Characteristics and climatic sensitivities of runoff from a cold–type glacier on the Tibetan Plateau

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ABSTRACT

Model calculations are made in order to understand the characteristics and response to climate change of runoff from a cold glacier on the Tibetan Plateau. Some 20% of meltwater is preserved at the snow–ice boundary due to refreezing, since the glaciers in mid to northern Tibet are sufficiently cooled during the previous winter.

Sensitivity to alterations in meteorological parameters has revealed that a change in air temperature would cause not only an increase in melting by sensible heat, but also a drastic increase in melting due to lowering of the albedo, since some of the snowfall changes to rainfall. In addition, it was suggested that a decrease in precipitation would cause a lowering of the surface albedo, with a resulting increase in the contribution of glacier runoff to the total runoff of river water. This study shows the first quantitative evaluation of the above effects, though they have been suggested qualitatively. The seasonal sensitivity of glacier runoff was examined by changing the dates given for a meteorological perturbation for a period of only 5 days. It was revealed that changes in both air temperature and precipitation during the melting season strongly affected glacier runoff by changing the surface albedo, though these perturbations only slightly altered the annual averages. **Citation:** Fujita, K., T. Ohta, Y. Ageta (2007), Characteristics and climatic sensitivities of runoff from a cold–type glacier on the Tibetan Plateau, *Hydrol. Process., 21*, 2882-2891, doi:10.1002/hyp.6505.

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INTRODUCTION

Glaciers located around the Asian highlands play an important role in the local water cycle by providing an abundance of meltwater to the adjacent arid/semi–arid regions. The contribution of melt water from glaciers in the west Kunlun Mountains was estimated to account for about half of the river water flowing to the Taklimakan Desert (Ujihashi *et al.*, 1998; Yao *et al.*, 2004). Fluctuations in glacier runoff and the consequent replenishing the water supply will, therefore, strongly affect human life in arid terrain. Although several case studies have been published on the hydrological system and its climatic sensitivities (e.g., Shi and Zhang, 1995; Ding *et al.*, 2000; Kang, 2000; Lan and Kang, 2000), few studies have been done with respect to glacier runoff on the Tibetan Plateau.

In the case of European and American glaciers, on which more studies have been carried out, runoff has been analysed on the basis of water storage and the delay of maximum flow, since temperate glaciers are more widely distributed (e.g., Fountain and Tangborn, 1985; Collins, 1987; Jansson *et al.*, 2003). On the other hand, it is considered that the runoff from Tibetan glaciers seems to be simpler with respect to runoff, because cold–type glaciers are dominant on the Tibetan Plateau (Huang, 1990), and no capacity for temporal storage is expected in cold glacier ice. Refreezing of meltwater, however, must be taken into account, since significant amounts of melt water have been captured as superimposed ice (Fujita *et al.*, 1996). In addition, although there is a dearth of measurement data on the Tibetan Plateau, abundant data exist for Europe and North America. Therefore, since a statistical approach is impossible, a practical glacier runoff model is an appropriate tool to determine how glacier runoff responds to climate change.

In the 1990s, intensive observations were carried out on glaciers, meteorology, permafrost, and river runoff in the Tanggula Mountains in the semi–arid central Tibetan Plateau (33°04'N, 92°04'E; Figure 1a), revealing the current status of the water cycle system including glaciers, soil water, and precipitation (Koike *et al.*, 1994; Ohta *et al.*, 1994; Seko *et al.*, 1994; Ueno *et al.*, 1994; Fujita *et al.*, 1996; 2000; Ageta *et al.*, 1997). Based on these observational results, the characteristics and climatic sensitivities of a glacier mass balance have been analysed using a numerical mass–balance model (Fujita and Ageta, 2000). We apply this model and discuss the characteristics and climatic sensitivities of glacier runoff.

GEOGRAPHICAL AND METEOROLOGICAL SETTING

The watershed of this study includes the Da and Xiao (large and small in Chinese) Dongkemadi Glaciers at their headwaters (Figure 1b). The total area and altitudinal distribution of the watershed are shown in Table I and Figure 2. Runoff measurements were carried out at Base Camp (10 km from the glaciers, BC in Figure 1b) for the 1993 melting period (Ohta *et al.*, 1994). The glaciers range from 5280 to 6104 m a.s.l. The average surface inclination is about 10° facing south, and there are few crevasses with no icefall (Figure 3). Table II and Figure 4 show the meteorological conditions observed at 5600 m a.s.l. on the glaciers during the period from October 1992 to October 1993 (Fujita and Ageta, 2000). The mean annual air temperature at 5600 m a.s.l. is about –10 °C, with an annual range exceeding 20 °C. Daily mean air temperatures exceed 0 °C for only 30 days a year, mainly in August. Most precipitation is supplied by the Indian monsoon during the summer melting season. The average shortwave radiation flux from June to August 1993 is 280 W m^{-2}, which is stronger than that of almost all mid–latitude glaciers (Ohmura *et al.*, 1992). Observational results and features of the glacier mass balance have been described by Seko *et al.* (1994), Ageta and Fujita (1996), and Fujita *et al.* (1996; 2000).

OBSERVATIONAL RESULTS

Runoff from both Dongkemadi Glaciers is estimated from the runoff obtained at BC, and the electrical conductivities of the river water, melt water on the glaciers, and water originating from the soil are given as:

$$\begin{aligned} R_s C_s + R_g C_g &= R_w C_w \\ R_s + R_g &= R_w \end{aligned} \tag{1}$$

