

# Land-surface hydrological processes in discontinuous permafrost region of the Western Tibetan Plateau

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**ABSTRACT:** In order to examine and record the effects of permafrost on hydrological processes in the cryosphere, the hydrology of the ground surface layer is investigated in the region of western Tibetan Plateau, a discontinuous permafrost region. The thermal/water regime is described. Water-budget components in the surface layer were calculated on a daily basis. Variability of the water-budget components, and its causes, are investigated in both seasonal and daily terms.

## 1 INTRODUCTION

The Tibetan Plateau, the most important example of a high-altitude cryosphere, is considered a region sensitive to global warming. Energy and water cycles over the Tibetan Plateau play an important role in the Asian Monsoon system, which in turn is a major component of both the energy and water cycles of the global climate. Of particular importance is the melt season, when snowmelt and evaporation occur simultaneously, but in different parts of the region. Through its effects on hydrology, the presence of permafrost increases the complexities of the land-surface hydrological processes through contrasting thermal and water regimes in the ground. Differential heating and evaporation, together with advection, are amplified by the hydrology of the permafrost terrain.

In a permafrost-dominated region such as the Tibetan Plateau, permafrost and the active layer control soil moisture and intimately link hydrology to the thermal balance of the soil. The active-layer thickness and permafrost conditions are largely controlled by surface heat fluxes, coupling the hydrological process in the ground surface layer to the surface energy budget to the extent that they cannot be quantified separately. Seasonality of the energy budget leads to alternating freeze-thaw of soil. These in turn interact with surface and sub-surface hydrological processes, playing a different role at different time-scales in the water cycle.

Observations of components of the hydrologic cycle, including precipitation, discharge, soil wetness and ground-surface meteorology conditions, have been carried out for several years on the Central Tibetan Plateau under the auspices of the GAME/Tibet project (Koike 2000). The atmospheric processes over the permafrost have been documented, but study of the hydrologic processes in the ground and their impact on the water cycle remains incomplete. This paper examines hydrologic processes in the ground surface layer and their impacts on the water cycle.

## 2 STUDYING REGION, DATA AND CALCULATION

### 2.1 *Studying region*

Data was collected at Gaize, located in western Tibetan Plateau ( $32^{\circ}18' N$ ,  $84^{\circ}03' E$ , 4420 m a.s.l.; Fig. 1). The ground surface of the observation site is almost bare soil covered by a few short grasses. Discontinuous permafrost develops in the region (Zhou et al. 2000). The study site is located in a rather dry region. In the hydrological year 1 October 2000 to 31 September 2001, precipitation was 169.4 mm, and the annual mean air temperature, relative humidity, and wind speed were  $0.2^{\circ}C$ , 32%, and 1.8 m/s, respectively.

Due to low and intermittent precipitation in the non-monsoon season, the snow cover on the Tibetan Plateau is irregular and discontinuously distributed over the ground surface, even in the middle of winter (Sato 2001). At the Gaize site, just 2.0 mm precipitation was observed during the winter half-year (November to next April). Therefore, the hydrological impact of snow cover was minimal and could be neglected.

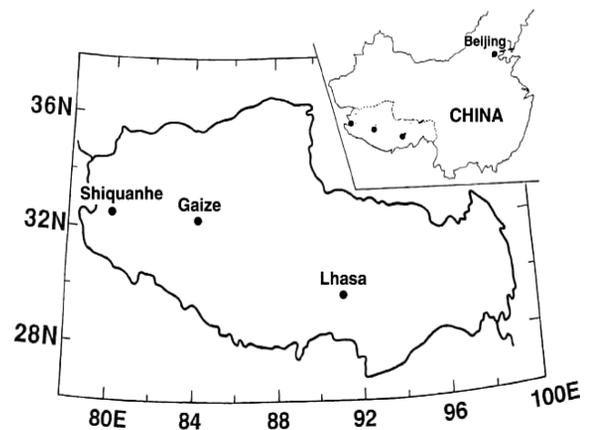


Figure 1. Location of study site Gaize.

