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Precipitation, melt and runoff in the northern Tien Shan

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Abstract

Precipitation and snow distribution, melt of glaciers and snow, and runoff formation in two basins in the northern Tien Shan are calculated and described. Major climatic features are a spring–summer precipitation maximum occurring simultaneously with ice and snow melt. Precipitation increases with altitude except in winter when an air temperature inversion occurs. Air temperature is a good predictor of glacial melt. Typical hydrographs have two floods: one is formed from melt of seasonal snow cover, and the other is formed from melt of glacial ice. The second flood is usually larger than the first. In mountain basins of the northern Tien Shan direct runoff from rainfall averages about 7–12% of annual volume. The glacial runoff is 18–28% of average annual runoff in basins with area of glaciation not less than 30–40%, but during summer it can increase to 40–70% of average annual runoff. Surface runoff from seasonal snow melt during spring and summer is 18% of average annual runoff; the groundwater component is 34–38% of average annual runoff.

1. Introduction

Among the problems that limit understanding of snow and glacier melt and runoff are shortcomings in our understanding of how natural processes vary depending on geographic location. Snow-covered and glaciated mountains occur from tropical to arctic regions and near coasts and within continents. Major continental, temperate ranges include the Tien Shan, Urals, Altay, Pamiro-Alay of Asia and Rockies of North America.

Our objective here is to evaluate the importance of the physical processes that control runoff formation in high-elevation mountain catchments in the northern Tien Shan. We consider two basins with long-term observations (Fig. 1). One is located in the Ala Archa River basin on the northern slope of the Kirgizskiy Alatoo, and the other is in the Bolshaya and Malaya Almaatinka river basins on the northern slope of the Zailiyskiy Alatau. Both

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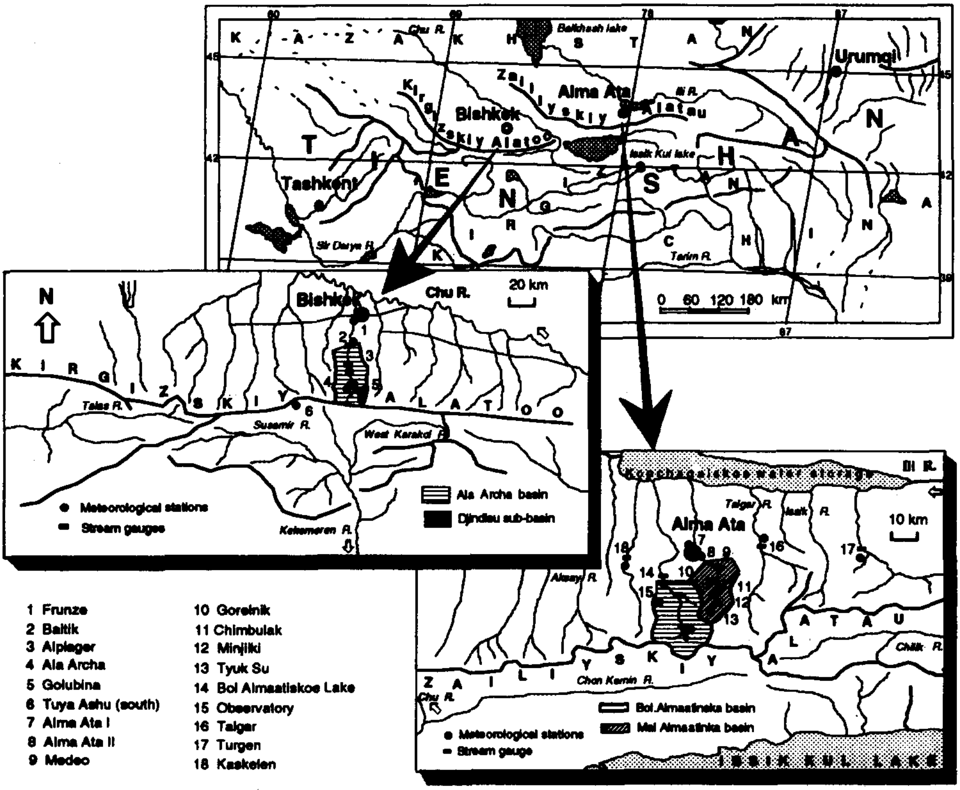


Fig. 1. The Kirgizskiy Alatoo and Zailiyskiy Alatau.

have glacierized and non-glacierized sub-basins. First we provide background information on precipitation, solar radiation and temperature, and then we examine sequentially larger hydrological units. Extending the calculations of snow and glacier melt and runoff from a small sub-basin to larger basins and to the entire range will help understanding of snow and glacier melt and runoff in continental mountain systems.

2. Geographical description

The Kirgizskiy Alatoo and Zailiyskiy Alatau are northern peripheral ranges in the Tien Shan (Fig. 1). In winter, this area is under the strong influence of the Siberian anticyclonic circulation, which decreases precipitation. The Siberian anticyclone weakens after March, and western cyclones bring precipitation that is favourable for snow accumulation and formation of glaciers in the alpine belt.

On the northern slopes of the Kirgizskiy Alatoo and Zailiskiy Alatau glacial and fluvioglacial relief occurs. Quaternary glacial formations such as troughs, cirques, moraines and roches moutonnées are observed above 3000 m. Permafrost is widerspread above

3000 m, with local permafrost beginning at 2000 m. Fluvioglacial relief is represented by moraine ridges and hills, outwash plains, debris-flow alluvial fans and inter-mountain basins.

Soil and vegetation occur except in the nival glacial belt, which occupies 75% of the northern slopes of the Kirgizskiy Alatoo and Zailiyskiy Alatau. Between 2800 and 3200 m there are alpine meadows, and between 1500 and 2800 m are sub-alpine meadows and forests. Below 1500–1600 m are mixed forests; steppe begins at 800–900 m.

The Kirgizskiy Alatau stretches over 400 km. Glaciers on the northern slope are located in the central part of the range, where orography and elevation are more favourable for snow accumulation and formation of glaciers. The total area of glaciation on the northern slope of the Kirgizskiy Alatoo is 471 km^2 (Anonymous, 1978). The highest summit (4895 m) in the Kirgizskiy Alatoo is located in the Ala Archa basin. The area of this basin is 233 km^2.

The length of the Zailiyskiy Alatau is 280 km. Glaciation is well developed in the central Zailiyskiy Alatau and Kirgizskiy Alatoo. In the Talgar massif are located the highest summits of the range; Talgar peak is about 5000 m. The areas of the Malaya and Bolshaya Almaatinka River basins (Fig. 1) are 118 km^2 and 155 km^2 with glaciation areas of 11.4 km^2 and 19.5 km^2, respectively.

3. Data collection

Major hydrometeorological stations in the Kirgizskiy Alatoo and Zailiyskiy Alatau used in our calculations are shown in Fig. 1. The period of observation for most meteorological stations and hydrological sites is not less than 30 years. The network of precipitation, stream gauges and snow cover measurements is presented in Table 1. Topographic maps of 1:25 000 scale were used.

Data were obtained also from year-round measurements at the Golubina glacier (3440 m), Ala Archa River basin, from 1958 to 1991. Long-term precipitation measurements at 44 locations in all landscape zones of the Ala Archa River basin and on 11 transects were obtained in glacial and non-glacial zones from 1600 m to 4200 m. Errors of precipitation measurements associated with evaporation are 2–8% and with wind influences are 4–10% of average values (Sosedov and Filatova, 1961; Cicenko, 1966; Sudakov and Tokmagambetov, 1968; Grigoriev, 1973). Year-round meteorological and glaciological measurements at the Tuyksu glacier in the Malaya Almaatinka River basin on the northern slope of the Zailiyskiy Alatau from the 1962 to 1982 were used (Avsuyk, 1984; Tokmagambetov and Erasov, 1985; Bochin and Krenke, 1987) in calculations of snow and ice melt and runoff. At all meteorological stations, measurements of solid, mixed and liquid precipitation amount were made.