where R (m^3 day^{-1}) and C (S m^{-1}) denote the amount of daily discharge and electrical conductivity respectively. Suffixes s, g, and w denote the values of water originating from the soil, glaciers, and the entire watershed respectively. Based on several measurements (8 for glacier water and 18 for soil water), the electrical conductivity of the melt water from the glaciers and soil is assumed to be constant at 20×10^{-4} S m^{-1} and 180×10^{-4} S m^{-1}. The runoff amount and electrical conductivity at BC were measured at 1 h intervals and averaged for daily values. Daily amounts of runoff from the Dongkemadi Glaciers and non–glacierized area are obtained by solving the above simultaneous equation (Figure 5a). In addition, the ratios of runoffs from glacierized and non–glacierized areas to the runoff from the whole area are also shown in Figure 5b. In July, the total runoff amount was suppressed, with almost all of it coming from non–glacierized areas. Since the glaciers are located above 5280 m a.s.l., where no melt had yet started, most of the runoff would be melt water from permafrost at a lower elevation. A rapid increase in runoff due to glacier melting is found, and glacier runoff contributes about half of the total runoff from the end of July to early September. During the observation period (from 1 July to 9 October, except for two days in August), contributions of the amount of runoff from glacierized and non–glacierized areas against the total runoff were 45% and 55% respectively (Table III). In contrast, the runoff depth from glaciers was 1.8 times that from the non–glacierized area (Table III). Although melting of the glaciers had started late due to their high elevation, a drastic melt supplied water to the river for a short period (Figure 5c). Electrical conductivity of the soil water could change spatially and temporally, while that of the glacier would be rather stable. Standard deviation of the measurements of soil water (50×10^{-4} S m^{-1}) could cause ±25% of the glacier runoff. However, temporal variability of the electrical conductivity of soil water may be rather small because thawing thickness of permafrost was less than 1 m in this region (Yabuki *et al.*, 1994), and this shallow depth guarantees that the permafrost does not yield multiyear interacted water. We believe, therefore, that this method is applicable as a preliminary estimation of glacier runoff from observational data.

GLACIER–RUNOFF MODEL

Based on their observations, Fujita *et al.* (1996) pointed out that a significant amount of melt water was refrozen at the interface of snow and ice, since the ice temperature was sufficiently cold (about –8°C at a 16 m depth at 5600 m a.s.l.). Runoff from a glacier, therefore, does not directly correspond to melt water at the glacier surface, whereas those two factors are equal in temperate glaciers which are defined as 0 °C ice temperature all year around. The amount of refrozen water differs with the altitude, since it depends on the melt intensity, thickness of the snow layer, and coldness of the ice. A numerical model should be useful, therefore, in evaluating the effect of melt water refreezing on glacier runoff in a watershed. Fujita and Ageta (2000) have discussed the features of glacier mass balance using a numerical model in which the refreezing process was taken into account. Their model obtains the daily amounts of melt water, refrozen water, and runoff, after solving the issues of surface energy balance and heat conduction in the glacier ice. The basic equations used in the model are described as follows:

$$[M,0] = SR_d(1-\alpha) + LR_d + LR_u + SH + LH + G \tag{2}$$

where M is heat for melting, SR_d is incoming solar radiation, LR_d is downward longwave radiation, SH is sensible heat, LH is latent heat, and G is conduction heat into glacier ice. The downward longwave radiation is calculated using air temperature, relative humidity and the ratio of solar radiation to that at the top of atmosphere (Kondo, 1994). The upward longwave radiation, sensible heat and latent heat are calculated by the bulk method as follows:

$$\begin{aligned} LR_u &= \sigma T_s^4 \\ SH &= c\rho CU(T_a - T_s) \\ LH &= l\rho CU[h_r q(T_a) - q(T_s)] \end{aligned} \tag{3}$$

where σ is the Stefan–Boltzmann constant, T_s is surface temperature, c is specific heat for air, ρ is air density, C is bulk coefficient for sensible and latent heat, U is wind speed, T_a is air temperature, l is latent heat for fusion of ice, h_r is relative humidity, and q is saturated specific humidity. Since all factors in Equation (3) are obtainable if the surface temperature is known, that temperature is estimated as follows:

$$T_s = \frac{SR_d(1-\alpha) + LR_d - \sigma(T_a + 273.2)^4 - l\rho CU(1-h_r)q(T_a) + G}{4\sigma(T_a + 273.2)^3 + \left(\frac{dq}{dT_a} l + c\right)\rho CU} \tag{4}$$

where we assume no heat for melting and the following approximations:

$$\begin{aligned} & T_s \approx T_a \\ & (T_s + 273.2)^4 \cong (T_a + 273.2)^4 + 4(T_a + 273.2)^3 (T_s - T_a) \\ & q(T_s) \cong q(T_a) + \frac{dq}{dT_a}(T_s - T_a) \end{aligned} \tag{5}$$

When the positive surface temperature is calculated in Equation (4), it is set at 0°C. The following iterative calculations are performed until the difference between surface temperatures becomes <0.1°C:

(1) surface temperature is obtained assuming no heat transfer into the glacier;

(2) heat transfer into the glacier is calculated using the calculated surface temperature;

(3) a new surface temperature is obtained using the calculated heat flux into the glacier.

A detailed description of the model was given in Fujita and Ageta (2000).

Input variables for the heat balance calculation are air temperature, incoming solar radiation, relative humidity, and wind speed in daily values. The daily amount of precipitation is also required for the mass balance calculation. Since solar radiation is strong in the region due to its low latitude, the surface albedo can drastically alter the heat balance of the glacier surface. In order to evaluate the effect of changes in meteorological variables on the glacier melt, the surface albedo in the model is calculated from the surface snow density, which changes with compaction, deposition of new snow, and the removal of upper snow by melting. This implies that the albedo is not an input parameter, but changes autonomously with the surface conditions. In addition, refreezing of melt water in snow will also affect the runoff amount. The model calculates the refrozen amount from changes in the ice temperature profile and water content in the snow, and runoff will be produced when excessive water exists in the snow. The amounts of refrozen and runoff water differ at each elevation depending on temperature of glacier ice and the amount of percolation water (Fujita *et al.*, 1996). The model provides plausible results, such as changes in surface and ice temperatures, relative levels of the surface and snow–ice interface, surface albedo, and the altitudinal profile of mass balance, all of which have been verified by observational data (Fujita and Ageta, 2000). In particular, demonstrations of changes in the albedo and levels of the surface and snow–ice interface (Figure 6) imply that the albedo, surface heat balance, and melt water refreezing are reliably calculated in the model.