## 2.2 Data

The data from the study site were obtained using three observation systems: an automatic weather station (AWS), a soil moisture and temperature monitoring system (SMTMS), and a precipitation gauge. The parameters measured by the AWS included air temperature ( $T_a$ ), wind speed ( $U$ ), air pressure ( $P$ ), and relative humidity ( $Rh$ ), all observed at 3.60 and 0.50 m above the ground and ground surface temperature ( $T_s$ ) and net radiation ( $Q_n$ ) observed at 1.50 m above and at ground surface. Volumetric soil water content ( $\theta$ ) and soil temperature at several depths ( $T_g$ ) were measured, using SMTMS, at depths of 0.04, 0.20, 0.60, 1.00, 1.60, 2.00, 2.60 m. Ground heat flux was measured at a depth of 0.10 m. Some of the precipitation data ( $Pr$ ) from a nearby meteorological station was used to extend the data series.

In this work,  $\theta$  denotes volumetric soil water content relative to the total volume of soil including dry matter. When soil temperature was above 0°C,  $\theta$  denotes volumetric soil liquid water content, which can be calculated from the TDR data. During the period when the soil temperature was below 0°C,  $\theta$  equals volumetric liquid water plus ice content, which was assumed to be the same as the observed liquid water content in the preceding fall freezing, as long as no significant change was seen in TDR data.

The same method was applied in the permafrost region of eastern Siberia (Sugimoto et al. 2001).

## 2.3 Calculation

The water budget in a soil layer from ground surface to a given depth can be expressed as:

$$\Delta W = P_r - E - (WF_H + WF_V) \quad (1)$$

where  $P_r$  is precipitation (mm),  $E$  is evapotranspiration (mm),  $WF_H$  is net lateral water fluxes both at surface and sub-surface (mm),  $WF_V$  is the vertical water flux at bottom of soil layer, and  $\Delta W$  is the water storage changes in the surface soil layer (mm). This can be written as  $\Delta W = \Delta W_L + \Delta W_S$ , subscript  $L$  and  $S$  denote liquid and solid water respectively.  $E$  is computed by the bulk formula (Brutsaert 1984):

$$E = \rho \beta C_E U (q_0 - q) \quad (2)$$

where  $\rho$  is air density,  $U$  is wind speed,  $q_0$  and  $q$  are the saturated and observed specific humidity respectively,  $\beta$  is a parameter determined by surface soil water content, which can be defined as (Stone & Quirr 1977):

$$\beta = \frac{\theta_{av} - \theta_{ec}}{\theta_{sat} - \theta_{ec}} \quad (3)$$

where  $\theta_{av}$  is the mean soil moisture averaged from two layers (at which depth),  $\theta_{ec}$  is a criterion for meeting the demands of soil evaporation. This is determined by the soil moisture data when soil is close to freezing, and it has been deduced in the range of 6.51–8.97% in the Tibetan Plateau region (Li et al. 2000).  $\theta_{sat}$  is saturated soil moisture, and it has been deduced in range of 33–37% in the Tibetan Plateau region (Li et al. 2000).  $C_E$  is the bulk vapor transfer coefficient that can be calculated from (Brutsaert 1984):

$$C_E = k^2 \left[ \ln \left( \frac{z_1}{z_2} \right) - \Psi_w \right]^{-1} \left[ \ln \left( \frac{z_1}{z_2} \right) - \Psi_m \right]^{-1} \quad (4)$$

where  $k$  is the von Kármán constant (0.4),  $z$  is height and subscripts 1 and 2 denote instrument heights of 3.6 and 0.5 m,  $\Psi_w$  and  $\Psi_m$  are stratification functions, which depend on atmospheric stability. To calculate  $\Delta W$ , the ground surface layer was divided into several sub-layers at 10 cm depth intervals. Water content  $\theta$  was calculated at the middle depth of each layer, using the profile of soil water content from SMTMS.

