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Article

Features of regime and mass exchange of some glaciers on central Asia periphery

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Abstract

We have analysed alimentation conditions and main features of the regime of mass exchange for some glaciers in Tien Shan, Pamir, South East Tibet and North slope of Himalayas (Fig. 1). The analysis used expeditionary data and long-term records from Golubina, Tuyuksu, Inilchek, Fedchenko, Abramova, Urumchi, Hailougou and Xixibangma glaciers and standard meteorological data from 36 stations in the region. Glaciation in these peripheral areas of Central Asia is less balanced and responds strongly to climatic changes. At the same time, each of these mountain areas has its own specific glacioclimatic conditions which depend on the orography, latitude and relative location in the Central Asia mountain system. Most of the glaciers studied in this work have a trend towards degradation of glaciation in Central Asia.

1. Introduction

The main purpose of this investigation is associated with study of the dynamics of mountain glaciation systems in contemporary conditions, their changes, and the forecasting of possible trends in various climatic regions of Central Asia. To resolve these questions we have obtained quantitative estimates of present variations in the main glaciers mass balance components which take into account the long term fluctuations in general atmospheric circulation. In this work we would like to record some general features that we have determined during expeditionary activities and through the analysis of the long term meteorological data at 36 stations (Fig. 1, Table 1).

2. Northern Periphery

On the NORTHERN PERIPHERY of Central Asia the main factors determining the glacier regime are interaction between the strong influence of south-west branch of the Siberia anticyclone and continuously west cyclonic interaction. In summer, the margins of Tien Shan are under the influence of warm air masses coming from deserts of Central Asia (Aizen, 1990). One of the peculiar feature of the Tien Shan is its location within the zone of the subtropical jet stream in June to August (Djan, 1987). At the same time the northern jet stream moves into this region and merges with the subtropical jet stream, which accords for heavy summer rainfalls.

During expeditionary activities in July - August 1989 and 1990 on the Central Tien Shan we made assessments of the atmospheric moisture in different glaciation zones of Central Tien Shan : in the ablation zone (2800–4000 m altitude sea level), at the average altitude of the equilibrium line position (4000–5000 m) and at firn fields (5000–7000 m). Samples were taken from 2 m above the surface at each observation point. The concentration of tritium varied at the ablation zone from 50 to 150 units during the observation period. Above 5000 m the concentration of tritium in atmospheric moisture of the surface layer varied from 600 to 1300 units. These high concentration of tritium might be associated with the penetration of stratospheric moisture through a rupture of the tropopause during the passage of jet stream atmospheric fronts, or with some other natural phenomena, including heliophysical ones. The preliminary isotope analysis has shown that atmospheric precipitation, over the period of observations in the region, is associated with western (Atlantic) moisture transport.