RUNOFF CHARACTERISTICS

Figure 7 shows that the glacier runoff is estimated from observations at BC (R_g in Equation (1), referred to as 'estimated runoff' hereafter), and the glacier runoff calculated by the model ('calculated runoff' hereafter) for the period from June to October of 1993. The calculated runoff shows a 1 day delay compared with the estimated runoff. Distance does not provide a plausible reason for this delay, since BC is located only 10 km from the glaciers. In the case of temperate glaciers, a sizeable amount of melt water will be retained within the glacier body, and thus temporal changes in the glacier runoff will differ significantly from those of the glacier surface melt. It is well known that the water storage in temperate glaciers will cause a delay in glacier runoff of several days to a few months (Fountain and Tangborn, 1985). However, it is not considered that melt water will be retained within the cold glacier ice, since the ice temperature is sufficiently cold. Since surplus non–refrozen water is immediately removed as runoff in the model, the delay seems to be caused by a failure to account for water infiltration in the snow layer. In any event, the changes in glacier runoff are reliably demonstrated in the model.

Changes in runoff based on a non–refrozen assumption (in which all melt water is immediately removed as runoff) are also calculated, as shown in Figure 7. Since melt water does not infiltrate into the glacier ice, the superimposition rate of refrozen water depends on how well the latent heat released with refreezing can be absorbed by the cold glacier ice body (Fujita *et al.*, 1996). Therefore, runoff through the refreezing process is considerably less than that through the non–refreezing process in the early melting season when enough cold ice refreezes a small amount of melt water. In addition, the difference also increases just after a short cooling. In the following August, however, the difference diminishes, since a significant amount of melt water reaches the warmed–up glacier ice. The runoff through the refreezing process was 20% less than that through the non–refrozen process during the main melting season in July and August (Table IV). This implies that the refreezing process must be considered in the runoff model for a cold–type glacier, whereas such capturing of water has not been taken into account in several runoff models for Himalayan glaciers (e.g., Fukushima *et al.*, 1991; Braun *et al.*, 1993). Since many cold–type glaciers are located around the Taklimakan Desert (Huang, 1990), where the glacier runoff contributes significantly to the river water (Ujihashi *et al.*, 1998), the refreezing process should be particularly taken into account in the runoff from cold–type glaciers.

SENSITIVITY TEST

Sensitivity to climatic changes in variables

In order to evaluate how glacier runoff is affected by the changes in climatic variables, the anomaly of each variable yielding a 10% increase in glacier runoff is calculated (Table II). Only one variable was changed without changing the other parameters. Although it is difficult to compare variables having different units and showing different fluctuations, it is notable that changes in air temperature affect the glacier runoff more sensitively than the other variables. In a normal melting season, a significant amount of precipitation due to the monsoon falls as snow, thus covering the glacier surface several times with high–albedo snow. Therefore, this suggests that the precipitation in summer prevents excessive melting and a loss of the glacier mass (Fujita and Ageta, 2000). In contrast, since the rain–snow boundary line fluctuates around the altitude of the glacier during the melting season (explaining why glaciers can exist there), changes in air temperature will determine whether precipitation falls as low–albedo rain or high–albedo snow on the glacier surface. If no snow covers the surface, the melt amount will increase drastically due to the absorption of strong solar radiation. In order to confirm the effect of the albedo, three runs were calculated for the altitude at 5600 m a.s.l. (Table V). Case 1 is the result of a control run. Case 2 is a warming test (+1 ºC) with the same albedo given in case 1. Case 3 is the same warming test (+1 ºC) but with the albedo calculated according to the model scheme. Case 2 provides the effect of air temperature warming only, which could cause a +174 mm w.e. increase in meltwater (40% over case 1). However, temperature warming might cause not only an increase in melting by sensible heat, but also an albedo decrease, as shown in Figure 8. Melt water would increase drastically (107% over case 1), since ice with a low albedo would appear during the melting season, whereas under the conditions in case 1 the surface was covered with snow.

Another interesting feature is that an increase in glacier runoff would result from a decrease in precipitation (which is unusual in river runoff), though considerable changes must occur in the amount of precipitation (corresponding to 24% of total glacier runoff and 16% of annual precipitation). At the high elevations where glaciers exist, a decrease in precipitation implies a decrease in snowfall with a high albedo. Hence, the surface albedo will decrease when snow does not cover the glacier surface, and the snow/ice melt will be accelerated even under the same temperature conditions (Fujita and Ageta, 2000). With respect to regional river runoff, therefore, the contribution of glacier runoff will increase/decrease when precipitation decreases/increases, whereas the precipitation falling on non–glacierized (permafrost) areas will emerge as runoff water with some delay. Although this phenomenon has been suggested qualitatively based on a statistical analysis of runoff from glacierized catchments (Collins, 1987), ours is the first quantitative evaluation showing the effect of precipitation on glacier runoff.

Seasonal sensitivity

In the above section, meteorological variables are changed homogeneously through the year in the model calculation. This is not plausible, however, since these variables will not change homogeneously except for a short period. 'Seasonal sensitivity' of the glacier runoff, therefore, was examined by changing the time when a variable changed for a short period. Daily means of air temperature (+1 °C) and daily precipitation (+10 mm w.e.) were changed from the input data for only a 5–day period. Figure 9 shows the seasonal sensitivity of glacier runoff on the change in each variable during those 5 days. The abscissa and ordinate are the dates when a perturbation occurred and the calculated total runoff (a) and summer mean albedo (b) respectively. Summer mean albedo is obtained by averaging the surface albedo with weighted–area distribution for the period from June to August. The calculated total runoff (61.3×10^5 m^3) and summer mean albedo (0.762) with no perturbation were also shown in the figures. Changes in air temperature in winter affect the glacier runoff hardly at all, whereas changes during the melting season increase it by nearly 10%. A warming of +1 °C during the 5 days corresponds to the +0.014 °C warming of annual mean air temperature, whereas the warming needs +0.1 °C in the case of a homogeneous warming as mentioned above (Table II). An increase in precipitation will bring about a decrease in glacier runoff through a year. Precipitation in winter will fall as snow on the whole glacier surface and thus delay the timing at which the ice surface with a low albedo appears in the following melting season. Precipitation in the early melting season (May to June) is most effective in decreasing glacier runoff, since in those months it is usually less than that in the highest melting season (July to August). Thus, a high albedo snow cover will effectively prevent surface melting under conditions of strong solar radiation. In contrast, some amount of precipitation in July and August will fall as rain due to the high air temperature, making it less effective in preventing melting. Since melting is almost over by September, an addition of snow will affect runoff the following year. Although changes in summer mean albedo may seem small (since they are obtained by averaging for the whole glacier area), they will significantly affect glacier runoff. The strong negative correlation between albedo and runoff depth (r = –0.98 with a 99% significance level) would also suggest that these perturbations affect runoff through changing the surface albedo of the glacier, as shown in Figure 10. These findings imply that the glacier surface conditions altered by a perturbation during only 5 days will greatly affect the heat/mass balance of the glacier and glacier runoff for the next melting period.