At Golubina and Tyuksu stations ablation was measured on slightly inclined ablation plots (2 m × 2 m) and with ablation stakes. On the plots, measurements were made in the morning and evening before and after sunrise at 121 points located in 20 cm cells. Simultaneously, snow density measurements were made in a snow pit located near the ablation plots. Density of melting ice was assumed to be 0.89 g cm^{-3} (Shumskiy, 1978). Measurements of areal ablation, and of spring and autumn refrozen melt water, were made at more

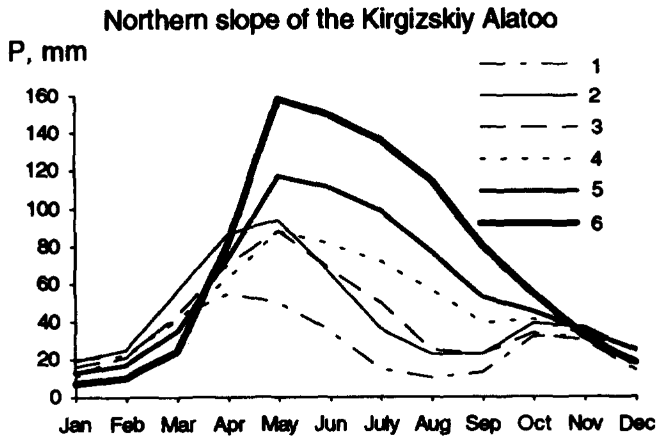
Table 1
Snow survey, stream gauges and precipitation sites on northern slopes of Kirgizskiy Alatoo and Zailiyskiy Alatau

River basins	Snow survey points		Remote points		Precipitation sites		Stream gauges	
	n	m a.s.l.	*n*	m a.s.l.	*n*	m a.s.l.	*n*	m a.s.l.
Central and eastern part of the north slope of Kirgizskiy Alatoo	30	1450–3120			25	2217–4120	70	1200–3300
West part of the North slope of Kirgizskiy Alatoo	21	1200–3300			19	2100–3200	32	1200–3000
Ala Archa and Alamedin basins, the north slope of Kirgizskiy Alatoo	13	1600–4200	10	1880–3530	44	1500–4400	26	1500–3300
Malaya, Bolshaya, Almaatinka and Talgar basins, the north slope of Zailiyskiy Alatau	14	1340–4196	50	2000–3640	24	1400-4196	28	1200–3200
Issik, Turgen, Kaskelen basins, the north slope of Zailiyskiy Alatau	15	1280–2680	33	2000–3850	14	1300–3400	14	1100–3100

n is number of points

than 200 stakes on both the Golubina and Tyuksu glaciers. Accuracy of ablation measurements and thickness of refrozen melt water was 5 mm.

Using hourly measurements of the gradient in air temperature, humidity and wind speed at 0.25, 0.5, 1.0 and 2.0 m, turbulent heat and humidity fluxes were calculated. All calculations were carried out using the theory of surface boundary layer in a homogeneous fluid developed by Monin and Obukhov (1954) and Kazanskiy (1965). Evaporation was measured also by repeated weighing of decimetre cubes of snow and firn using an electric balance with accuracy of 0.1 g. Our estimates of the proportion of ground water to total runoff have been described by Aizen et al. (1995a).



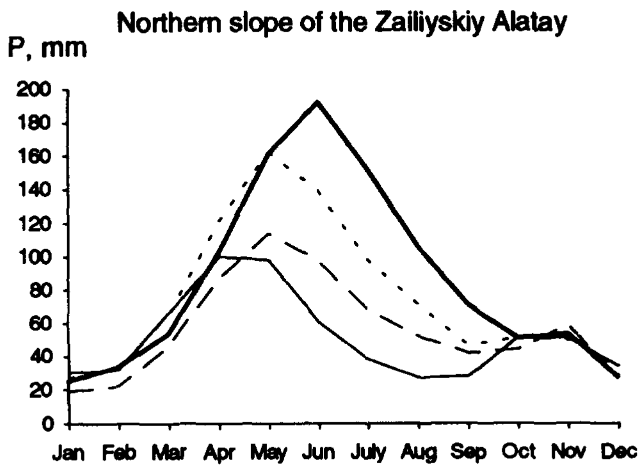
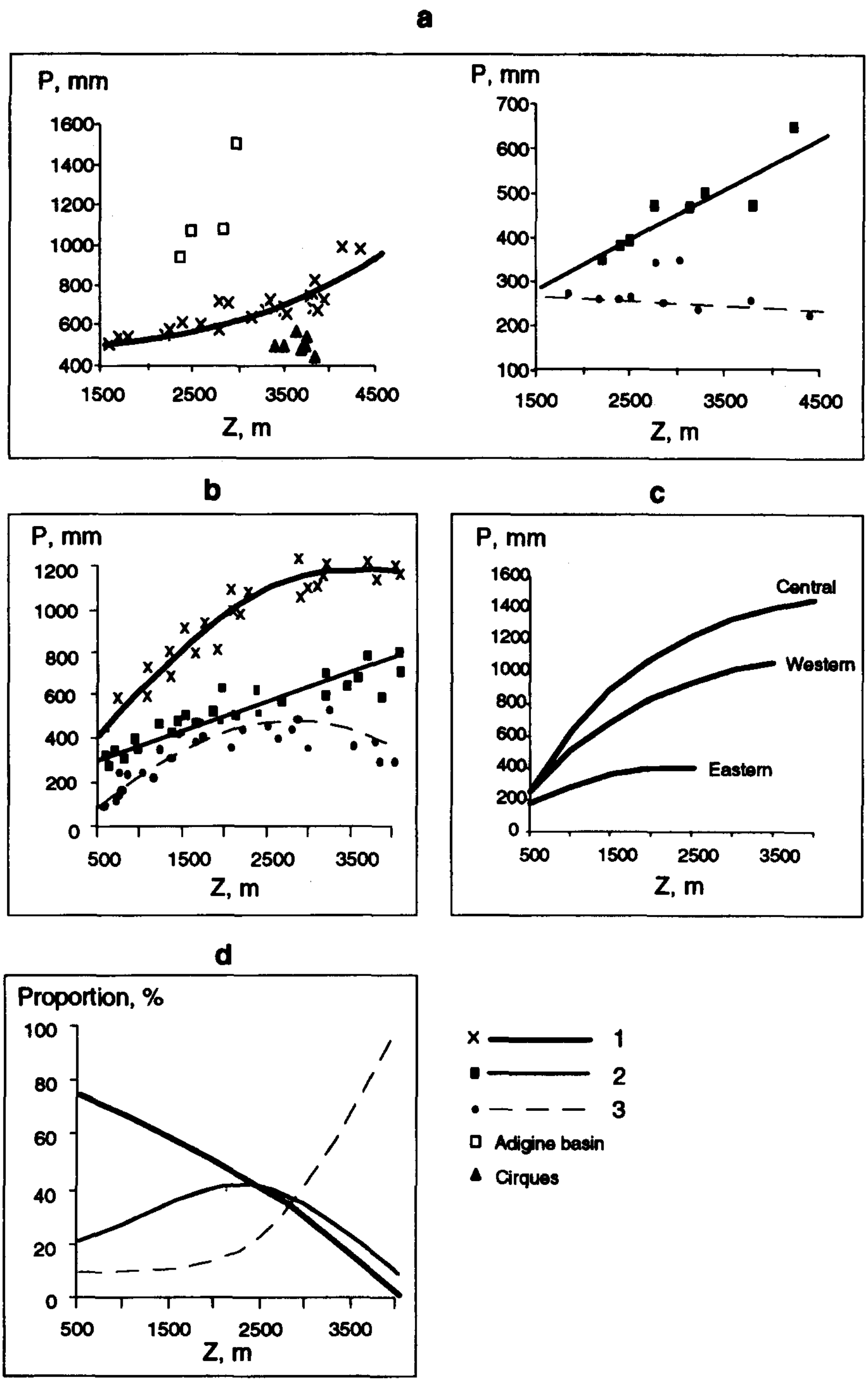


Fig. 2. Annual variation of long-term mean precipitation (P) as a function of altitude on the northern slope of the Kirgizskiy Alatoo and Zailiyskiy Alatay: 1, less than 1000 m; 2, 1001–1500 m; 3, 1510–2000 m; 4, 2001–3000 m; 5, 3001–3500 m; 6, 3501–4500 m.



a
P, mm
1600
1400
1200
1000
800
600
400
1500
2500
3500
4500
Z, m
P, mm
700
600
500
400
300
200
100
1500
2500
3500
4500
Z, m
b
P, mm
1200
1000
800
600
400
200
0
500
1500
2500
3500
Z, m
c
P, mm
1600
1400
1200
1000
800
600
400
200
0
Central
Western
Eastern
500
1500
2500
3500
Z, m
d
Proportion, %
100
80
60
40
20
0
500
1500
2500
3500
Z, m
1
2
3
Adigine basin
Cirques

4. Regime of precipitation

Open northern slopes of the Kirgizskiy Alatoo and Zailiyskiy Alatau are under weak cyclonic activity and moderate influence of cold air from the west, north-west and north. At elevations up to 2500 m, two maxima of precipitation occur (Fig. 2). The main maximum is observed in March–May (35–45% of annual total), and the second is observed in autumn. At high altitude, one spring–summer maximum of precipitation occurs, in May–July (45–55% of annual total). The precipitation minimum in the glacial zone is observed in December and January (2–5% of total).