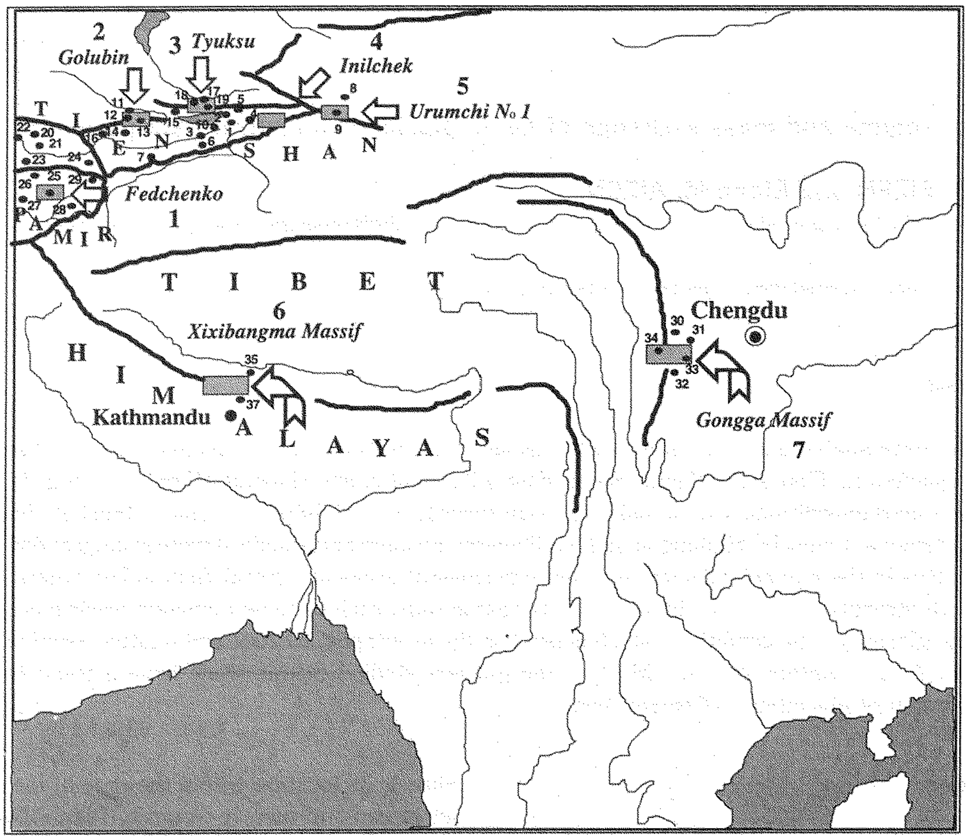

Fig. 1. The map-scheme of Central Asia mountain regions : 1 - Fedchenko glacier (Central Pamir), 2 - basin of Ala-Archa river (Kirgiz Alatoo), 3 - basin of Tuyksu river (Zailiysky Alatoo), 4 - Pobeda-Khan Tengry glacial massif (Central Tien Shan), 5 - Urumchi No 1 glacier (East Tien Shan), 6 - glaciation of Xixibangma massif (North Himalayas), 7 - glaciation of Gongga massif (South-East Tibet).

Table 1. Descriptions of meteorological stations.

Region	No	Name	H, m	Year of beginning
West Tien Shan	20.	Chatkal	1937	1932
	21.	Angren	2286	1930
	22.	Tashkent	477	1872
	23.	Pskem	1256	1932
	24.	Chaar Tash	2748	1960
North Tien Shan	2.	Prjevalsk	1716	1881
	10.	*Karabatkak glacier*	3780	1957
	11.	Baitik	1579	1914
	12.	Ala Archa	2945	1958
	13.	*Golubina glacier*	3440	1958
	14.	Tuya Ashy (North)	3225	1958
	15.	Novorossiyskaya	1510	1926
	16.	Ala Bel	3213	1960
	17.	Mindjilki	3017	1936
	18.	B. Almaatinskoe ozero	2110	1931
	19.	*Tuyksu glacier*	3440	1946
East Tien Shan	8.	Urumchi	918	1940
	9.	*Urumchi glacier No. 1*	3588	1958
Central Tien Shan	1.	Tien Shan	3614	1930
	3.	Narin	2039	1890
	4.	Koily	2800	1950
	5.	Chon Ashu	4005	1960
	6.	Ak Sai	3140	1956
	7.	Chater Kul	3540	1940
Pamir	25.	*Fedchenko glacier*	4169	1933
	26.	*Abramova glacier*	3400	1970
	27.	Anzob pass	3373	1940
	28.	Horog	2075	1899
	29.	Cara Kul	3930	1939
South East Tibet	30.	Kangding	2615	1952
	31.	Shintua	3060	1952
	32.	Tongen	3727	1952
	33.	Luding	1321	1952
	34.	Hailougou	2940	1986
North Himalayas	35.	Tingri	4300	1955
	37.	Nyalam	3810	1955