CONCLUSION

The model calculations revealed that melt water refreezing could not be a negligible factor in the glacier runoff from cold–type glaciers on the Tibetan Plateau. Although in temperate glaciers the amount of glacier runoff was considered to be equal to that of melt water at the glacier surface, refreezing at the snow–ice interface of the glacier captured 20% of the melt water generated at the surface. This result suggests that it is not suitable to describe runoff water as equivalent to melt water, although this is a common mistake made in previous studies.

Model calculations were conducted for climatic and seasonal sensitivities. A warming of the air temperature most effectively increased the glacier runoff not only by increasing the sensible heat flux, but also by changing the phase of precipitation from snow to rain, which directly affects the albedo of the glacier surface. A decrease in precipitation, in contrast, increased the glacier runoff by reducing the chance of snow cover with a high albedo, which should prevent melting at the glacier surface. Since solar radiation is the main heat source of heat balance on the glacier surface, the surface albedo of the glacier is the most significant variable for glacier runoff.

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Table I. Area of watershed at BC (Figure 1b) and the non–glacierized and glacierized areas.

Watershed	Area (km^2)
Whole at BC	50.5
Non–glacierized	34.6
Glacierized	15.9

Table II. Averages and summation of meteorological variables measured at 5600 m a.s.l. on Xiao Donkemadi Glacier from 10 October 1992 to 9 October 1993 (Fujita and Ageta, 2000). Column 3 shows anomalies of climatic variables (climatic sensitivities) resulting in a 10% increase in runoff from the Dongkemadi Glaciers.

Meteorological variables	Average / summation	Anomaly to yield +10% runoff
Air temperature (°C)	–10.3	+0.1
Precipitation (mm w.e.)	672	–108
Global solar radiation (W m^{-2})	240	+20
Relative humidity (%)	77.9	+2.7
Wind speed (m s^{-1})	4.1	–2.6

[a] w.e.: water equivalent.

Table III. Runoff amount and depth for each watershed from 1 July to 9 October of 1993 (except 2 and 3 August).

Watershed	Runoff amount (10^5 m^3)	Runoff depth (mm w.e.)
Whole at BC	156.9	311
Non–glacierized	86.5	250
Glacierized	70.4	443

Table IV. Runoff amounts from Dongkemadi Glaciers estimated from observations, calculated taking account of refreezing and non–refreezing processes, between July and August of 1993 (except two days in August).

	Runoff amount (10^5 m^3)
Estimated from observation	59.5
Refreezing process	58.4
Non–refreezing process	69.0

Table V. The calculated mass balances at 5600 m a.s.l. for the period from 10 October 1992 to 9 October 1993. Case 1 denotes mass balance in the case of control calculations. Cases 2 and 3 denote mass balances when air temperature is warmed by +1 °C from input data with the same albedo as in case 1 (case 2) and with the albedo calculated according to the model (case 3). Units of all variables are mm w.e. Differences from case 1 are also shown.

	Calculated results			Difference from case 1	
	Case 1	Case 2	Case 3	Case 2	Case 3
Snow	634	592	592	–42	–42
Rain	38	80	80	+42	+42
Balance	220	32	–265	–188	–485
Melt water	440	614	910	+174	+470
Runoff	385	578	868	+193	+483
Evaporation	67	63	70	–4	+3
Refrozen water	93	116	122	+23	+29

Figure legends

Figure 1. (a) Location of Tanggula Mountains and (b) watershed of the Da and Xiao Dongkemadi Glaciers (DD and XD) on the central Tibetan Plateau. Broken line, hatched area, and BC in (b) respectively denote the watershed, glacier area, and Base Camp where river runoff was measured.

Figure 2. Distribution area of the watershed at altitude intervals of 100 m. Black and gray denote non–glacierized and glacierized areas respectively.

Figure 3. Photograph of the Da (left) and Xiao (right) Dongkemadi Glaciers.

Figure 4. Daily means of air temperature (line in a), global solar radiation (solid line in b), wind speed (dotted line in c) and relative humidity (solid line in c) measured at 5600 m a.s.l. of the Xiao Dongkemadi Glacier from 10 October 1992 to 9 October 1993 (after Fujita and Ageta (2000)). Daily amount of precipitation (bars in a) was obtained using a tipping bucket beside the glacier during summer, and estimated using an automatic snow–level gauge during winter. Global solar radiation calculated for the top of atmosphere also shown as a broken line (b).

Figure 5. Daily amount of runoff (a), contribution ratio (b) and runoff depth (c) of watershed of the Dongkemadi Glaciers for summer of 1993. Thin solid lines, thick solid lines, and broken lines denote the whole watershed, glaciers and non–glacierized areas respectively. Text describes separation of glacier and non–glacierized runoffs from runoff observed for the whole watershed.

Figure 6. Relative levels of (a) surface and snow–ice interface and (b) albedo at 5600 m a.s.l. of Xiao Dongkemadi Glacier from October 1992 to October 1993 (after Fujita and Ageta (2000)). Gray and black circles in (a) denote observed surface and snow–ice interface respectively. Solid black and gray lines in (a) denote the calculated surface and snow–ice interface respectively. Gray and broken lines in (b) denote calculated and observed albedo respectively.

Figure 7. Runoffs from the Dongkemadi Glaciers estimated from observations at BC (gray line), calculated from the model (solid line), and calculated with a non–refreezing assumption (broken line) from June to October of 1993.

Figure 8. Temporal changes in albedo calculated for the altitude at 5600 m a.s.l.

Gray and black lines denote the control (Cases 1 and 2) and warming (Case 3) calculations respectively.

Figure 9. Seasonal sensitivity of (a) the glacier runoff and (b) the summer mean albedo examined by changing time at which a variable changed for 5 days. Abscissa and ordinates are respectively the dates when a perturbation was given and the calculated total runoff (a) and summer mean albedo (b). Daily means of air temperature (+1 °C, black line) and daily precipitation (+10 mm w.e., broken line) were altered for 5 days from the input data. The total runoff (61.3×10^5 m^3) and albedo (0.762) of control calculation are depicted by gray line.

