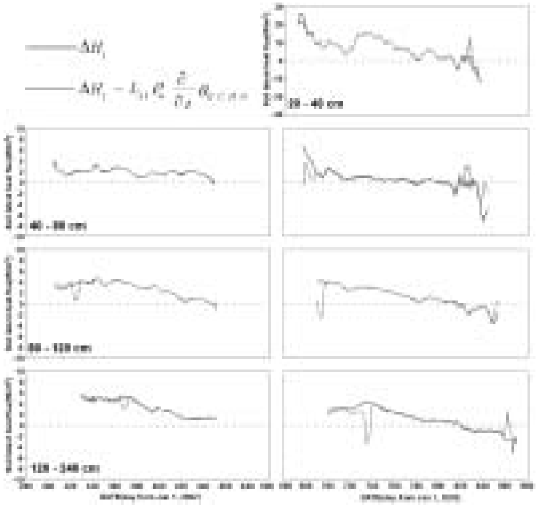

Fig. 3 Nature of winter non-conductive heat fluxes.

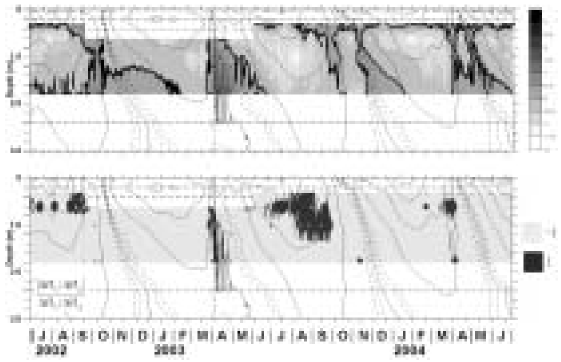

Fig. 4 Contribution of non-conductive heat.

depths during the first and second thawing, and second freezing periods. This was well agreed with soil water variations.

Non-conductive heats mostly prevented soils warming or cooling that were promoted by conductive heat throughout the period (Fig. 4). Two exceptions have occurred at the first thawing period of March 2003 and during late summer 2003. The former is doubtlessly due to snow melting infiltration, by which temperatures of all soil have been increased.

III Discussion and implication

The dry active layers contain large percent of open and closed pores. Rapid snowmelt infiltration during the first thawing period is the evidence for the occurrence of large open pores passing to the atmosphere. Soil internal vaporization might occur through open pores and within the closed pores. This is because that soil internal evaporation and sublimation were more pronounced when soils were drier.

This is distinctive hydro-thermal feature of dry active layer. It has been widely accepted that freezing and thawing are the main contributions of non-conductive heat production at circumpolar regions, where active layer were generally wet and its thickness is small. Due to semiarid climate conditions, the active layers in the southern boundary of Eurasian permafrost regions were extremely dry and sometimes reach to several meter in thickness. Freezing and thawing are the second importance. Soil internal vaporization consume nearly ten times larger amount of latent heat than solidification, and should be considered for modeling soil hydro-thermal dynamics of this region.

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