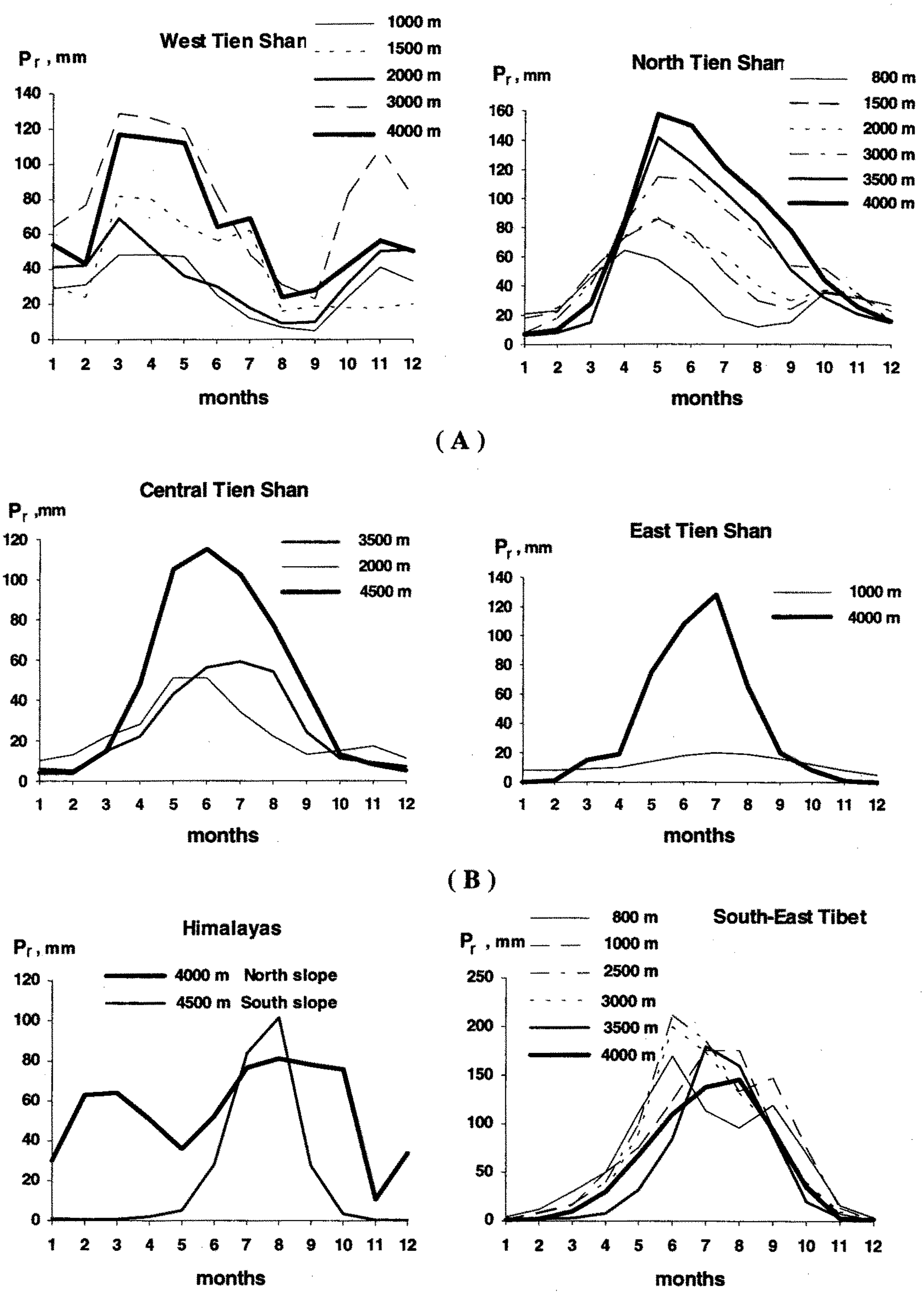

Fig. 2. Annual variation in precipitation on the *North* (A), *South* (B), periphery of Central Asia.

Precipitation which might be caused by the penetration of monsoons from the Pacific and Indian Oceans has not been detected, possibly because of the insufficient observation period.

On the basis of long-term data (1960—1991) from meteorological stations, we may explain the distribution of precipitation on Tien Shan Mts. In the valleys of West and Northwest Tien Shan Mts., precipitation occurs during winter, while in Central and East Tien Shan Mts., winter precipitation accounts for only 8—10% of the annual total (Fig. 2A). In spring, the precipitation may spread to the glacier zone of Northwest Tien Shan. By the end of May 500—1000 mm of snow layer may be accumulated here. At the same time on the East Tien Shan Mts. 80—85% of the annual precipitation falls during summer. We can also see that annual precipitation decreases from West to East and from North to South and its increase with altitude and displacement time maximum, changes from winter to summer. The change in period of maximum precipitation is caused by gradual weakening of the winter anticyclonic activity, and its influence on the west moisture carrying streams in the spring-summer season. The slope orientation towards the main moist air masses and distance from orographical centres influence the altitudinal distribution of precipitation is represented on the Fig. 3. At the same time the precipitation increase with altitude is disproportional for all studied regions. On the North and West Tien Shan the main moist air streams are moving 3000 m and 5000 m altitudes, causing different precipitation distribution through these altitudes Fig. 4. The precipitation type is strongly associated with air temperature in summer. According to curves of rain or snowfall probability (Fig. 5A), received for the glacial areas of North Tien Shan (23 years of observation on Golubina glacier), the liquid precipitation occurs when air temperature are above 5 °C. Thus, rainfall is a very rare phenomenon at the higher equilibrium line. However heavy rainfall at the ablation zone in West and North Tien Shan takes place two or three times during summer and greatly increases the ablation process on the local glaciers.