Based on the long-term data in the Ala Archa and Bolshaya and Malaya Almaatinka River basins, we determined the distribution of total, summer (from May to September) and winter (from October to April) precipitation as a function of altitude (Fig. 3(a)–(c)). On average, yearly precipitation increases with altitude along the main valleys, but there are small-scale variations in distribution within the basins. Forests on north and north-west aspects and windward slopes have increasing moisture, where local annual precipitation reaches 1500 mm as a result of adiabatic processes occurring at the level of interaction between warm air rising from the valleys and katabatic cold winds from glaciers (Grigoriev, 1973). In cirques on leeward slopes, the quantity of precipitation is only about 400–500 mm. If these deviations are excluded (Fig. 3(a)), the main factor determining distribution of precipitation on the northern slopes of the Kirgizskiy Alatoo and Zailiyskiy Alatau is elevation (Fig. 3(a) and (b)). Altitudinal distribution of total precipitation can be approximated by a parabolic function (Fig. 3(a)–(c)):

$$P(Z_i) = P(Z_0)[a_1 + a_2(Z_i - Z_0) + a_3(Z_i - Z_0)^2] \tag{1}$$

where $P(Z_0)$ and $P(Z_i)$ are precipitation at altitude Z_0 and Z_i, and a_1, a_2 and a_3, are the parameters determined by observational data using the least-squares method (Aizen, 1988). Based on measurements during 10 years, the root-mean square errors of calculated precipitation were on average 12% of average in the Ala Archa and in the Bolshaya Almaatinka. On the Kirgizskiy Alatoo, more rapid increase of precipitation occurred in the upper elevations (Fig. 3(a)), because high branches of the range block the passage of westerly moisture-containing air masses (Fig. 1) into the lower and middle part of the basin. On the Zailiyskiy Alatau, larger precipitation gradients occurred at elevations below 1800 m than at upper elevations (Fig. 3(b)). This precipitation distribution is opposite to the distribution in the northern slope of the Kirgizskiy Alatoo, because valleys located on the northern slope of the Zailiyskiy Alatau are more open to air flows from the west and north than are basins on the Kirgizskiy Alatoo.

Along the northern slope of the Kirgizskiy Alatoo the largest precipitation is observed in the central and eastern part of the range; on the Zailiyskiy Alatau, the greatest precipitation is observed in the central part (especially in the basin of the Malaya Almaatinka River). On

Fig. 3. Long-term mean total (1) and seasonal (2, May–September; 3, October–April) distribution of precipitation in the Ala Archa basin (northern slope of the Kirgizskiy Alatoo) (a), in the Bol. Almaatinka basin (northern slope of Zailiyskiy Alatau) (b), and on the northern slope of the Zailiyskiy Alatau (c); altitudinal changes in proportion of liquid (1), mixed (2) and solid (3) precipitation in the northern Tien Shan (d).

the Zailiyskiy Alatau precipitation decreases to the west and to the east but altitudinal distribution is similar (Fig. 3(c)).

On the Kirgizskiy Alatoo above 2600 m during the cold season (October–April), the altitudinal variation in monthly precipitation is negative—about -2 mm per 100 m—because of the stable inversion in the bottom of valleys which maintains a low level of condensation. On the Zailiyskiy Alatoo above 2600 m the altitudinal relation is also negative (-4 mm per 100 m) but below 2600 m it is about 12 mm per 100 m. During winter, about 15–20% of annual precipitation is observed in the nival glacial zone and 22–27% of annual precipitation is observed at 2000–2200 m. The foehn cloudiness on the crest-line of the Kirgizskiy Alatoo provides about 5% of annual precipitation in a 1 km belt along the north slope near the crest-line.

During the warm season (May–September), the altitudinal precipitation relation is positive in both regions: about 12–22 mm per 100 m. The cyclonic activity is intensified and the level of condensation is higher. Throughout summer there are, on average, two cases with more than 25 mm day^{-1} of precipitation. During the warm season, the average quantity of precipitation at 3400 m is about 62–72% of the annual total, and about 65% of precipitation falls as snow (Fig. 3(d)).

5. Snow accumulation

The average characteristics of the snow cover regime, i.e. maximum snow water equivalent and dates of the maximum and duration of snow cover, on northern slopes of the Kirgizskiy Alatoo and Zailiyskiy Alatau are shown in Fig. 4 and Fig. 5, based on long-term averages (Anonymous, 1966) from precipitation sites and snow surveys. The average duration of snow cover on the glacial surface is not less than 9 months, but

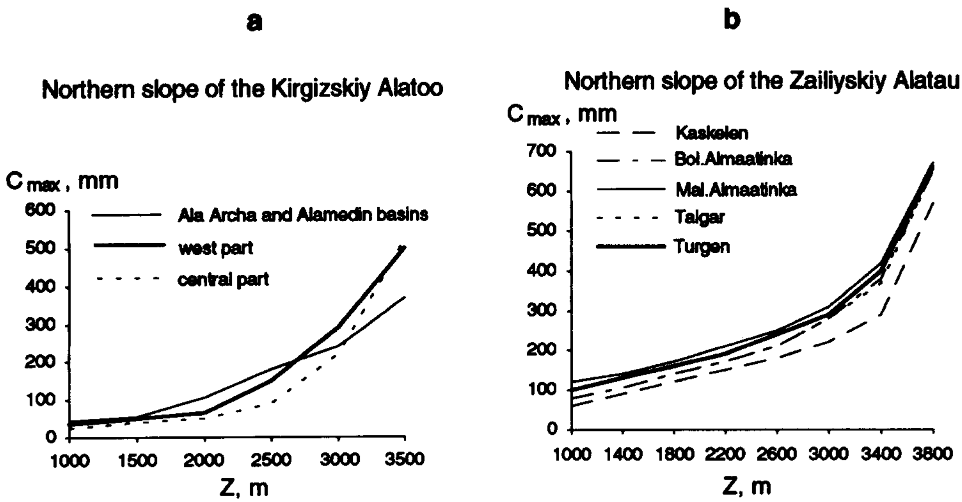


Fig. 4. Distribution of maximum snow water equivalent (C_{max}) on the northern slope of the Kirgizskiy Alatoo (a) and Zailiyskiy Alatau (b).

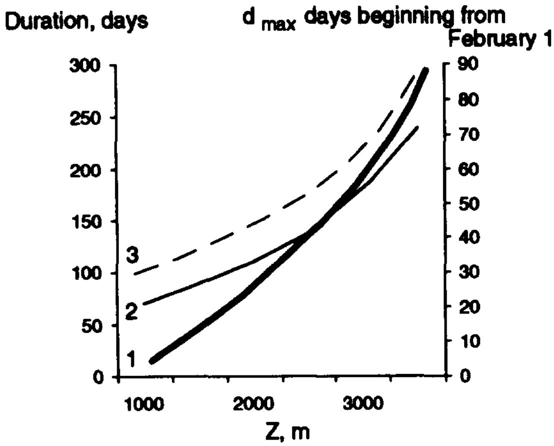


Fig. 5. Distribution dates of snow water equivalent maximum (1) in the northern Tien Shan, and duration of snow cover on the northern slope of the Kirgizskiy Alatoo (2) and Zailiyskiy Alatau (3).

on the moraines the duration is 7.5 months with thickness of not more than 20 cm (Getker, 1988).

The wind regime on the northern slope of the Kirgizskiy Alatoo plays an important role in snow redistribution. The speed of foehn winds can reach 30–40 m s^{-1} in valley bottoms, and air temperature can increase 15°C within 3 h. During a day with foehn winds, 25 cm of snow with a density about 0.18 g cm^{-3} can be melted and evaporated (Aizen, 1988).

Development of katabatic winds in nival glaciation zones is also important for snow redistribution. Comparison of snow survey data and measurements of total precipitation from 1960 to 1982 on the northern slope of the Kirgizskiy Alatoo (Aizen, 1988) and on the Zailiyskiy Alatoo (Tokmagambetov and Erasov, 1985) indicates the same peculiarities in snow redistribution. Snow is removed from windward steep slopes and crests and deposited on firn fields. In lower zones of glaciers and on moraine surfaces, the ratio between values of measured accumulation and precipitation are 0.3–0.8; in cirques of large glaciers, values are 1.0–1.6; at the foot of northern slopes, values are 1.4–2.0. The spatial coefficients of variation of snow accumulation are small (coefficient of variation (CV) is 0.2–0.4) on glaciers. On the basis of monthly avalanche observations, avalanches on the glaciers are not more than 2–3% of the volume of accumulation (Aizen, 1988).

