

## Evaporation of river water in West Kunlun Mountains, China

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### Abstract

During a Sino-Japanese joint expedition in 1985, evaporation was observed to be at a rate of about  $0.4 \mu\text{m s}^{-1}$  from the water surface in the West Kunlun Mountains. Heat balance observation indicated that the latent heat loss through evaporation was roughly balanced with the short wave radiation which accounted for most of the heat source. The fractionation factor of oxygen isotope for evaporation was obtained to be about 1.03 in association with the kinetic effect.

### 1. Introduction

The West Kunlun Mountains are located in a vast dry area, but a number of glaciers exist there. It is still in question why the glaciers have developed in the arid region, and how the glaciers can be nourished. Either by the local water cycle, or by the water vapor from the west? Since the region is in the drought, evaporation from the surfaces of rivers, lakes and glaciers should be intense. It is important to estimate the amount of evaporation from the local area in order to examine the water source feeding the glaciers. This paper describes briefly the role of evaporation in the West Kunlun Mountains.

### 2. Observations

Detailed observations were made during the daytime on July 23, 1985 near a camp site at an elevation of 5112 m above sea level, which was located close to the Kunlun Peak, the highest mountain in the Kunlun Mountains. Meteorological variables measured were incoming and reflected solar radiation, air temperature, humidity and wind speed at a height of 1.5 m, wind direction, cloud type, and cloud amount. These observations were made at about every hour. In addition, the following data were collected: soil tempera-

ture at a depth of 70 mm and at the surface, and water temperature of the river, which flowed toward south with a width of about 1 m. The water depth was a few tens of a mm and the flow speed was about  $0.2 \text{ m s}^{-1}$ . The river was described in detail by Nakawo and Watanabe (1987).

Figure 1 shows the observed data of these variables. The time-scale is given in Chinese Standard Time, which is about 2.5 hours ahead of Local Mean Time at the observation site.

It was a relatively fine day on July 23, 1985: almost no cloud except for a rather short period in the evening. The variation of incoming solar radiation hence seems to depend only on the zenith angle of the sun, exhibiting a typical curve for a fine day. The maximum value was about  $1 \text{ kW m}^{-2}$  at 2 to 3 p.m., around noon, Local Mean Time. A small disturbance can be seen at around 4 to 5 p.m., which could have caused a sharp drop in soil surface temperature.

Air temperature varied in a range of 0 to  $13^\circ\text{C}$ , showing a maximum at around 5 to 6 p.m., which lagged by about 3 hours behind the peak of solar radiation. The relative humidity decreased in the morning and increased in the evening: an inverse trend of the air temperature. Absolute humidity was hence rather stable, roughly 0.7 kPa.