For analysis of main ablation components we consider the temperature as well as solar radiation and evaporation. According to expeditionary measurements at different altitudes in the Central Tien Shan during 3 years, and 23 years in the North Tien Shan, it was established that the typical feature of temperature regime is the small range of the air tem-

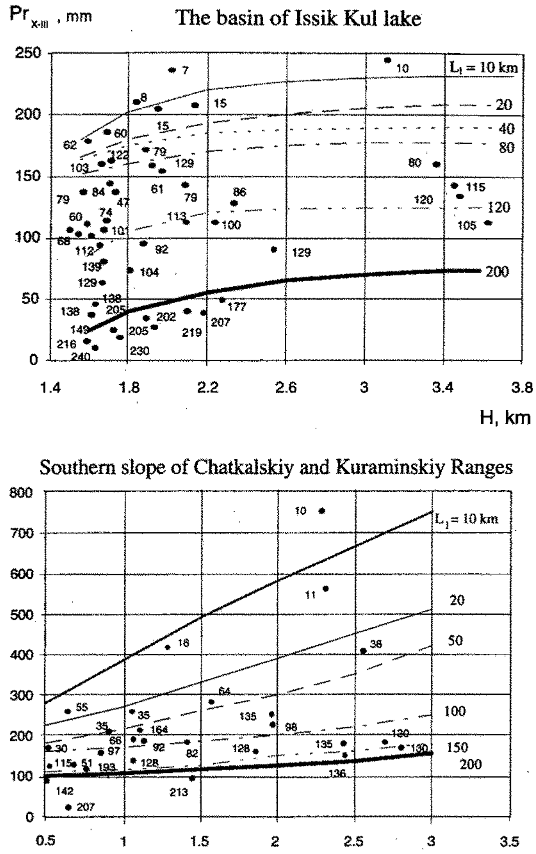

Fig. 3. The changes of precipitation with altitude throughout October-March under the fixed values of distance up to orographycal centres (L_1).

perature gradient over the glacial surface. In Central and Northern Tien Shan its value is about 0.36—0.4 °C/100 m, because of katabatic winds influences.

On the basis of the expeditionary data, including gradient measurements of air temperature, humidity and wind speed, the values of evaporation and condensation flows were calculated and measured by the weight method (Aizen *et al.*, 1991). Also, on the basis of expeditionary data, all components of heat balance during summer ablation period were calculated (Table 2) (Aizen, 1987 ; Aizen and Aizen, 1993). The typical peculiarity for glacial regions of North periphery is compensation between evaporation and condensation from glacial surface.

In the North Tien Shan the heat of turbulent exchange is equalled 15% (Kirghizkiy Alatoo) in the