Figure 10. Runoff depth versus summer mean albedo. Black and gray dots result from air temperature and precipitation perturbation shown in Figure 9. Regression line is obtained for both results ($R^2 = 0.96$).

Figure 1. Fujita et al.

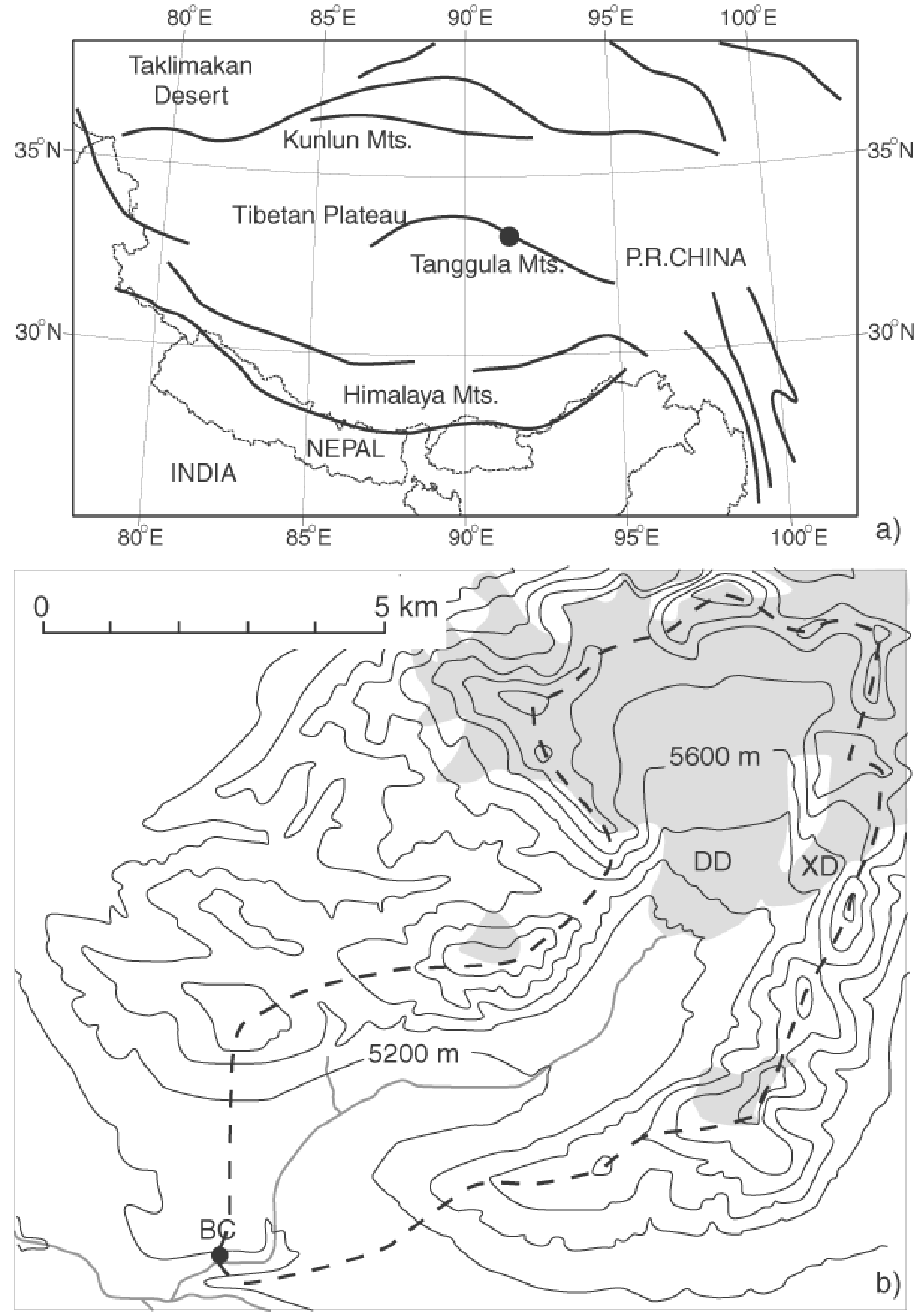

Figure 2. Fujita et al.

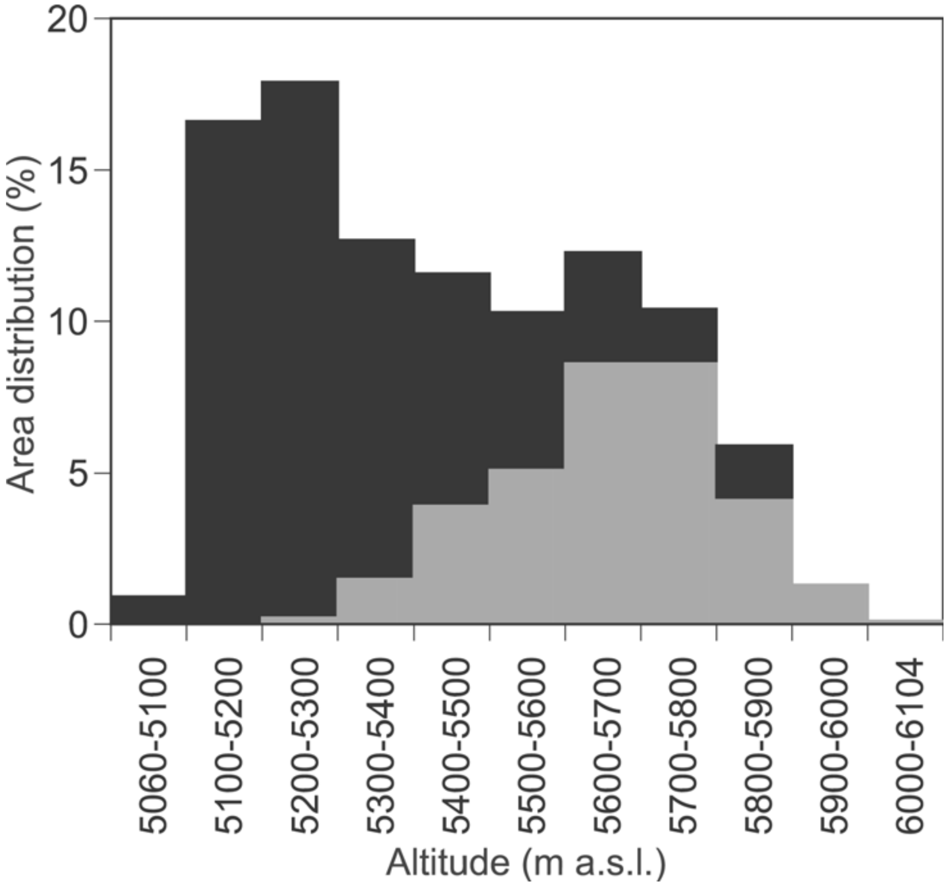

Figure 3. Fujita et al.