The vertical water flux at depth of 2 m ( $WF_V$ ) was determined by the gradient of soil-water potential (Kondo & Xu 1996):

$$WF_V = -\rho_w k_w \frac{\partial \phi}{\partial z} \quad (5)$$

where  $\phi$  is soil-water potential,  $\rho_w$  is water density,  $k_w$  is hydraulic conductivity, and  $\phi$  is the sum of the matric and gravity potentials. The latter is equal to the depth  $-z$ ; the matric potential is a function of the volumetric water content of the form  $a \exp(b\theta)$  (Liu et al. 1999), where the coefficients  $a$  and  $b$  are determined from an empirical relationship between matric potential and volumetric water content applicable to the Tibetan Plateau, originally derived for water and heat balance estimation (Xu & Haginoya 2001). The calculation of  $k_w$  follows the work of Iwata et al. (1995) and Kondo & Xu (1996).  $WF_V$  was calculated at bottom depth of 2.00 m, and assumed to be zero when the soil was frozen. The TDR data was used for estimating soil water potential gradient.

$Ep$ , the potential evaporation, is given by:

$$Ep = \frac{\Delta}{\Delta + \gamma} Q_{ne} + \frac{\gamma}{\Delta + \gamma} Ea \quad (6)$$

where  $\Delta$  is the slope of the saturation water vapour pressure curve.  $\gamma$  is given by:

$$\gamma = C_p P / 0.622 L_e \quad (7)$$

where  $C_p$  is the specific heat of air at constant pressure,  $P$  is air pressure, and  $L_e$  is latent heat of vaporisation of water.  $Q_{ne}$  indicates the heat energy available for evaporation, determined by the net radiation ( $Q_n$ ) and ground heat flux ( $Q_g$ ):

$$Q_{ne} = (Q_n - Q_g)/L_e \quad (8)$$

$E_a$  is a function of wind speed ( $U$ ) and saturation deficit ( $e_0 - e$ ):

$$E_a = 0.26(1 + 0.45U)(e_0 - e) \quad (9)$$

### 3 RESULTS AND ANALYSIS

#### 3.1 Soil thermal/water regime

Contours of soil temperature and volumetric soil water content at the study site from October 2000 to October 2001, down to 260/278 cm depth, are shown in Fig. 2. The ground starts to freeze in November and reaches a maximum frost depth in March. Compared to the freezing process, the ground-thawing process was rather fast at the study site, starting at the surface at the beginning of April. Ground melt of whole layer was completed by the end of April.

The effects of the soil thermal regime on water content variations are seen by comparing the upper and lower panel of Fig. 2; the latter shows the

spatial-temporal variation of liquid volumetric soil water content at the study site. The soil became wetter as the ground thawed, and drier as the ground froze, which lead to clear seasonal variation in soil moisture. Vertical profile of soil water content was irregular as the ground thawed, with layers of high water content observed at 0.60 and 2.00 m, respectively.

#### 3.2 Water budget of ground surface layer

The accumulative values of water budget components in the ground surface layer for the hydrological year Oct. 2000 to Sept. 2001 are shown in Fig. 3. Due to the monsoonal climate, precipitation occurred predominantly in the summer monsoon period (May to September). Summer precipitation constituted 99% of the annual total at the site. The annual precipitation of 169.4 mm was much smaller compared to the eastern part of the Plateau (300–600 mm/yr).

Evaporation showed clear seasonal variation related to precipitation. The monthly evaporation exceeded precipitation except during the period from June to August. On an annual basis, precipitation was exceeded evaporation by about 42 mm. The total amount of  $\Delta W$  reveals that the water storage in the soil increased 50.1 mm over the year. The vertical water flux ( $WF_v$ ) at depth of 2 m totalled 6.2 mm over the year, and was therefore negligible in the water balance.

#### 3.3 Evaporation and soil moisture

Obvious seasonal variations of evaporation are related to the ground surface condition, soil thermal processes, and precipitation. During the period of ground surface freezing (October to March), sublimation from the ground surface was small, with a daily sum of less than 0.8 mm. At the beginning of the ground surface thaw period (April to May), evaporation was quite large, as a