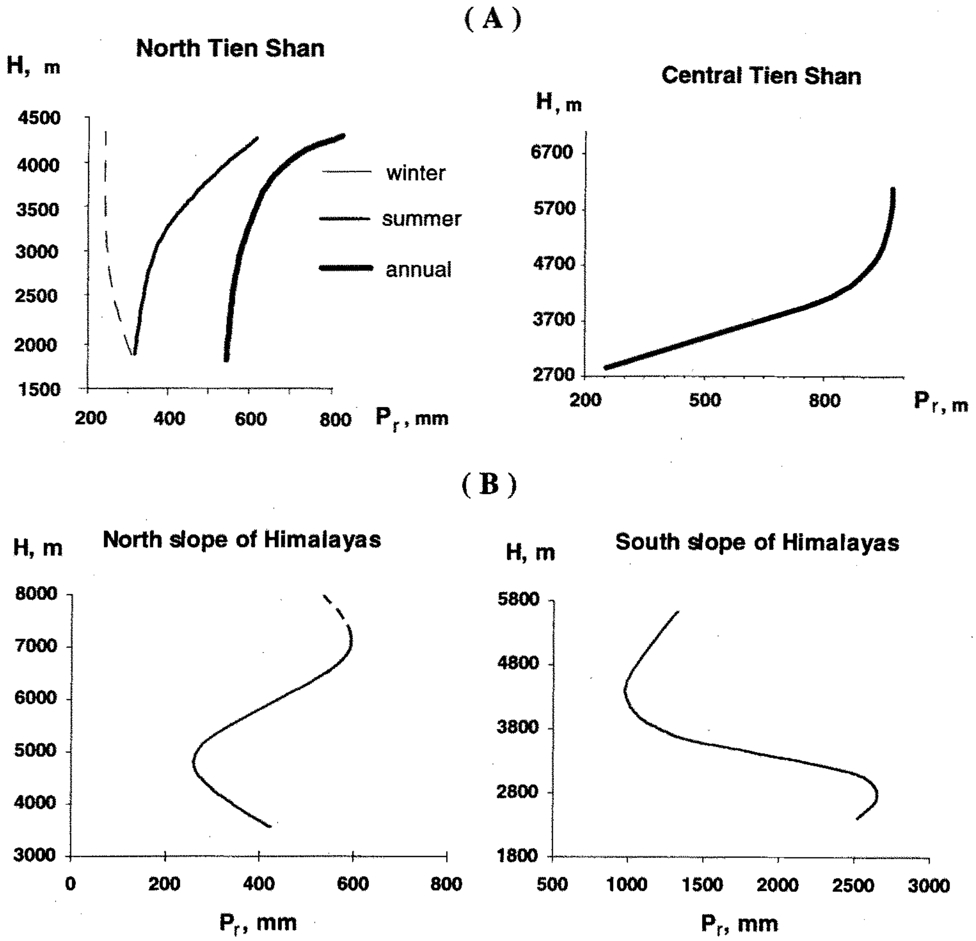

Fig. 4. The vertical change of annual precipitation on North (A) and South (B) periphery of Central Asia.

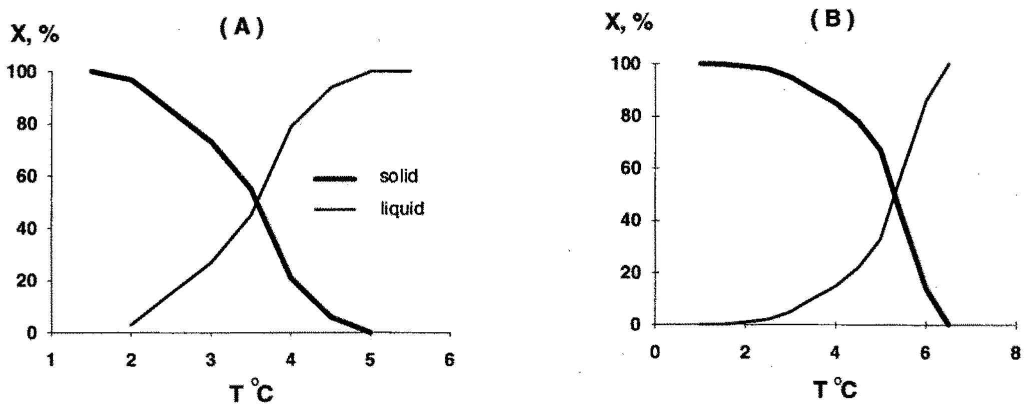

Fig. 5. Probability curves of different summer precipitation falls versus, the general numbers of cases X in relation to the mean daily air temperature on the North (A) and South (B) periphery of Central Asia.

Table 2. Average components of the balance during melting period corresponding to period of expeditionary activities.

Region	B $MJm^{-2}day^{-1}$	%	P_t $MJm^{-2}day^{-1}$	%	LE $MJm^{-2}day^{-1}$	%	IW $MJm^{-2}day^{-1}$	%	Month of observat.	Years of observat.
C. Tien Shan	8.5	96	0.4	4	−0.6	7	−8.3	93	VIIVIII	1989-90
N. Tien Shan	9.3	85	1.6	15	−0.8	7	−10.1	93	V-IX	1958-92
S-E Tibet	3.6	45	2.8	35	+1.6	20	−8.0	100	IX-X	1990
N. Himalayas	2.4	86	0.4	14	−1.6	57	−1.2	43	VIII-IX	1991

B is total radiation balance, P_t is turbulent heat fluxes, ± LE heat fluxes connected with water exchange + condensation − evaporation, IW is heat used for melting.