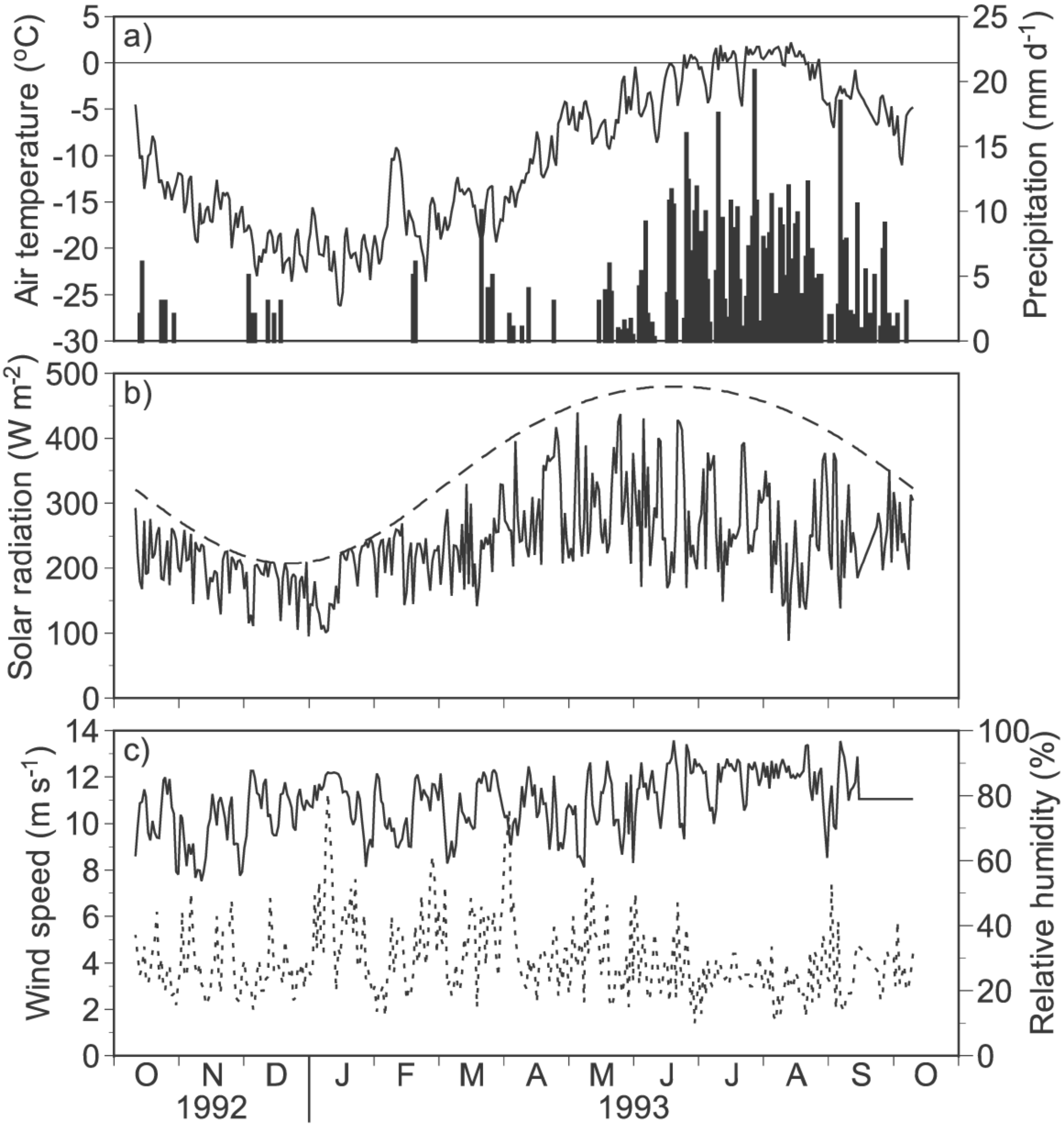
a)
b)
c)
Air temperature (°C)
Precipitation (mm d-1)
Solar radiation (W m-2)
Wind speed (m s-1)
Relative humidity (%)
5
0
-5
-10
-15
-20
-25
-30
25
20
15
10
5
0
500
400
300
200
100
0
14
12
10
8
6
4
2
0
100
80
60
40
20
0
O N D J F M A M J J A S O
1992
1993

Figure 5. Fujita et al.