Wind speed was about  $10 \text{ m s}^{-1}$  in the early

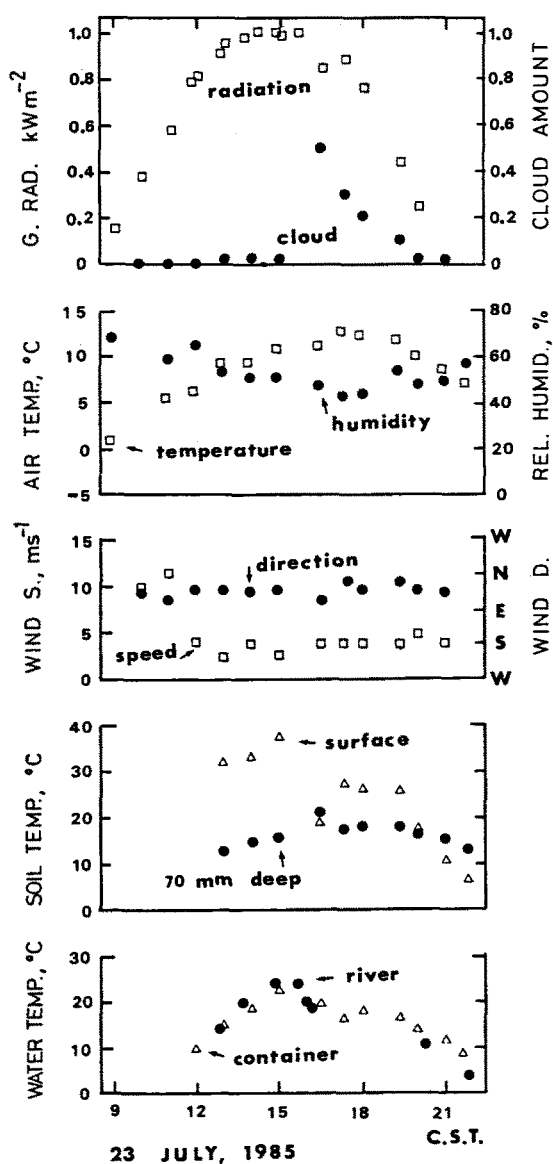


Fig. 1. Variation of meteorological variables and the temperatures of soil and river water.

morning but it decreased at around 11 p.m., and kept a value of 3 to 4  $\text{m s}^{-1}$  afterwards. Wind direction was almost constant, around NE, throughout the day of the observation.

Surface temperature of the soil seemed to depend mainly on radiation rather than air temperature, since the peak value of surface temperature, which reached as high as 40 °C, was recorded at around noon of LMT.

At a depth of 70 mm, however, the maximum value was observed at 5 to 6 p.m., resulting from the thermal diffusion process through the soil. The temperature gradient at the surface was about 300  $\text{deg m}^{-1}$  around noon, which decreased gradually until the surface temperature became lower than the temperature at the depth of 70 mm. A negative gradient developed afterwards. The temperature of the river water varied with a similar trend to the surface temperature of the soil.

### 3. Evaporation

A water filled container was placed at the soil surface, and the change in the water level was measured for estimating the amount of evaporation from the water surface. Figure 2 represents the obtained rate of evaporation. It showed a negative value (condensation) in the morning, but it increased to positive soon after: the mean value of the rate was about 0.4  $\mu\text{m s}^{-1}$  (or  $\text{g s}^{-1}\text{m}^{-2}$ ) in the afternoon. The rate varied with time, as can be seen in Fig. 2. The variation, however, may not be very significant, since the error of the observation was as large as 0.1  $\mu\text{m s}^{-1}$ .

The temperature of the river water was nearly equal to the water temperature in the container as shown in Fig. 1. It is considered, hence, that the rate of evaporation from the river would be similar to the rate observed with the container.

The rate of evaporation was also estimated by solving a two-dimensional turbulent diffusion model numerically. The following assumptions were made for the calculation: 1) steady flow of air crossing a river of 1 m wide orthogonally; 2) the specific humidity of the upwind air is independent on height from the surface, *i.e.*, no vertical flux of moisture before the air reaches the river. This assumption is based on an observation that the surface soil was very dry; 3) specific humidity at the river surface is given by the humidity at saturation corresponding with the water temperature.

Continuity equation for the water vapor is given by

$$u \frac{\partial q}{\partial x} = \frac{\partial}{\partial z} (K_E \frac{\partial q}{\partial z}) \quad (1)$$

where  $x$  is the distance along the wind direction with its origin at the upwind edge of the river,  $z$  is the

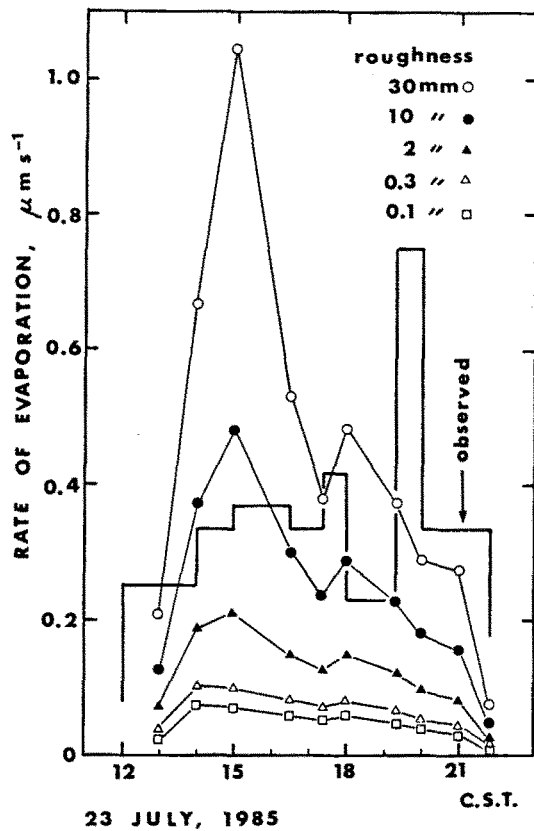


Fig. 2. Rate of evaporation from the river surface.