6. Solar radiation

In the bottom of valleys on the northern Tien Shan, the intensity of solar radiation is low in winter (Fig. 6) (Anonymous, 1966) as a result of a strong temperature inversion and development of fog. According to an analysis of records for 23 years (Aizen, 1988), even in summer, the atmospheric clarity above the Golubina glacier decreases from 0.832 at 08:00 h to 0.77 at 17:00 h, because in the second half of the day the dominant wind from the Chuyskaya valley brings loess dust and industrial aerosols. The coefficient of

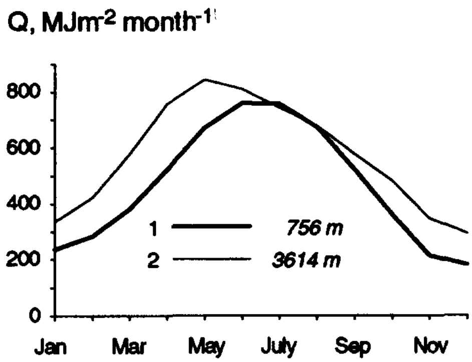


Fig. 6. Long-term mean short-wave radiation (Q) at Bishek (1) and Tien Shan (2) stations.

atmospheric clarity was calculated as the ratio between solar radiation measured during cloudless weather by actinometer and the extraterrestrial value.

Direct radiation under clear skies changes from 59.5 to 67.3 kJ m^{-2} min^{-1} because of variation in atmospheric clarity and height of the sun (Table 2) (Aizen, 1988). The average direct solar radiation changes from 2.1 to 31 MJ m^{-2} day^{-1} depending on cloudiness (Avsuyk, 1984; Aizen, 1988). Sums of total short-wave radiation change during the year from 8.4 to 47.7 MJ m^{-2} day^{-1}, which is 21%–81% of possible radiation income. Solar radiation parameters measured on glaciers are presented in Table 2.

Table 2
Monthly means of short wave radiation (Q, MJ m^{-2} $month^{-1}$), albedo (A,%), net radiation (B, MJ m^{-2} $month^{-1}$), direct solar radiation under clear skies (I, kJ m^{-2} min^{-1}), duration of solar radiance (A, hours) and average cloudiness amount (N,%) on glaciers of the northern slope of Zailiyskiy Alatau (after Avsuyk, 1984) and Kirgizskiy Alatoo (Aizen, 1988)

Glacier	Year	June			August			September		
		Q	A	B	Q	A	B	Q	A	B
Zailiyskiy Alatau, accumulation area										
Constitucii	1964	608	30	423						
Toguzak	1964	687	53	277						
Korjenevskogo	1965	888	30	570	704	35	453			
Tuyksu	1966	511	68	155	465	60	201	432	52	100
	1969	461	71	365	675	56	331	473	78	155
	1971	683	61	356	737	24	377			
	1973	876	53	415	758	48				
		Jun	Jul	Aug	Jun	Jul	Aug	Jun	Jul	Aug
Kirgizskiy Alatoo, 3450 m			I			A			N	
Golubina	1972	57.0	60.8	63.4	115	177	162	61	60	51
	1973	59.5	61.6	67.9	242	268	234	54	31	26
	1975	60.3	61.2	64.5	187	276	212	54	35	40

Table 3
Average monthly air temperature gradients (°C 100 m^{-1}) on northern slopes of Kirgizskiy Alatoo and Zailiyskiy Alatau. Z_l and Z_u are elevations of lower and upper stations, ΔT_s (°C 100 m^{-1}) is the average gradient from June to August

Stations	Z_l	Z_u	ΔT_s	Jan.	Feb.	Mar.	Apr.	May.	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
N. Kirgizskiy Alatoo	726	3120	0.6	-0.2	-0.02	0.3	0.5	0.6	0.6	0.6	0.5	0.4	0.2	0.01	–0.1
N. Zailiyskiy Alatau	818	3750	0.5	0.3	0.4	0.4	0.5	06	0.5	0.6	0.5	0.4	0.5	0.3	0.2
Sub-alpine meadows—spruce forest	1580	2100	0.96												
Spruce forest	2100	2750	0.22												
Forest—alpine meadows	2850	2945	0.49												
Sub-alpine meadows—nival/glacial zone	1580	3450	0.72												
Nival/glacial zone	3450	3800	0.41												
Alpine meadows—nival/glacial zone	2945	3450	0.8												

7. Air temperature regime

Altitudinal distribution of air temperature was approximated by linear functions using the least-squares method. The correlation coefficient between measured and calculated data was 0.83, and the standard error was 6.4 mm. The mean air temperature gradients on the northern slopes in the Kirgizskiy Alatoo and Zailiskiy Alatau are presented in Table 3. In the Ala Archa basin during summer the average daily air temperature difference between valley (2200 m) and glacier (3400 m) changed from 0.2 to −0.7°C per 100 m at night and from 0.9 to 0.75°C per 100 m during the day (Aizen, 1988). Katabatic winds from glaciers decrease altitudinal air temperature differences at night especially in winter. Large differences in gradients are observed between different landscape zones (Table 3).

8. Characteristics of runoff formation

8.1. Melt

In the northern Tien Shan, air temperature can be used as a predictor of snow and ice melt (Eq. (3) is from Konovalov (1979), and Eq. (2) and Eq. (4) are from Aizen (1988)):

$$W_d = 6.4T_d + 8.7 \quad r = 0.71, \ \mathrm{SE} = 1.9 \tag{2}$$

$$W_{de} = 62T_{de} + 54 \quad r = 0.70, \ \mathrm{SE} = 8.3 \tag{3}$$

$$W = 0.32(T_s + 9.5)^{3.46} \quad r = 0.79, \ \mathrm{SE} = 78 \tag{4}$$

where T_d, T_{de} and T_s (°C) are daily, 10 day and summer average air temperature; W_d, W_{de} and W (mm) are daily, 10 day and total annual melt of ice, snow and firn. SE (mm) are standard errors of W_d, W_{de} and W. Eq. (1) and Eq. (3) are based on 32 years of observations on the Golubina glacier, and Eq. (2) is based on 380 days of observations on the Tuyksu glacier. The deviations between calculated and measured values of W_d, W_{de} and W are 3–5%. Relative error of ice and snow melt was 2.3–7.6% of the average (Aizen, 1988).

It is possible to approximate the seasonal melt as a linear function, but the power-law approximation is more statistically significant (Eq. (4)). We expect that the difference in type of approximation of daily, 10 day and seasonal melt occurred because at the same air temperature, the intensity of melt could be different with different albedo or humidity. The linear equation for daily melt is averaged for all summer months (Eq. (2)). In June, with relatively high albedo, the intensity of melt is 4.6 mm °C^{-1}; in July and August, it is 7.1 mm °C^{-1} and 7.4 mm °C^{-1}, respectively, when albedo decreases. On average, it is 6.4 mm °C^{-1} during summer. As assumed, an approximation of seasonal melt by a power law takes into account different factors of melting. Higher temperatures influenced seasonal mean temperature observed in August, when the intensity of melt was highest.

Melt calculated by the daily linear equation (Eq. (2)) is 154 cm at mean summer air temperature of 2.0°C at the altitude of the equilibrium line, and the mean number of days when melt occurred at this altitude is 72 days. Melt calculated by seasonal power-law (Eq. (4)) is 150 cm.

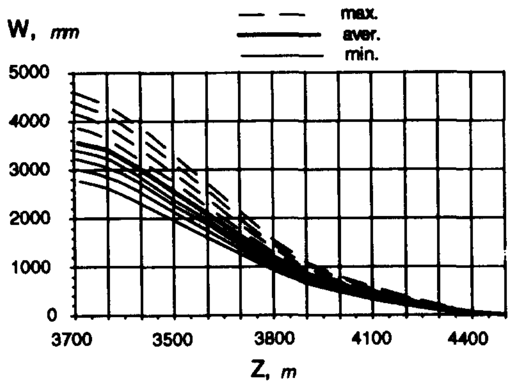


Fig. 7. Altitudinal changes of melt (*W*) on glaciers of the northern slope of the Kirgizskiy Alatoo.

In spite of a linear distribution of air temperature with altitude, the altitudinal gradient of melt on glaciers was not constant (Eq. (5); Fig. 7), because moraine cover on glacial tongues decreases melt intensity. Altitudinal distribution of glacier melt can be approximated as a parabolic function Eq. (5)). The parameters in Fig. 7 were determined by long-term observational data using the least-squares method. Different empirical coefficients of curves from year to year are dependent on meteorological conditions during summer. On the basis of these curves, distribution of total melt with altitude in year may be calculated for a given year using mean of measured or calculated melt at an altitude:

$$W(Z) = a + a_1 Z + a_2 Z^2 \quad (5)$$

where *W*(*Z*) (mm) is annual melting at the *Z* (km) altitude, $a = 37\,482$ (rate: from 28 832 to 48 727), $a_1 = -16\,045$ (from −20 859 to −12 343) and $a_2 = 1710$ (from 1316 to 2223) are experimental coefficients. The correlation coefficient between measured and calculated values is 0.83, and the error of calculation is 60 mm per season. We now consider runoff formation in sequentially larger basins.