Central Tien Shan it is not more than 4% (Table 2). Therefore the melt regime is determined mainly by solar radiation. So it is better to calculate the melt and runoff through the solar radiation data (1) or to use air temperature with differentiation of weather types (2−3).

Central Tien Shan :

$$W_i = 1.59 \cdot 10^{-3} \cdot [23.9 \cdot Q(1-A)]^{1.68} \quad (1)$$

where W_i is daily value of snow and ice melt, Q is net short wave radiation, A is albedo.

$$W_{i(Aw,Cw)} = 12.1 + 12.5\ T \text{ when } T > 0.9\ ^{\circ}\mathrm{C} \quad (2)$$

$$W_{i(Ac,Cc)} = 40.0 + 4.6\ T \text{ when } T < 0.9\ ^{\circ}\mathrm{C} \quad (3)$$

where $W_{i(Aw,Cw)}$, mm is the melt during warm anticyclone and cyclones weather types, $W_{i(Ac,Cc)}$ is melt during cold types of weather.

North Tien Shan

$$W_i = \begin{cases} 0.64 \cdot T + 0.87 & \text{when } T > -1.4\ ^{\circ}\mathrm{C} \\ 0 & \text{when } T < -1.4\ ^{\circ}\mathrm{C} \end{cases} \quad (4)$$

where T is daily average air temperature.

These equations, as well as the values of gradients of the temperature, solar radiation parameters and precipitation, were prepared on the base of expeditionary activities and long term meteorological information. Through the calculation of melting the elevation, angles and aspects were taken into account in equation (5).

$$Q_{i(slope)} = Q_{i(hor.)} \cdot \{K_{n,s,e,w} \cdot [\sin(\alpha)] + \cos(\alpha)\} \quad (5)$$

where $Q_{i(slope)}$ is net short wave radiation on the slope of N, S, E, W exposition, K is the coefficient of recalculation taken after Rusin (1979), α is angle of inclination of slope.

The detailed description of calculations by this equation is represented in Aizen (1987) and Aizen and Aizen (1993). It is interesting to note that at the altitude of the equilibrium line the summer air temperature on the Central Tien Shan, as well as on the Central Pamir, has negative values (Table 3). This illustrates the mainly radiation component of melting of this area, which is proved by the quantitative estimations : on the C. Pamir and C. Tien Shan the radiation component in heat balance is equalled more than 90% (Suslov *et al.*, 1980 ; Aizen and Aizen, 1993).

Analysis of summer air temperature and annual precipitation space-time oscillations has shown that all runs have a series of synchronous short periodic variations of precipitation and air temperature (Fig. 6). This periodicity is expressed in interchange of arid or humid years. It is interesting to note that the nature of precipitation and air temperature opposite run on the large territory of Tien Shan. The increase of air temperature is generally accompanied by decrease of precipitation (Table 4). This peculiarity is result of frequent spring and summer snowfalls, influencing the temperature regime in the nival glacial zone. According to data received from instrumental measurements from the meteorological stations over the past 100 to 200 years the mean of air temperature in North Tien Shan has risen 0.9°C in summer and annual precipitation has decreased 100 to 150 mm on the North Tien Shan (Fig. 6A). From 1890 until 1990 the curves were calculated from instrumental meteorological records, and before 1890 by the dendrochronological analyses of tree rings. For summer air temperature estimates the Turkestanian cedar (*Juniperus Turkestanica Kom.*) and for annual precipitation the Tien Shan firtree (*Picea Schrenkiana Fisch. et May*) were sampled in the forest belt at the heights of 2400−2800 m (Aizen and Solomina, 1993).

According to the main received components of

Table 3. Main statistical indexes of the Central Asia glacial systems regime.