height from the surface,  $u$  is wind speed,  $q$  is specific humidity, and  $K_E$  is the turbulent eddy diffusivity for vapor. It was assumed that  $u$  is independent of  $x$ , and given by

$$u = \frac{u_*}{k} \left( \ln \frac{z+z_0}{z_0} + \frac{kz}{\lambda} \right) \quad (2)$$

where  $z_0$  is the roughness length (taken to be constant over the soil and the river),  $u_*$  is the friction velocity (determined by  $z_0$  and the measured wind speed at 1.5 m in height),  $k$  is von Karman's constant (0.4),  $\lambda$  is a scaling height which gives an upper bound on  $K_E$  to avoid an unrealistic increase of  $K_E$  outside the internal boundary layer (Blackadar, 1962). According to a mixing length theory (Monin and Yaglom, 1971), the diffusivity  $K_E$  is given by

$$K_E = ku_*z/\phi_E \quad (3)$$

where  $\phi_E$  is the dimensionless vertical gradient of specific humidity expressed as a function of Monin-Obukhov's length  $L$ . Using the notation of  $\xi = z/L$ ,

$$\phi_E = 1 + 5.2\xi \quad (4)$$

for stable conditions (Webb, 1970), and

$$\phi_E = (1 - 16\xi)^{-1/2} \quad (5)$$

for unstable conditions (Businger, 1966; Dyer and Hicks, 1970). The calculation showed that the air stratification was mostly stable over the river surface except for the parts near the upwind edge of the river at the end of the day.

With the data given in Fig. 1, the rate of evaporation was estimated using the above model. The estimation was made only for the afternoon because of a lack of data on soil and water temperatures in the morning. The results of the calculation are shown in Fig. 2, where the rate of evaporation averaged over the width of the river is presented for various roughness length, an unknown parameter between 0.1 mm and 30 mm.

In comparison with the direct measurement with a water-filled container, a value of 10 mm for  $z_0$  seems to predict the rate of evaporation most compatible with the observation. The roughness length of 10 mm may be rather large, but would not be unrealistic. The river surface was relatively rough, which was caused by bottom undulations of the river, since the water depth was only a few centimeters. Since there grass was growing sporadically to a height of several centimeters over the soil surface, the value of 10 mm would be also realistic for roughness length.

#### 4. Heat budget

Heat budget at the surfaces of the soil and the river was considered to have the following form.

$$Q_S + Q_L + Q_C + Q_E + Q_H + Q_R = 0 \quad (6)$$

where  $Q_S$  is short wave radiation balance,  $Q_L$  long wave radiation balance,  $Q_C$  conductive heat flux,  $Q_E$  latent heat flux,  $Q_H$  sensible heat flux, and  $Q_R$  the remainder term. Here the fluxes toward the surface are taken positive.

The short wave radiation balance was obtained by direct measurements of incoming solar radiation