8.2. Djindisu sub-basin

The area of the Djindisu sub-basin is 12.6 km^2, and its average altitude is 3700 m (Fig. 1). The stream gauge is located 1 km from the glacial tongue of the Golubina glacier in the rocky canyon of the Djindisu River at 3000 m; this gauge was operated from 1958 until 1991. No permanent snow-patches occur in the sub-basin. In addition to the main Golubina glacier, with an area of 6.2 km^2, there are two small glaciers that are each 0.02 km^2 in area. To calculate glacial runoff we averaged the morphometric characteristics for each of the glaciers within the Ala Archa basin, and a glacier with characteristics which correspond to average parameters was chosen, i.e. Golubina glacier. The Golubina glacier includes the full altitudinal range and morphologic peculiarities of glaciation in the region, and its aspect corresponds to that for most glaciers in the region (Aizen, 1988).

Annual runoff from glaciers (*QRg*) in the Djindisu sub-basin is described by the equation

$$W + Ps \pm E - J = QRg \quad (6)$$

Table 4
Glacial runoff from the Golubina glacier, and water balance in the glacial mountain basins of Djindisu and Ala Archa rivers

Golubina glacier			Djindisu River basin			Ala Archa River basin		
	ave	aw		ave	aw		ave	aw
W	5.1	819	*QRg*	6.5	516	*W*	31.4	878
Wi	0.1	16	*Pn*	4.8	381	*Ps*	8.8	247
Ps	1.4	225	inc	11.2	897	*J*	-3.6	-101
E	0	0	*En*	0.7	-56	*Wi*	0.8	22
J	0.08	-13	*QRs*	8.9	706	*QRg*	37.5	1047
QRg	6.5	1043	*QR*	10.9	865	*Pn*	170.8	866
			exp	11.6	921	inc	208.3	894
			dis	-0.4	-32	*En*	46.1	234
						QRs	87.7	376
						QRs	143.2	615
						exp	189.3	812
						dis	19.0	82

W, melt of snow, ice and firn; *Wi*, glacier melt in crevasses and at bottom; *Ps*, summer precipitation on the glacier surface; *E*, evaporation and condensation on the glacial surface; *J*, refrozen melt water; *QRg*, glacial runoff; *Pn*, precipitation on non-glacial surface; inc, sum of *QRg* and *Pn*; *En*, evaporation and condensation from non-glacial surface; *QRs*, runoff during summer; *QR*, runoff during year; exp, the sum of *En* and *QR*; dis, discrepancy in balance; ave, 10^6 m^3, is long-term average during 1959–1991; aw, mm, average weighted values related to: Golubina glacier area of 6.2 km^2, and area of ablation zone on the glacier, 4.76 km^2, for values of aw (*Wi*); Djindisu basin area of 12.6 km^2; Ala Archa basin area of 233 km^2 for aw (*Pn, En, QRs, QRs*), and glaciarized area of 36 km^2 for aw (*W, Wi, Ps,QRg*).

where *W* is melt of ice, *Ps* is liquid precipitation falling anywhere on the glacier and solid precipitation falling on the area below the snow line position during summer, $\pm E$ is evaporation or condensation, and *J* is repeated ice formation. All values were measured during 32 years. The average annual value of melt from glaciers was 819 mm (Table 4). Daily melt of snow and ice at 3400 m on the glacier changes from 8 to 45 mm day^{-1} depending on meteorological conditions. The inner and under glacier melt (*Wi*) is 2.7% of surface melt. From the tongue of the Golubina glacier to the equilibrium line, all precipitation falling during summer goes into runoff. During summer, liquid precipitation above the equilibrium line is observed on average two or three times (Aizen and Aizen, 1993). The average quantity of summer precipitation (*Ps*) that contributed to glacial runoff was 225 mm. Cicenko (1966) reported that evaporation from the entire glacial surface was compensated by condensation, which is typical for most glaciers on the northern periphery of Central Asia (Konovalov, 1979; Krenke, 1982; Aizen et al., 1993). Average refrozen melt water (*J*) for the whole glacier is 13 mm (Table 4). Between the average lowest level of the glacier terminus, 3200 m, and equilibrium line, 3800 m, the spring and autumn refrozen melt water is 50–80 mm; at altitudes above 3800 m it can be as much as 120 mm $year^{-1}$. The value of spring ice formation is taken into account only during calculations of daily runoff. For calculations of annual runoff this value is not considered, because repeatedly frozen spring ice melts in summer. The average annual runoff from glaciers was 1043 mm (Table 4).

Annual water balance for the sub-basin of the Djindisu River is described by the equation

$$QRg + P_n - E_n = QR \tag{7}$$

QRg is runoff from glacial surfaces calculated by Eq. (6), P_n is precipitation falling on the non-glacial surface during the year, E_n is evaporation from the non-glacial surface, and QR is annual total runoff from the sub-basin measured by the stream gauge. Total runoff is formed by surface and ground water. Evaporation from the non-glacial surface equalled 56 mm (Table 4). Losses of glacial runoff during transit from glaciers to stream gauge were minimal. The river channel from the Golubina glacier to the stream gauge is located on solid bedrock and has several tributaries beginning from lateral moraine terraces. Melt water and liquid precipitation from non-glacial surfaces do not increase river discharge as sharply as snow and ice melt from glaciers. In the nival glacial zone, melt waters completely infiltrate into friable moraine sediments and form ground water which enters the main watercourse and several springs.

The annual glacial runoff from the Djindisu sub-basin is 42–49% of the total and during summer it can increase to 70%. The annual snow melt component from the non-glacial surface amounts to about 8–10%. Variation of annual values of glacial runoff is more than annual variation in total runoff. The coefficient of variation is 0.31 for glacial runoff and 0.17 for total runoff. This difference is caused by greater variations in glacial melt than variations in runoff from the non-glacial surface which are regulated by ground water. The ground component is 303–320 mm, or 36–38% of average annual runoff in the sub-basin (Aizen et al., 1995a).

Typical hydrographs for the Djindisu River (Fig. 8) have two floods: one formed from melt of seasonal snow cover on the non-glacial surface, and the other formed from melt of glaciers. The separation of components of snow melt from the non-glacial surface and glacier melt was based on simultaneous measurements on the glacier and non-glacial slopes. The beginning of snow melt from non-glacial slopes is associated with discharge increasing from low water levels. The first snow melt flood is not large because water saturates the soil first. The second flood is usually larger than the first one because it occurs during the highest air temperatures, when melt becomes sufficiently intense on the glacier and the soil has been saturated. The beginning of a noticeable increase in river discharge usually coincides with the time when seasonal snow cover disappears on the non-glacial surface and glacier melt starts. The beginning of glacier melt coincides with disappearance of snow at the end of glacial tongue. The end of glacier melt and runoff decline occurs at the end of August or middle of September, when glaciers are completely covered by snow. Direct runoff from rainfall is clearly observed on hydrographs only when storm precipitation with an intensity of more than 20 mm day^{-1} occurs. During spring and summer, snow melt and precipitation saturate the soil, and storm precipitation can form flood flashes. The average component of runoff directly from rain around the Djindisu sub-basin is 7–12% of annual runoff volume. The average error in closure of water balance was not more than 5% of total runoff (Table 4).

8.3. Ala Archa, Bolshaya and Malaya Almaatinka river basins

Geomorphologic formation of the Ala Archa basin, glacier distribution and orientations (Aizen et al., 1995a), landscape belts and climatic conditions are similar to those in the central parts of northern slopes in the Kirgizskiy Alatoo and Zailiyskiy Alatau.