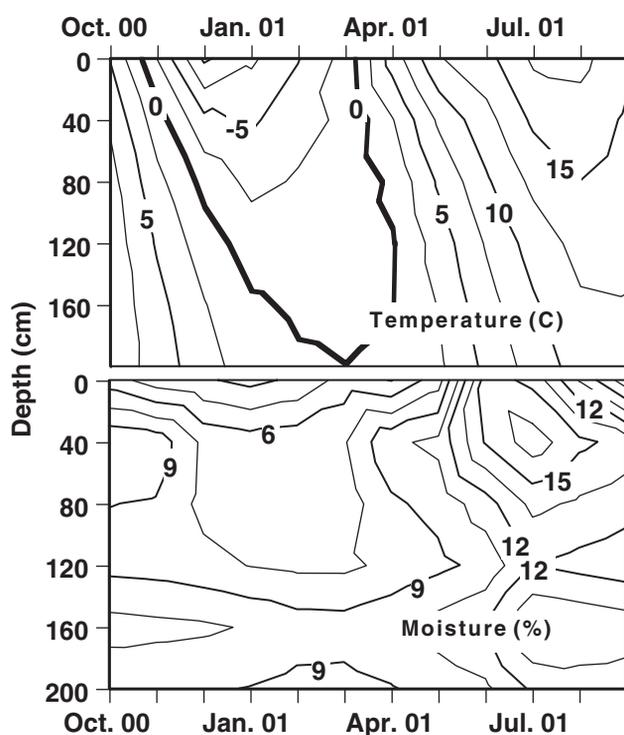


Figure 2. Contours of soil temperature and volumetric liquid and solid at Gaize site, western Tibetan Plateau.

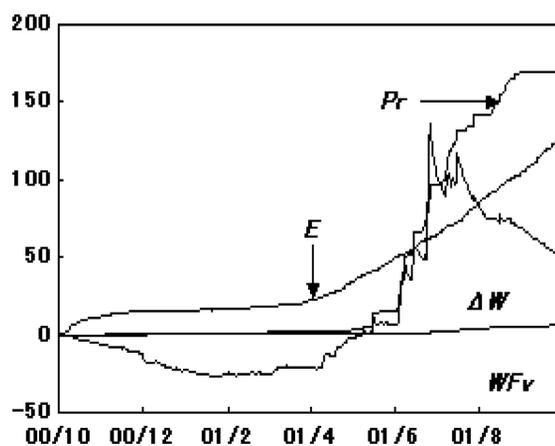


Figure 3. Accumulative values of water budget components for Oct. 2000 to Sept. 2001 (mm) for Gaize site.

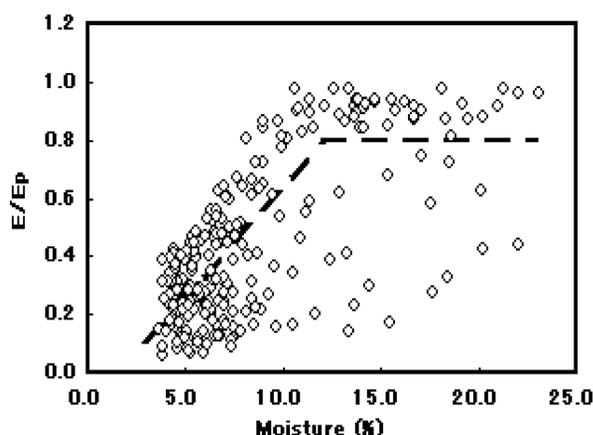


Figure 4. The variation of ratio of daily evaporation to potential evaporation ( $E/E_p$ ) versus volumetric surface moisture.

result of the thaw of the very thin layer near the ground surface.

The ratio of actual evaporation to potential evaporation ( $E/E_p$ ) is a measure of the extent to which the land surface exerts control over the evaporation process. To investigate the dependence of evaporation efficiency on the availability of stored water, the variation of  $E/E_p$  with surface soil moisture (at 0.04 m) is shown in Fig. 4. It is clear that  $E/E_p$  increased linearly with ground surface moisture when the volumetric water content was less than 30%, but shows no tendency to increase with moisture content beyond that level. This stresses the importance of the critical value of soil moisture of 30%.

### 3.4 Changes in water storage and soil moisture

Precipitation and evaporation are complementary components of the water balance, being the input from the atmosphere, and the return flow from the ground back to the atmosphere. The difference between precipitation ( $Pr$ ) and evaporation ( $E$ ) should reflect the change in water stored in the ground surface layer ( $\Delta W$ ), and the value of  $\Delta W$  must therefore be related to the difference between  $Pr$  and  $E$ . The variation in  $\Delta W$  versus ( $Pr-E$ ) at the study site is shown in Fig. 5. The change in water storage generally increased with ( $Pr-E$ ). It was not expected that the water stored in the ground surface layer would react linearly to atmosphere forcing, but such a result might provide a simple parameterisation for model development.