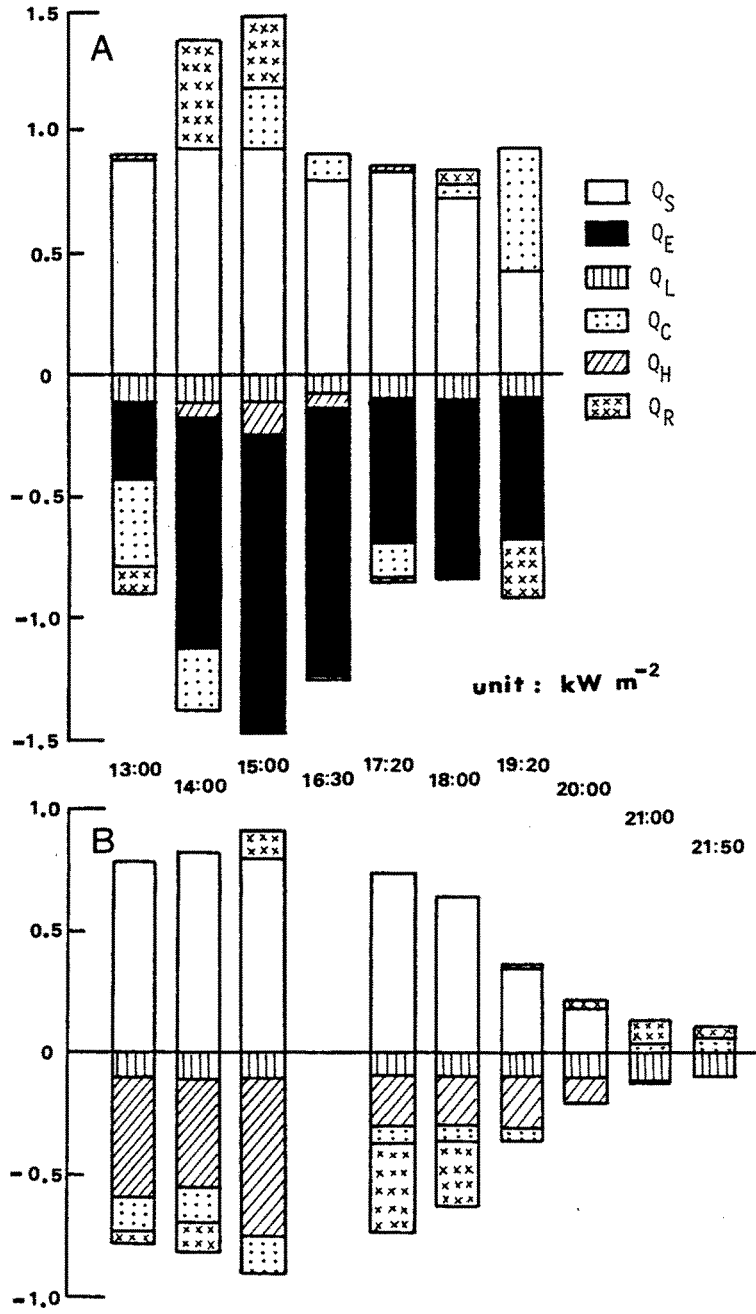


Fig. 3. Heat balance at the river surface (A), and at the soil surface (B).

(Fig. 1), and the reflected radiation at the soil and the river surfaces. The albedo of the soil surface was about 0.18 for most of the day, but increased up to 0.27 near the sunset, presumably owing to the increase of the solar zenith angle. The albedo of the river was about 0.06, which also increased up to 0.07 toward the end of the day. The reflection, however, would have taken place not only at the water surface but also at the river bed. This will be discussed later.

The long wave radiation balance was calculated by the following equation given by Kondo(1976).

$$Q_L = \epsilon \sigma T_a^4 (0.066 \sqrt{e_a} - 0.49) (1 - nch) \quad (7)$$

where  $T_a$  is air temperature in K,  $e_a$  vapor pressure in mb,  $\epsilon$  emissivity (taken to be 0.90 and 0.95 for the soil and the water surfaces respectively),  $\sigma$  Stefan-Boltzmann constant,  $n$  cloud amount in tenth. The value of  $c$  is given as a function of  $e_a$ , and  $h$  is a function of cloud amount, cloud type and the frequency of precipitation. The conductive heat flux in the soil was calculated using the temperature data at depths of 70 mm and zero (surface) given in Fig. 1, with an assumption of linear temperature profile. The thermal conductivity of the soil was assumed to be  $0.5 \text{ W m}^{-1}\text{K}^{-1}$ , which corresponds with the sand soil of 40 % in pore space and a few per cent in water content.

For the river, it was difficult to estimate the conductive heat flux. It was considered conventionally that a change in stored heat in the river water is the result of "heat conduction", in which the heat convection is included. This would not be correct because the river water was warmed up not only by the conduction/convection but also by the extinction of short wave radiation. The solar radiation would also heat the river bed, and the water should be warmed up through conduction/convection from the bed. Short wave radiation balance, however, was overestimated by the amount of the absorption at the river bed and the river water. It would be reasonable, hence, to subtract the heat as a part of the "heat conduction". An essential assumption, through this procedure, is that the conductive heat below the river bed can be neglected. This assumption may cause an underestimation in "conductive heat flux", in particular around noon LMT, while the temperature could be high at the river bed because of the strong solar radiation.