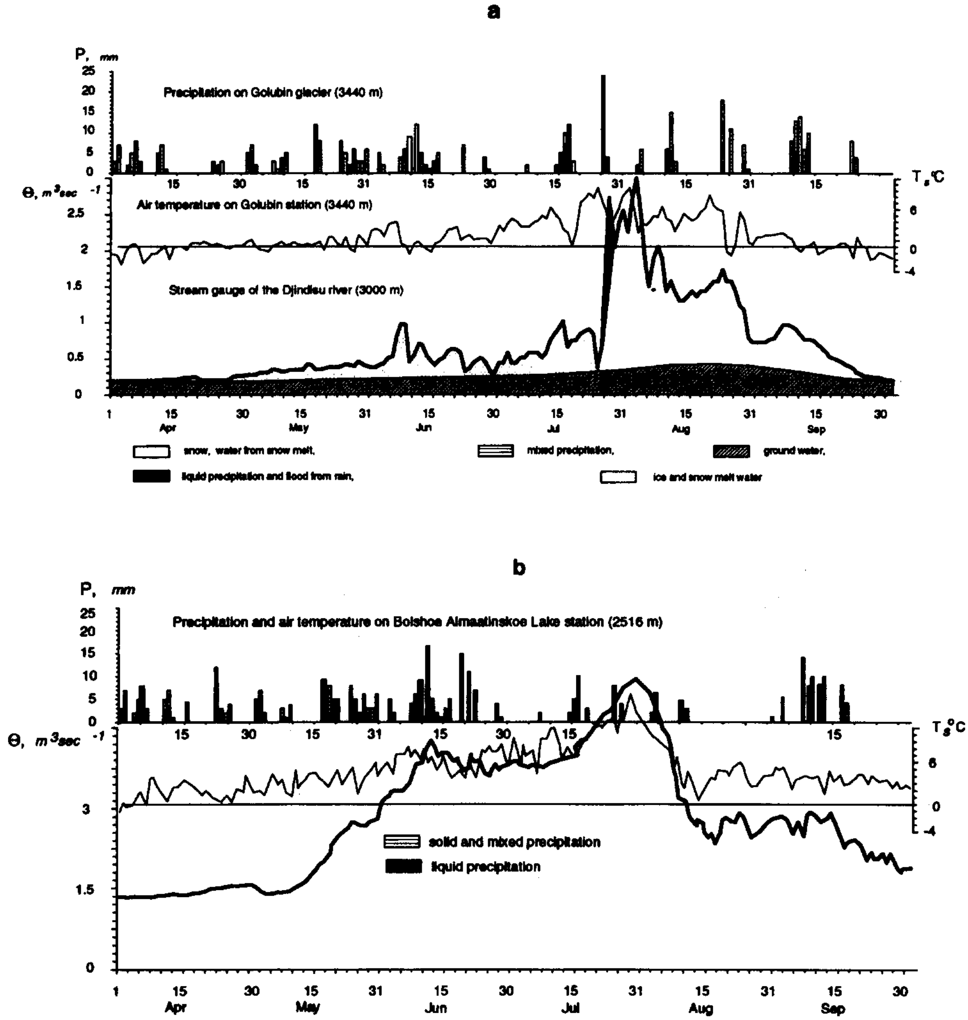


Fig. 8. Typical hydrograph, daily precipitation and air temperature for April–September 1965 of the Djindisu River, Ala Archa basin, northern slope of the Kirgizskiy Alatoo (a), and of the Bol. Almaatinka River, on Bol Almaatinskoe Lake, northern slope of the Zailiyskiy Alatau (b).

The area of the Ala Archa river basin and its glaciation were determined using 1:25 000 scale maps, which were constructed in 1981 and verified by aerial photography in 1991. There were 44 glaciers ranging in area from 0.02 to 6.2 km^2 in the Ala Archa river basin. The total area of glaciers without ice buried under moraines was 35.85 km^2. The nine large glaciers occupied 83% of the total glacial area. The largest glacier was Golubina. The vertical range of glaciation in this basin was 1550 m. The average level of the glacial terminus was 3710 m, but 60% of the glacial area was in the altitudinal range between 3650 m and 4100 m. The centre of glacial distribution coincides with the average elevation of firn line at 3800 m. Lengthy and gently sloping northern and north-western valleys, windward of the main moisture-containing air masses, created conditions for which 87% of total glaciation area occurred on slopes of northern and north-western aspects. Only 0.7% of total glacial area occurred on the steep short southern and eastern slopes (Fig. 1).

Ten years of measurements on six Ala Archa glaciers showed that average annual curves of melt with altitude have similar distributions (Aizen, 1988; Aizen et al., 1995a), and this allows us to calculate melt and glacial runoff in the Ala Archa basin using only measurements at one representative glacier, the Golubina glacier. The initial variable in the nomogram is a value of melt at any altitude that could be measured or calculated by air temperature (eqns (2)–(4)). On the basis of the nomogram (Fig. 7), the altitudinal distribution of melt $W(Zi)$ in the basin was calculated. The calculation of total glacial runoff in the Ala Archa basin was based on the relation

$$QRg = \sum_{i=1}^{N} r(Zi)Fi \tag{8}$$

where QRg (m^3) is volume of runoff, $r(Zi)$ is measured or calculated runoff in one of 100 m altitudinal zones (Zi) on the glacier, N is number of altitudinal zones in the basin from which a volume of runoff is calculated, and Fi is glacierized area of an altitudinal zone. During 32 years, the average annual volume of glacial runoff was 37.510^6 m^3 or 1047 mm (Table 4). The annual quantity of precipitation was calculated on the basis of precipitation data for the averaged weighted altitude of the non-glacial part of the basin (Fig. 3(a)). Annual evaporation from the non-glacial surface was 27% of annual precipitation (Cicenko, 1966). The total runoff from the basin was calculated on the basis of water discharge measurements in the stream gauge at the location 'Ala Archa River—mouth of Kashkasu River'. The average annual discharge equalled 4.54 m^3 s^{-1}. The proportion of glacial runoff was 24% of the annual total and 40% of the summer total. The coefficient of variation of glacial runoff (0.36) was larger than that of total runoff (0.1). The error in closure of the water balance averaged 9% of inputs.

Water losses were caused by infiltration and under-channel runoff of the Ala Archa River. According to hydrogeologic investigations (Cicenko, 1966), the average annual value of losses during transit of water through the basin was 0.42 m^3 s^{-1}, or about 8–10% of input to the water balance. Similar runoff losses were observed on other rivers of the northern slope of the Kirgizskiy Alatoo and Zailiyskiy Alatau (Sumarokova, 1987). The ground water component was about 34–36% of the annual total runoff. Surface runoff of seasonal snow melt from the nival–glacial belt during spring and summer was 18%, and runoff from liquid precipitation was 8%. Liquid precipitation in the forest, meadow and steppe belts are almost entirely excluded from surface runoff. Rain forms surface runoff only when storm precipitation with an intensity of more than 20–30 mm day^{-1} occurs.

On the northern slope of Zailiyskiy Alatoo, glacial runoff in the Malaya Almaatinka River was 18% of the total and on the Bolshaya Almatinka River was 28% of the total. Runoff variations in the Malaya and Bolshaya Almaatinka River basins are presented in Fig. 8.

8.4. Northern slope of the Kirgizskiy Alatoo and Zailiyskiy Alatau

From June to August, the proportion of glacial runoff in most rivers on the northern slopes of the Kirgizskiy Alatoo and Zailiskiy Alatau was on average 18–28% of total runoff and reached 40–60% during summer (Bochin and Krenke, 1987). At stream gauges located close to glaciers the proportion of summer runoff (July–September) was about 52–73%, and at stream gauges located at outlet of mountains the proportion was 30–50%.

Major hydrological events, i.e. flood, peak of flash flood, recession and low water, occurred almost synchronously among the sub-basins on the Kirgizskiy Alatoo and Zailiyskiy Alatau. During summer, southern and south-western cyclonic circulation enters the Tien Shan from south of the Caspian Sea, the head of Murgab River and the head of Amu-Dariya River, and is accompanied by intrusions of warm tropical air masses (Bugaev et al., 1957). Therefore, flash floods on rivers of the Zailiyskiy Alatau are often observed from a day to a week later than in the Kirgizskiy Alatoo, because of a delay in synoptic weather patterns between the ranges. During autumn, when frontal cyclonic circulation develops on cold fronts and influxes of cold air masses do not pass through the mountain ranges, flash floods on rivers of the Zailiyskiy Alatau end earlier.

On tongues of glaciers, melt begins in the middle or end of May. However, melt water does not form runoff at this time because there is sufficient reserve low-temperature material for repeated ice formation from melt water both inside the snow cover and on contact with the glacial surface. Melt water gradually fills cavities in glaciers which have emptied from the previous year. The level of water in rivers originating from glaciers begins to rise at the end of May or middle of June when melt becomes sufficiently intense. At this time, water flows out of glacial cavities and moraines, and daily variations of runoff are directly proportional to intensity of melt. The beginning of a noticeable increase in river discharge usually coincides with the time when seasonal snow cover disappears from lower levels of glaciers. Therefore, in hydrological calculations, the beginning of runoff on the glaciers is the date of snow disappearance on the lowest third of glacier tongues. The end of melt and runoff is the day, usually at the end of August or beginning of September, when glaciers are completely covered by snow (Shulc, 1965; Konovalov, 1979). The duration of melt period on glaciers varies from 35 to 118 days, and is 72 days on average (Aizen, 1988). Sudakov and Tokmagambetov (1968) reported that disappearance of snow cover on the glacial surface, i.e. beginning of glacier ablation, is observed 20 days later than on non-glacial surfaces. During summer, about 3 m of ice and winter snow melt at the lower boundary of the glaciers. Avalanches onto glacial surface are very common in the beginning of summer. Water-saturated masses of snow move down glacial surfaces with 5–7° of inclination angle and clear all the snow along the way. These events, which uncover the glacial surface, hasten ice melt and can produce flash floods in rivers. Average values of glacial melt and duration of melt period at different altitudes of glacial zones of the northern Tien Shan are presented in Table 5. The main hydrological characteristics of the basins and glaciation areas in each one are presented in Table 6.