No	Glacial system	Prevalent aspect	φ^o N.I.	S_g km²	H_h-H_l m	T_s,°C ave.	σ	min.	max	P_r, mm ave.	σ	min.	max.	d	$H_{e.l.}$, m ave.	σ	min.	max.	B, g/sm² ave.	σ	min.	max.
1.	N. Tien Shan	N, N-W	42	1651	5020 3000	2.0	0.5	0.8	3.6	697	127	392	1058	5	3848	120	3560	4220	-15.0	32	-110	52
2.	C. Tien Shan	W	41	4320	7500 2600	-1.8	0.4	-2.7	-0.6	840	61	708	988	6	4476	47	4416	4646	-31.8	13	-47	16
3.	C. Pamir	N	38	3480	7120 2580	-0.2	0.9	-2.0	2.1	1192	284	728	2060	3	4780	186	4338	5187	-41.0	78	-207	150
4.	S-E Tibet	E, S-E	29	90	7400 2900	4.9	0.7	2.1	6.2	2207	110	1758	2149	6	5200	171	4839	5647	-9.9	49	-123	99
5.	N. Himalayas	N	28	4840	8000 5500	2.4	0.4	1.7	2.9	558	94	436	901	8	5900	84	5700	6300	-3.4	27	-84	111

φ - average mean of glacial system latitude, S_g - area of glaciation system, H_h-H_l- upper and lower levels of glaciation, T_s- means of summer temperature at the equilibrium line, σ- standard deviation, P_r- annual precipitation at the equilibrium line, $H_{e.l.}$-position of equilibrium line, B- mass balance glacial system, d-month of precipitation maximum.

Table 4. Correlation coefficients of regime indexes in glacial systems of Central Asia.

Glacial systems	T_s/P_r	$T_s/H_{e.l.}$	$P_r/H_{e.l.}$
N. Tien Shan	-0.17	0.85	-0.35
C. Tien Shan	-0.10	0.79	-0.59
C. Pamir	-0.32	0.90	-0.60
S-E Tibet	-0.45	0.73	-0.53
N. Himalayas	-0.42	0.77	-0.58

T_s^o- summer air temperature at the equilibrium line, P_r - annual precipitation, $H_{e.l}$- equilibrium line position.

mass-balance (Table 3), we calculated the index of mass-balance for all studied glaciers during the period when instrumental records were available. The nearest meteorological stations which correlate with corresponding glaciers or glaciation area were selected. The principle of the selection of stations was the value of the correlation coefficient in the characteristics of summer air temperatures and annual precipitation between glacier and station (not lesser than 0.68). For calculation of mass-balance index (I_{bi}) equation (6) was used :

$$I_{bi} = [P_{H_o} + \gamma(p) \cdot (H_{e.L} - H_o)/100] - [A_{H_o} - \gamma(a) \cdot (H_{e.L} - H_o)/100] \qquad (6)$$

where P_{H_o} is the annual precipitation (October-September) on the level of the meteorological station, $\gamma(p)$ and $\gamma(a)$ are the altitudinal gradients of the precipitation and ablation determined through the results of field studies, H_o and $H_{e.L}$ are altitudes of the meteorological station and long term average value of the equilibrium line position calculated by Kurowsky's method (Kruowsky, 1891), A_{H_o} is the ablation value calculated by equations (1-4), considering that $A = W$, because of compensation between evaporation and condensation here.

According to calculations all glaciers have a negative mass balance on average. The highest values of negative balance are typical of glacier systems in the Northern periphery of Central Asia, especially for Pamir (Fig. 7), where glaciation conditions are determined, to a considerable extent, by the influence of Atlantic air. At the same time the variability of the mass balance and the equilibrium line position for Central Pamir are the greatest (Table 3), whereas greatest stability of these characteristics is typical for Central Tien Shan. Consequently it proves that the outlying areas of the Central Asian mountain glacier systems are subjected to climatic changes more intensively than internal ones.