The latent heat flux was estimated, for the river surface, by solving the two-dimensional turbulent diffusion equation for water vapor numerically, as

was described in the previous section. The roughness length of 10 mm was used in the calculation. For the soil surface, it was assumed that the vapor pressure was independent of height, *i.e.*, the latent heat flux was zero.

Also, the sensible heat flux was calculated with the two-dimensional turbulent diffusion equation with the roughness length of 10 mm, in which

$$u \frac{\partial T_a}{\partial x} = \frac{\partial}{\partial z} (K_H \frac{\partial T_a}{\partial z}) \quad (8)$$

where  $K_H$  is the turbulent eddy diffusivity for heat. Based upon the semi-empirical similarity model (Monin and Yaglom, 1971),

$$K_H = ku_* z / \phi_H \quad (9)$$

The calculation was made with  $\phi_H = \phi_E$ , which has been adopted by many workers (*e.g.* Webb, 1970 ; Businger, 1966 ; Dyer and Hicks, 1970).

Variations of the heat balance components thus estimated are shown in Fig. 3. The major heat source is the short wave radiation both at the water and the soil surfaces. The heat is released, on the other hand, mainly through the evaporation at the water surface, but, at the soil surface by the sensible heat flux.

The remainder term shows significant positive values, at 14 : 00 and 15 : 00 and a negative value at 19 : 20, at the river surface. This would be due to the inaccurate estimates for conductive heat flux, including the unrealistic assumption that the heat conduction toward the soil underneath the river was neglected. The amount of the remainder term is also large (negative value) at 17 : 20 and 18 : 00 at the soil surface. Since the temperature at the soil surface was subjected to a disturbance caused by a transient presence of cloud in this period, the estimate of the heat conduction with the temperature data only at two depths including the surface level could be inaccurate.

It can be seen in Fig. 3, however, the amount of heat source is generally in good agreement with the amount of released heat at each time. This would indicate that the roughness length of 10 mm employed in the calculation and hence the calculated rate of evaporation are not far from reality.

## 5. Solute and $^{18}\text{O}$ concentration of the river

Time variation of the river discharge was observed on July 23. The flow flux decreased gradually

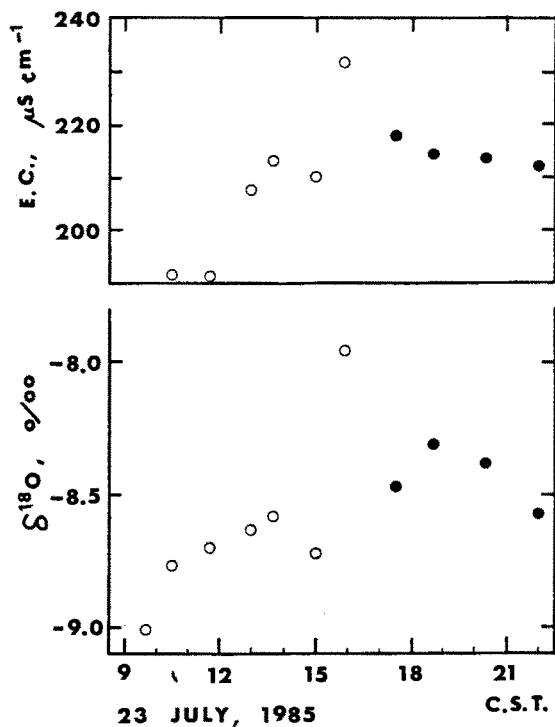


Fig. 4. Electric conductivity (E.C.) and  $\delta^{18}\text{O}$  of the river water.

with time, and the stream was dried up at about 15:00 at an observation site, forming a water tongue upstream (Nakawo and Watanabe, 1987). Another site was chosen at about 100 m upstream from the previous site for further observation. A bottle of water was sampled from the river at every 1–1.5 h. The site for the sampling was where the discharge was measured: two different sites, each for earlier and later halves of the day; the sites were apart one another by about 100 m.