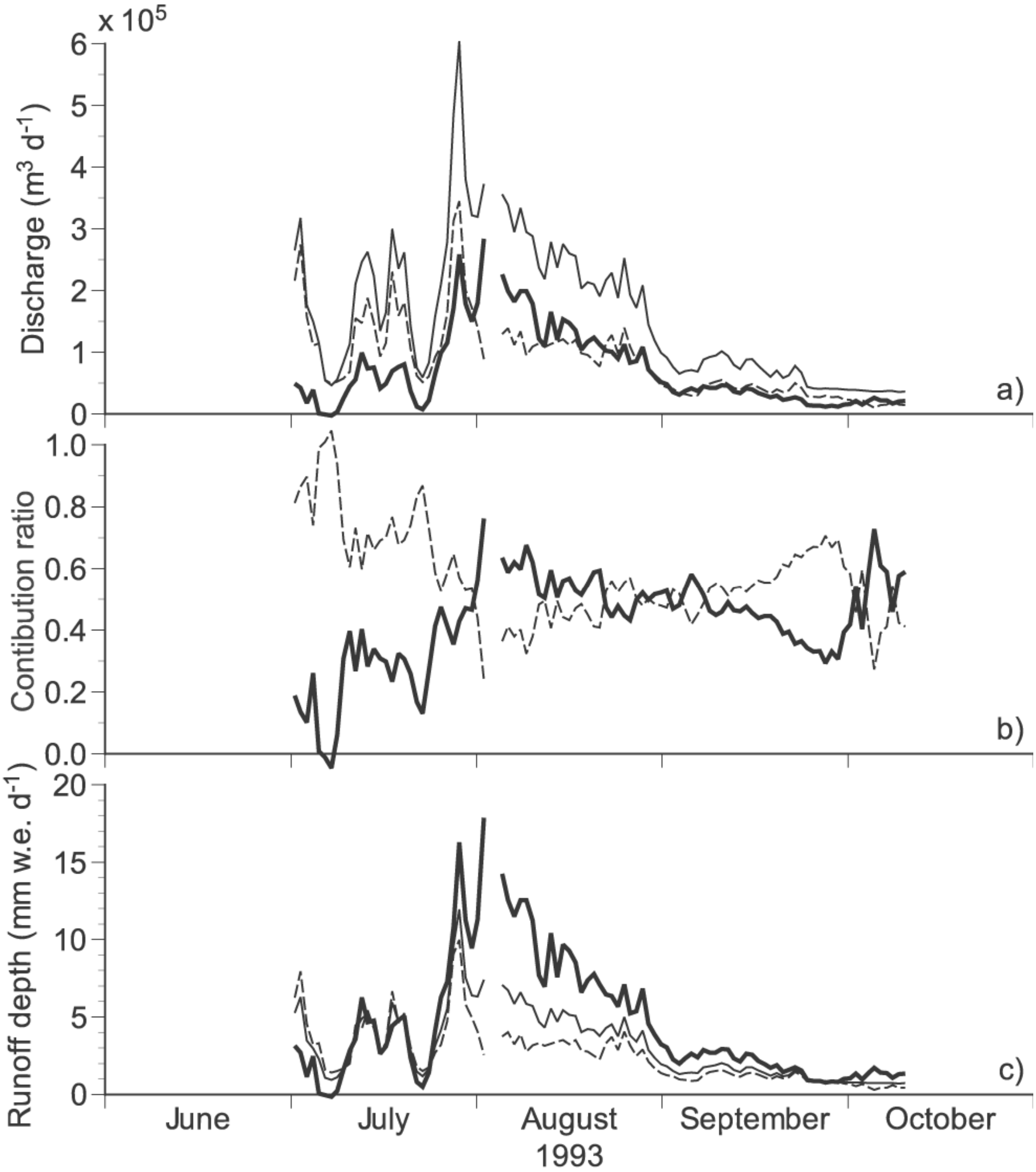

Figure 6. Fujita et al.

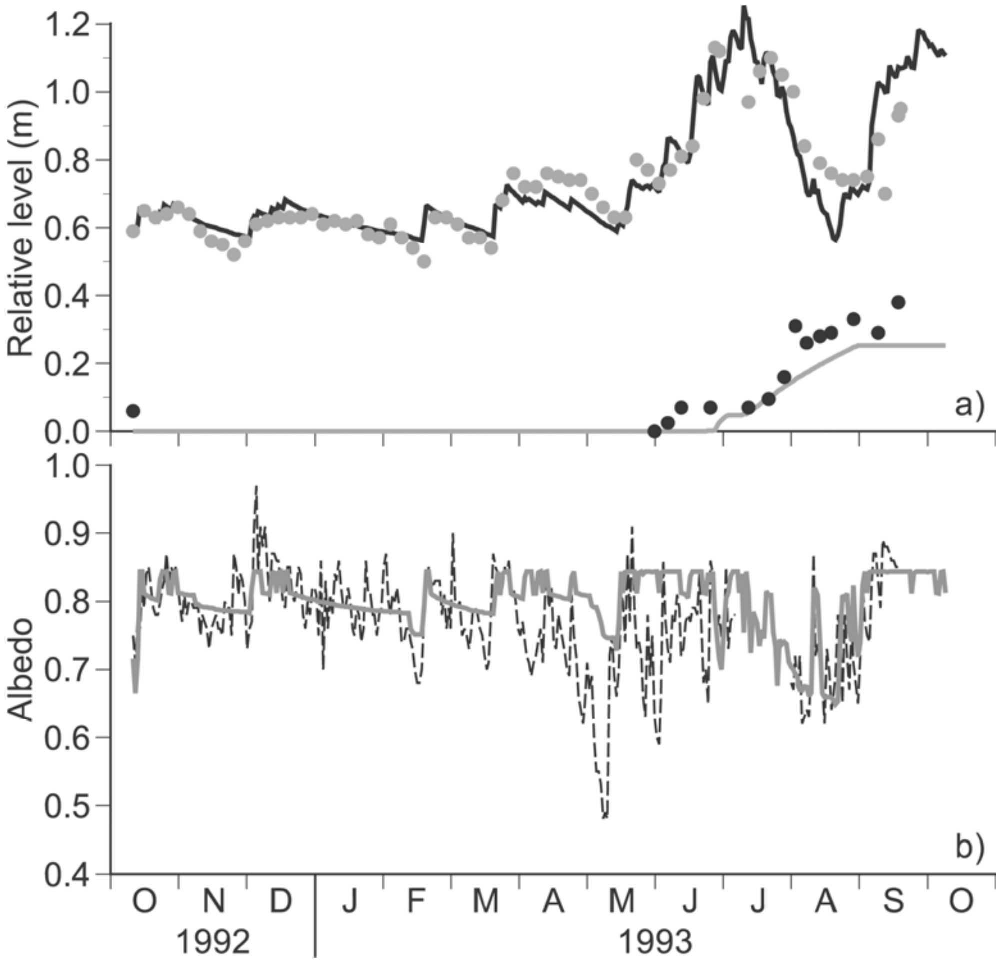

Figure 7. Fujita et al.