To extend hydrological conditions of the Ala Archa basin to the whole northern slope, we determined the correlation of runoff in this basin with basins in the central part of the Kirgizskiy Alatoo and basins without or with very small glaciation. Daily runoff in rivers in all basins have high correlation with runoff of the Ala Archa River and Djindisu River (Golubina glacier). The correlation coefficients are not less than 0.78, with standard error of 5.9 mm. High correlation was observed in May and September for all basins (Table 6). During spring, runoff increases everywhere because of snow melt. In the first month of autumn, runoff is regulated by continuing income of ground water. Temporary increases in

Table 5
Melt period of snow/ice (D, days) and melt (W, mm) at different altitudinal zones on the glaciers of Kirgizsliy Alatoo and Zailiyskiy Alatau

	Location	Dates	ΣT_{sp}	D	W_{4150}	W_{4050}	W_{3800}	W_{3580}
ave	Golubina glacier 3400 m	16 June–8 September		72	218	444	1134	2147
σ				20	208	247	359	369
ave	Tuyksu glacier 3017 m	5 July–12 September	249.7	69				

ΣT_{sp}, total of positive temperatures from April to June; Dates are dates of beginning and end of melt, ave, average values, σ, standard deviation.

runoff are associated with autumn precipitation. In summer, the same factor causes opposite effects in glacial and non-glacial basins. For example, increase of air temperature and decrease of precipitation in the middle of July and in August result in decrease of runoff in non-glacial basins. However, the volume of runoff increases in the basins with glacial nourishment.

9. Discussion

By examining runoff from a small sub-basin to larger basins we verified some scales of similarity in the northern Tien Shan where mathematical relationships describing the physical processes are the same, such as altitudinal distribution of precipitation and melt. Major hydrological events, i.e. floods, recession and low water, occurred almost synchronously among the sub-basins and full basins in the northern Tien Shan. The share

Table 6
Hydrological characteristics in some basins of the northern slope of Kirgizskiy Alatoo and correlation to daily discharge in the Ala Archa River basin ($\Theta_{Ala\ Archa}$) and on rivers (Θ_{bas}) in the basins

Basin		May	Jun.	Jul.	Aug.	Sep.	Θ_{bas}	Θ'_{bas}	$Z_{a.w.}$	F	F_g
Alamedin	a	0.21	5.90	5.80	5.14	0.22	6.47	20.4	3260	317	15
	b	1.47	0.82	1.10	1.74	1.34					
	r	0.97	0.88	0.75	0.85	0.99					
Issik Ata	a	0.87	2.51	3.2	6.34	1.10	7.75	14.2	3030	546	10
	b	0.59	0.34	0.41	0.45	0.91					
	r	0.95	0.86	0.85	0.78	0.97					
Karabalta	a	0.62	1.74	7.69	1.93	0.85	6.31	10.9	2910	577	1
	b	0.16	0.12	-0.30	0.22	0.49					
	r	0.81	0.75	0.82	0.74	0.84					
Tyuk	a	0.73	1.54	3.23	5.44	1.92	2.25	12.9	3160	174	6
	b	0.91	0.67	0.22	0.24	0.34					
	r	0.83	0.72	0.76	0.77	0.91					

a and b are regression coefficients ($\Theta_{bas} = a+b.\Theta_{Ala\ Archa}$), r, correlation coefficient; Θ_{bas}, $m^3\ s^{-1}$, average discharge, Θ'_{bas}, $l\ s^{-1}\ km^{-2}$; specific discharge, $Z_{a.v.}$, m, avĕrage weighted altitude of basins; F, basin area km^2; F_g, % is proportion of glacial area of total basin area.

of the main components in the total runoff in different sized basins is also about the same. However, the proportion of water losses caused by infiltration and under-channel runoff varies as a function of basin size. Glacial melt and snow melt from the non-glacial surface also varies as a function of glacial area in each basin (Table 7).

Extending sequentially to larger hydrological units, a divergence in the water cycle components is increasing evaporation from snow on non-glacial surfaces from less than 1% to 27% of annual precipitation and decreasing runoff variability. The coefficient of variation is 0.31 for runoff in a small glacial sub-basin and 0.17 for total runoff in the large basin. In the northern Tien Shan, a key factor determining the runoff in different scale basins is the proportion of glacier area in the total area of the basin.

Results of our analysis of the Tien Shan can be compared with physical processes controlling the terrestrial water balance in other continental mountains at temperate latitudes, such as the Rocky Mountains, Urals, Altay and Pamiro-Alay. These mountains have similar hydroclimatic characteristics because they are located at the same latitudes and in the centre of the North American and Eurasian continents.

Snow and glaciers dominate the montane hydrology. In the Tien Shan, Rocky Mountains, Urals, Altay and Pamiro-Alay, snow accounted for 60–80% of precipitation (Khodakov, 1962; Shulc, 1965; Gray and Male, 1981; Aizen et al., 1995b). Snow and glaciers provide 40–70% of total river runoff in the Pamiro-Alay (Krenke, 1982), Tien Shan (Aizen et al., 1995b; present paper) and Altay (Shpin, 1987), and snow provides about 80–90% in the Rocky Mountains above 2740 m (Goodell, 1966) and in the Urals (Khodakov, 1962)

Western air flows from the Atlantic or Pacific Ocean are the main source of precipitation in these continental mountains. Non-linear increases of precipitation and snow accumulation with altitude are typical in continental mountains (Meiman, 1970; Armstrong and Ives, 1976; Rhea and Grant, 1974; Leavesley et al., 1983; Getker, 1988; present paper), and the highest altitudinal gradient of snow accumulation is observed on the open windward slopes, reaching 100–200 mm per 100 m (Rhea and Grant, 1974; Shpin, 1987; Aizen, 1988; Kotlyakov, 1996). In the Pamiro-Alay, northern Tien Shan, Altay, Urals and Rocky Mountains maximum snow water equivalent is about 800–1200 mm (Armstrong and Ives, 1976; Shpin, 1987; Getker, 1988; present paper). In the continental mountain regions above 2000 m, the temporal variability of snow accumulation is not large; the coefficients of variation are 0.2–0.3 (Aizen et al., 1995b; present paper; Kotlyakov, 1996). Net radiation contributes more energy for snow and glacier melt than the combination of all other forms of heat transfer (Tronov, 1966; Gray and Male, 1981; Aizen and Aizen, 1994).

The location of the Tien Shan and Altay in the centre of the large Eurasian continent and the strong Siberian anticyclonic influence result in hydro-climatic distinctions in the snow regime between Pamiro-Alay, Tien Shan and Altay, and the Rocky Mountains and Urals. Spring–summer precipitation maxima occur with snow and ice melt in the Pamiro-Alay, Tien Shan and Altay, whereas winter precipitation maxima occur in the continental Rocky Mountains and Urals. This results in relatively low spring and summer temperatures in the Pamiro-Alay, Tien Shan and Altay. Snow variations with latitude are seen only within the mountains oriented approximately north–south, such as the Rocky Mountains and Urals. There is extensive glaciation with areas of about 9668 km^2 in the Pamiro-Alay, 7273 km^2

Table 7
Proportion of river runoff components in comparison to total runoff (P,%) in the different sized basins of the northern Tien Shan

	Djindisu river sub-basin	Ala Archa river basin
F_b, km^2	12.6	233
F_g,km^2	6.4	36
$P(QR_g)$	42–49	28
$P(QR_s)$	8–10	18
$P(QR_{g+s})$	50–59	46
P(QRr)	7–12	8
P(QRgr)	36–38	34-36
P(Losses)	0	8-10

F_b, area of basin; F_g, area of glaciers in basin; $P(QR_g)$, proportion of glacial runoff; $P(QR_s)$, proportion of snow melt runoff from non-glacial surface; $P(QR_{g+s}) = P(QR_g)+P(QR_s)$; P(QRr), proportion rain runoff; $P(QR_{gr})$, proportion of ground water runoff; P(Losses), proportion of losses during transit of water through the basin.

in the Tien Shan including 1434 km^2 in the northern Tien Shan, and 910 km^2 in the Altay (Krenke, 1982) maintained by a spring–summer maximum of precipitation and low spring–summer air temperatures. The typical hydrographs in the Rocky Mountains and Urals have one flood formed from melt of seasonal snow cover (from April to July) (Khodakov, 1962; Gray and Male, 1981; Rango, 1988). On the Pamiro-Alay, Tien Shan and Altay rivers there is a second flood formed from melt of glaciers (July–August).