Electric conductivity and  $\delta^{18}\text{O}$  have been measured with the water samples; the results are presented in Fig. 4, in which open and solid circles indicate the earlier and the later sites respectively for the sampling. Both the conductivity and  $\delta^{18}\text{O}$  show a similar trend: a gradual increase in the morning until about 16:00, and a slight decrease toward evening. Since the evaporation of water vapor from the river surface increased with time in the morning (Fig. 2), the increase in the conductivity is considered to be due to the concentration of various salts in the river water caused by the evaporation of water vapor. Isotopic fractionation during vaporization would have caused

the increase in  $\delta^{18}\text{O}$  also by the increase of rate of evaporation. The decrease in the both variables in the evening, hence, can be attributed to the decrease in the rate of evaporation.

When the concentration of various salts in the river water is expressed by  $c$ , the continuity equation is given by

$$\frac{ce}{h} = v \frac{\partial c}{\partial x} + \frac{\partial c}{\partial t} \quad (10)$$

when the salts are well mixed in the river water including the water soaked into the soil underneath. In eq.(10),  $v$  is flow speed of the river,  $e$  is rate of evaporation,  $x$  is distance along the river, and  $t$  is time. It is assumed implicitly that the river width and its height  $h$  (water depth) are independent of time, which is unrealistic for the river under consideration (Nakawo and Watanabe, 1987). In further discussion, however, only a short period, say several minutes, will be examined, and hence the above assumption is considered applicable. Also, the river flow can be regarded as in the steady state for the rather short period. Assuming that the electric conductivity  $K$  is proportional to the solute concentration,

$$\frac{Ke}{h} = v \frac{dK}{dx} \quad (11)$$

At 17:40, three samples were taken over a span of about 100 m. A gradient of the conductivity of about  $-0.3 \mu\text{S cm}^{-1}$  per m was obtained from an analysis of these samples. With this value and the other measured data, using eq.(11), the rate of evaporation  $e$  was calculated to be  $5 \mu\text{m s}^{-1}$ , which is larger than the rate given in Fig. 2 by one order of magnitude. This indicates that the solute concentration increased much faster, as the river water flowed downward, than the rate expected from the process of concentration through evaporation. This would perhaps be caused by solute addition to the river water from the soil, which was neglected in eq.(10), since the river could dry up sometime as expected by the observations by Nakawo and Watanabe (1987).

Similar to eq.(10), the following equation can be given for isotopic composition of oxygen.

$$\frac{ae}{h}(1-\alpha) = v \frac{\partial a}{\partial x} + \frac{\partial a}{\partial t} \quad (12)$$

where  $\alpha$  is fractionation factor for evaporation process, and  $a$  is concentration of  $^{18}\text{O}$  expressed with  $\delta^{18}\text{O}$  value as follows.

$$a = a_s \left( 1 + \frac{\delta^{18}\text{O}}{1000} \right) \quad (13)$$

where  $a_s$  is the concentration of  $^{18}\text{O}$  in standard mean

ocean water (SMOW).

The rate of evaporation  $e$  was calculated, using the data on  $\delta^{18}\text{O}$  for the water samples, for steady condition by eq.(12) with the same method as employed for the conductivity. In the calculation,  $\alpha$  was taken to be 1.0091, the value for equilibrium process at 20°C (Dansgaard, 1964). Obtained was the value of  $1.3 \mu\text{m s}^{-1}$ , which is about three times larger than the observed rate of evaporation. The river water, however, was in contact with the soil for at most several hours, which would not be enough for significant isotope exchange between the water and the soil. It is considered, therefore, that the disagreement is mainly due to an assumption of equilibrium condition, which would not be the case for evaporation in an arid area. Kinetic effect would be involved in the rapid evaporation. Effective fractionation factor was, hence, calculated by substituting  $e = 0.4 \mu\text{m s}^{-1}$  (observed rate of evaporation) and the observed gradient of  $a$ ,  $-7 \times 10^{-9}$  per m, into eq.(12), assuming the condition of steady state. The value of 1.03 was obtained for  $\alpha$ , which is compatible with the values obtained by other workers (e.g. International Atomic Energy Agency, 1981), but much larger than the value for equilibrium process. This should be noted in such a dry region as the West Kunlun Mountains, when analyzing the water cycle in terms of  $\delta^{18}\text{O}$ .

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