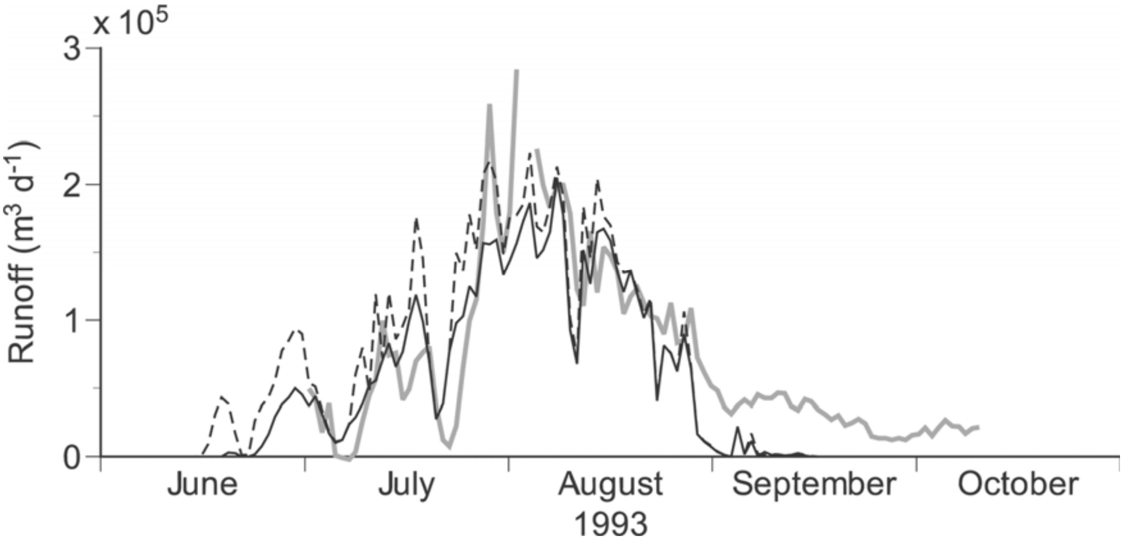

Figure 8. Fujita et al.

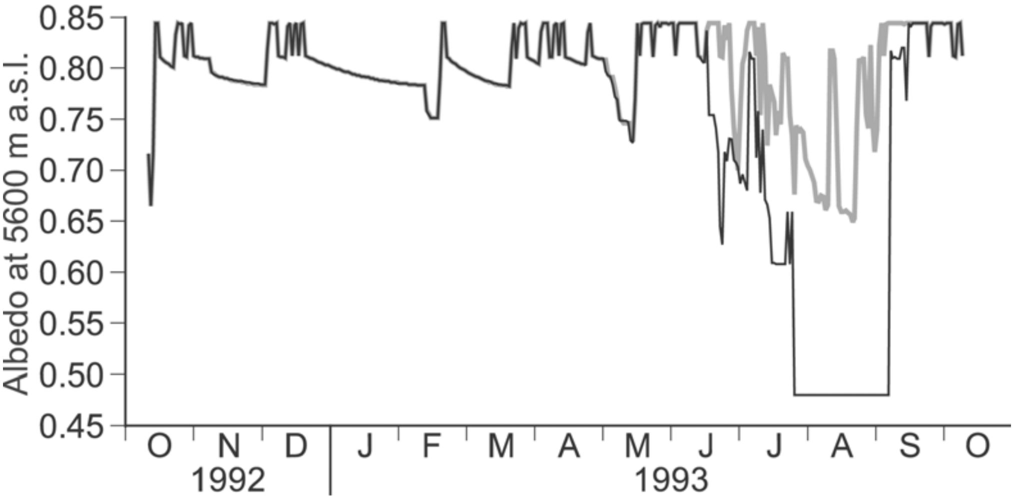

Figure 9. Fujita et al.

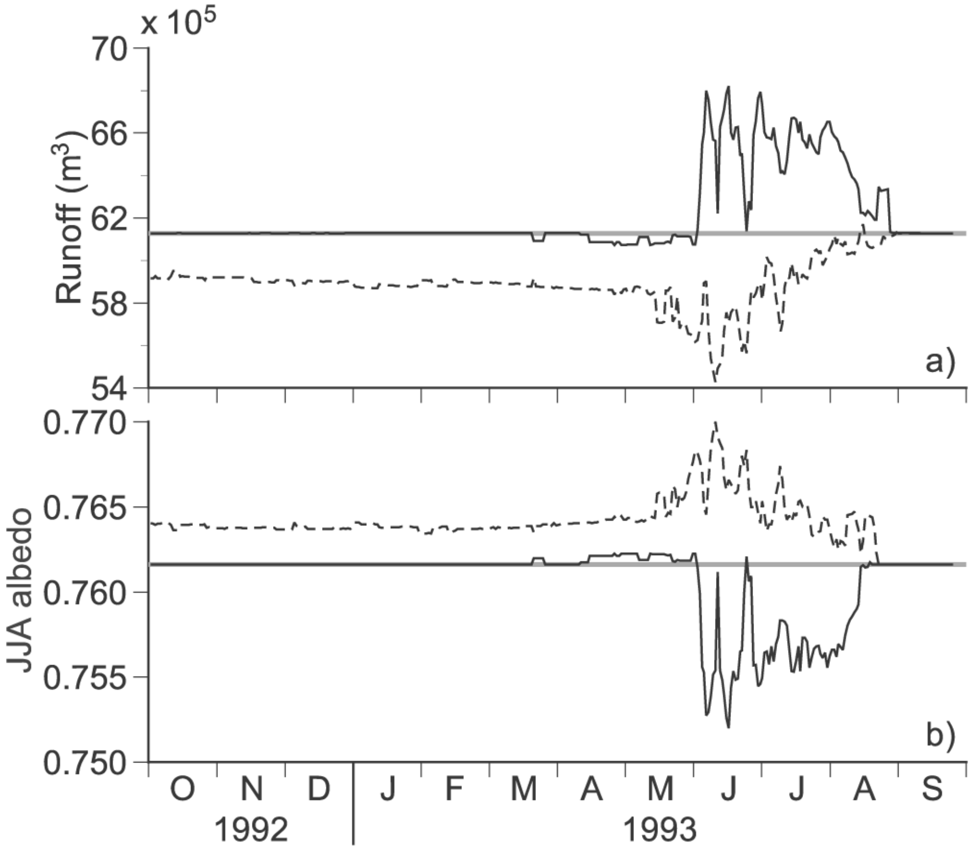

Figure 10. Fujita et al.