10. Conclusions

Hydrological and climatological characteristics of mountain areas in two basins of the northern periphery of Middle Asia are presented. The main factor determining distribution of precipitation in the northern Tien Shan is elevation. Distribution of precipitation is opposite on the northern slopes of the Kirgizskiy Alatoo and Zailiyskiy Alatau, with larger precipitation gradients at upper elevations on the Kirgizskiy Alatoo and at lower elevations on the Zailiyskiy Alatau. Monthly precipitation is greatest in June.

The foehn and katabatic wind regime on the northern Tien Shan plays an important role in snow redistribution. The ratio between values of measured accumulation and precipitation is 0.3–2.0. Avalanche nourishment for observed glaciers is not more than 2–3% of the entire volume of accumulation.

During ablation, the average daily sum of total short-wave radiation is about 79–81% of possible radiation income at these latitudes. In the northern Tien Shan, the air temperature is a predictor of glacier runoff. The duration of melt period on glaciers of the northern Tien Shan is 72 days on average. During summer, about 3 m of ice and winter snow melt at the lower boundary of the glaciers. In spite of a linear distribution of air temperature with altitude, altitudinal distribution of glacier melt can be approximated as a parabolic function, because of moraine cover on glacial tongues.

In high mountain river basins of the northern Tien Shan with areas of glaciation not less than 30–40%, typical hydrographs have two floods: one formed from melt of seasonal snow cover on the glacial tongue and non-glacial surface, and the other formed from melt of the glacier. The second flood is usually larger than the first. Direct runoff from rain is observed on hydrographs only when storm precipitation with an intensity of more than 20 mm day^{-1} occurs. The average contribution of rain is about 7–12% of annual runoff volume. Glacial runoff is 18–24% of average annual runoff in these basins, and during summer it can increase to 40–70%. Surface runoff from seasonal snow melt during spring and summer is 18%, the ground water component is 36–38% of average annual runoff, and 10% is lost during channel transit in the basin.

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References

Aizen, V.B., 1988. The glaciation and its evolution in the North Tien Shan (in Russian). Ph.D. Thesis, Academy of Sciences of the USSR, Moscow, 218 pp.

Aizen, V.B. and Aizen, E.M., 1993. Glacier runoff estimation and simulation of stream flow in the peripheral territory of Central Asia. IAHS Publ., 218: 167–179.

Aizen, V.B. and Aizen, E.M., 1994. Features of regime and mass exchange of some glaciers on central Asia periphery. Bull. Glac. Res., 12: 9–24.

Aizen, V.B., Aizen, E.M., Nesterov, V.N. and Sexton, D.D., 1993. A study of glacial runoff regime in Central Tien Shan during 1989–1990. J. Glaciol. Geocryol., 3: 442–459.

Aizen, V.B., Aizen, E.M. and Melack, J.M., 1995a. Characteristics of runoff formation at the Kirgizskiy Alatoo. IAHS Publ., 228: 413–430.

Aizen, V.B., Aizen, E.M. and Melack, J.M., 1995b. Climate, snow cover and runoff in the Tien Shan. Water Resour. Bull., 31: 1–17.

Anonymous, 1966. Reference Book of Climate USSR, 1966. Kirgiz SSR, Vols. 18, 19, 31, 32, parts 1, 2, 4 (in Russian). Hydrometeo, Leningrad.

Anonymous, 1978. Catalogue of Glaciers in USSR, 1978. Hydrometeo, Leningrad.

Armstrong, R.L. and Ives, J.D., 1976. Avalanche release and snow characteristics, San Juan Mt., Colorado. Inst. Arct. Alp. Res. Univ. Colo. Occas. Pap., 13.

Avsuyk, G.A. (Editor), 1984. Tuyksu Glaciers (in Russian). Hydrometeo, Leningrad, 170 pp.

Bochin, N.A. and Krenke, A.N. (Editors), 1987. Observational Data on Mountain–Glacier Basins of the Soviet Union under the International Hydrological Decade Program, 1969–1974, Vol. 2 (in Russian). Hydrometeo, Leningrad, 300 pp.

Bugaev, V.A., Djordjio, V.A., Kozik, E.M., Petrosyanz, M.A., Pshenichnii, A.Ya., Romanov, A.A. and Chernishova, O.N., 1957. Synoptic Processes in Middle Asia, Tashkent (in Russian). Hydrometeo, Leningrad, 114 pp.

Cicenko, K.B., 1966. About runoff calculation in altitudinal zones of mountain basins. In: Questions of Kirgizskiy Geography (in Russian). Ilim, Frunze, pp. 62–63.

Getker, M.I. (Editor), 1988. Methodological recommendations for calculations of snow cover characteristics in Middle Asia mountains. Hydrometeo, Tashkent, 145 pp.

Goodell, B.C., 1966. Snowpack management for optimum water benefits. Conf. Water Res. Eng. Am. Soc. Civ. Eng., Denver, CO, October 1965. ASCE, New York.

Gray, D.M. and Male, D.H. (Editors), 1981. Handbook of Snow. Pergamon, Ottawa, Ont., 751 pp.

Grigoriev, A.A., 1973. Hydrometeorological analysis of precipitation distribution in basins of Chu River and Issik Kul lake (in Russian). Ph.D. Thesis, Tashkent, USSR, 214 pp.

Kazanskiy, A.B., 1965. About Richardson's critical number. In: The Physics of Atmosphere and Ocean, Vol. I (in Russian). Academy of Sciences of the USSR, Moscow, pp. 875–879.

Khodakov, V.G., 1962. Snow Cover and Modern Glaciation in the Urals, Vol. 2 (in Russian). Academy of Sciences of the USSR, Moscow, pp. 41–49.

Konovalov, V.G., 1979. Calculation and Forecast of Mountain Glaciers in Middle Asia (in Russian). Hydrometeo, Leningrad, 232 pp.

Kotlyakov, V.M. (Editor), 1996. Atlas of Snow and Ice Resources in the World. Kiev, 820 pp. (in press).

Krenke, A.N., 1982. Mass Exchange in Glacial Systems on the USSR Territory (in Russian). Hydrometeo, Leningrad, 287 pp.

Leavesley, G.H., Lichty, R.W., Troutman, B.M. and Saindon, L.C., 1983. Precipitation–runoff modelling system—users' manual. US Geol. Surv. Water Resour. Invest. Rep., 4238, 207 pp.

Meiman, J.R., 1970. Snow accumulation related to elevation, aspect and forest canopy. Proc. Workshop Snow Hydrol. Queen's Printer of Canada, Ottawa, Ont., pp. 35–47.

Monin, A.S. and Obukhov, A.M., 1954. Main characteristics of turbulent mixing in atmospheric surface boundary (in Russian). Tr. Inst. Geofiz. Akad. Nauk SSSR, 24: 3–17.

Rango, A., 1988. Progress in developing an operational snowmelt–runoff forecast model with remote sensing input. Nord. Hydrol., 19: 65–76.

Rhea, J.O. and Grant, L.O., 1974. Topographic influences on snowfall patterns in mountainous terrain. Interdiscip. Symp. US Nat. Acad. Sci., Washington, DC, April 1973, pp. 182–192.

Shpin, P.C., 1987. Snow anomalies in the North of Eurasia. In: Glaciers and Climate of Siberia (in Russian). Academy of Sciences of the USSR, Tomsk, pp. 44–61.

Shulc, B.L., 1965. Rivers of Central Asia, 2nd edn. (in Russian). Hydrometeo, Leningrad, 291 pp.

Shumskiy, P.A., 1978. Dynamic Glaciology. Translated from Russian by U. Radok and V.J. Vinokuroff. Amerind, New Delhi (original publication 1969).

Sosedov, I.S. and Filatova, L.N., 1961. Results of observations of snow cover evaporation in mountains of Zailiyskiy Alatau (in Russian). J. Meteorol. Hydrol., 8: 18–24.

Sudakov, P.A. and Tokmagambetov, G.A., 1968. Wind regime and snow in nival glacial zone of the Zailiskiy Alatau. In: Hydrological Regime of Kazakhstan's Glaciers (in Russian). Nauka Kazakn., Alma-Ata, pp. 87–95.

Sumarokova, V.V., 1987. About correlation between glacier ablation with precipitation and runoff in high mountain regions. In: Achievement of Soviet Glaciology (in Russian). Ilim, Frunze, pp. 144–147.

Tokmagambetov, G.A. and Erasov, N.V., 1985. Calculation and Forecast of Mountain Glacier Distribution and Regime (in Russian). Nauka Kazakn., Alma-Ata, 159 pp.

Tronov, M.V., 1966. Glaciers and Climate (in Russian). Hydrometeo, Leningrad, 191 pp.