### 3.5 Water budget and water cycle

To improve understanding of the significance of hydrological processes through studying the interaction between the ground surface layer and the atmosphere,

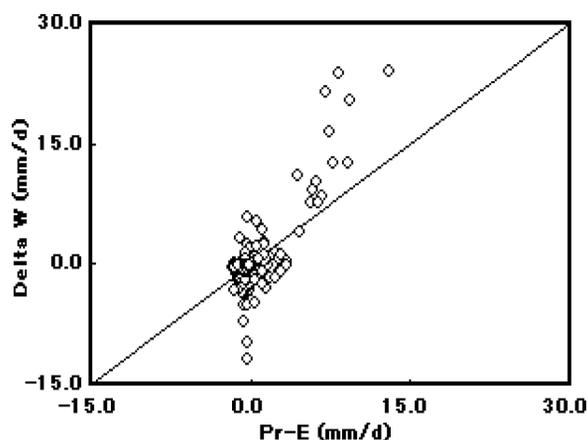


Figure 5. Changes in water storage versus  $Pr-E$ , (mm) at Gaize, western Tibetan Plateau.

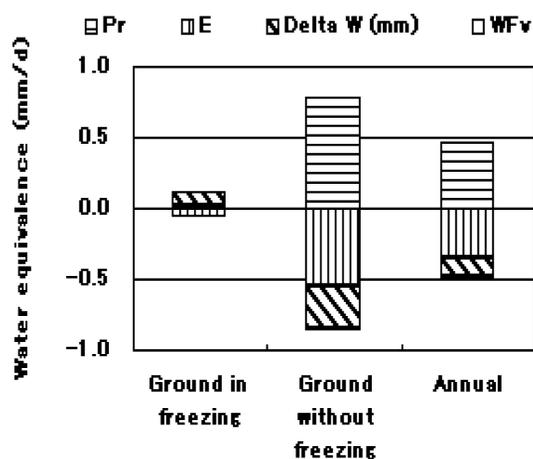


Figure 6. Water budget of the ground surface layer (0–2.00 m) at Gaize, western Tibetan Plateau.

the hydrological year was divided into two periods according to the seasonal state of the ground surface thermal regime. These are the ground-frozen period (Nov. to Mar.) and ground-thawed period (Apr. to Oct.). To illustrate how the water budget is composed in annual terms, and over these seasonal periods, the composition is given in Fig. 6. Amounts were normalised to a daily time scale to permit comparison.

During the ground-frozen phase, there was little change of water storage (0.1 mm/d). Upward vapour flow occurred in the form of sublimation amounting to 0.1 mm/d. When the active-layer was completely frozen, the water cycle almost stopped, except for little sublimation. As the ground began to thaw from the surface, the water cycle became dashing: precipitation averaged 0.8 mm/d coupled to evaporation of 0.5 mm/d and an increase in water storage of 0.3 mm/d. Regarding the annual water budget of the surface ground layer, precipitation and evaporation were the dominant components, accounting for 100% and –70% respectively.

#### 4 CONCLUSIONS

The water movement in the surface ground layer (0 to 2.00 m) in the discontinuous permafrost region of the western Tibetan Plateau is predominantly controlled by the thaw-frost cycle. When ground starts melting, soil moisture increases due to precipitation relate to the summer monsoon climate.

At the commencement of ground surface thaw, the concentration of water within a rather thin surface layer leads to a rapid increase in evaporation. Evaporation is then reduced when the wet soil zone moved downwards.

In the dry permafrost region, with annual precipitation less than 300 mm, the vertical water flux is too small to affect the water budget.

The freeze-thaw cycle, which affects seasonal soil moisture, water storage, evaporation, and the mobilisation of water through soil and vegetation during the summer monsoon season, is a dominant feature of the land-surface hydrology in the permafrost region of the eastern Tibetan Plateau.

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