During 26 years of synchronous observations the balance mass sign was the same on the glaciers Central Tuyuksu and Golubin in 85% of cases ; Golubina, Inilchek and Tuyuksu, Inilchek in 69% of cases ; Inilchek, Urumchi in 85% ; Tuyuksu-Urumchi in 50% ; Golubin, Urumchi in 69% (Fig. 7). High synchronous changes are typical for mass balance of Central Tuyuksu, Golubin and Inilchek, Urumchi No. 1 glaciers. The two first glaciers are on the North periphery of Tien Shan in the zone of temperate precipitation with a spring-summer maximum. The two second glaciers are in zone of insufficient precipitation with a summer maximum. The similarity of all glaciers is determined by the dominant role of west and north-west atmospheric flows over the Tien Shan. The quantitative and qualitative differences are associated with geographical location of glaciers in mountain systems and specific morphology. According to (Fig. 6) the decrease in precipitation quantity is

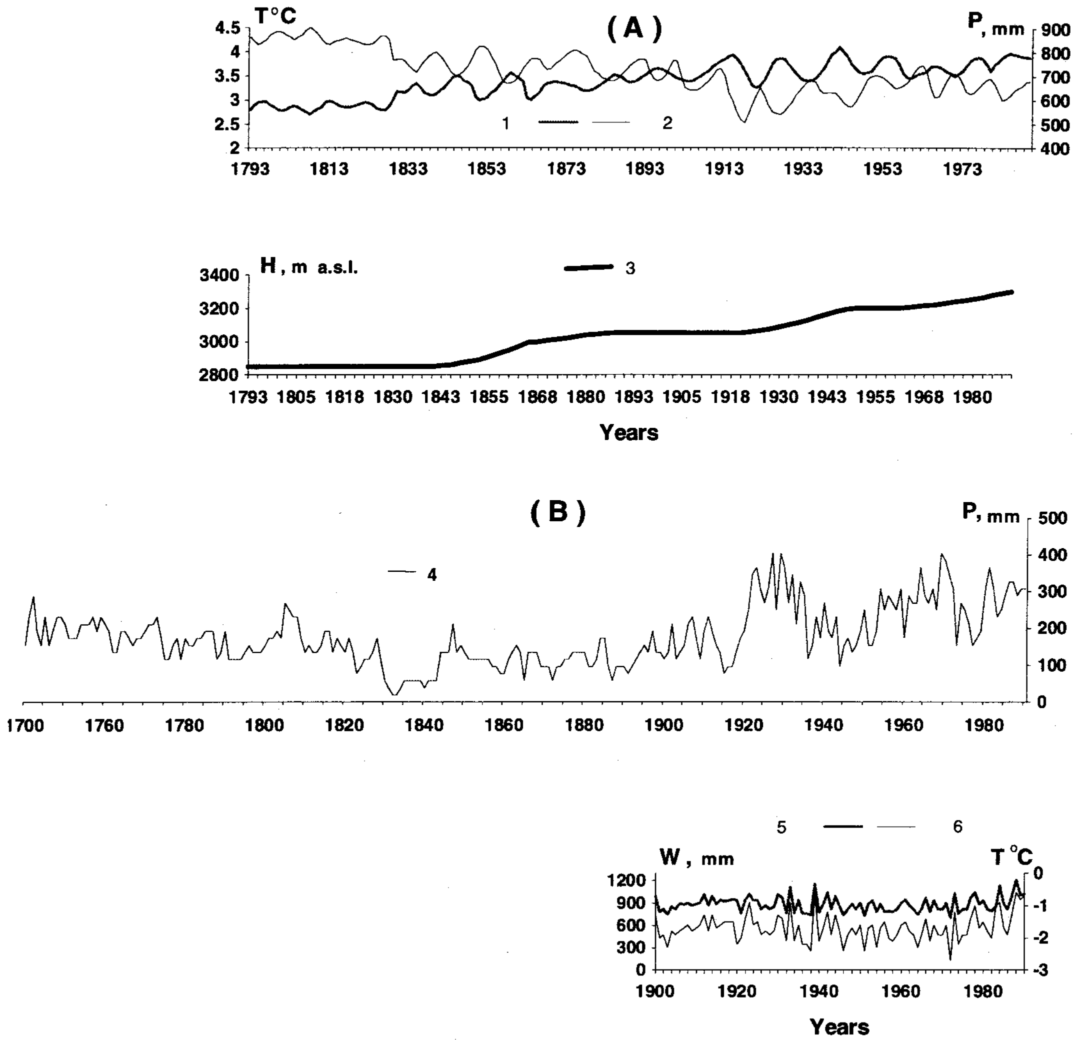

Fig. 6. Reconstructed summer air temperature - 1, annual sums of precipitation - 2 glacier Golubin tongue position - 3 on North Tien Shan (A), and annual sums of precipitation - 4 on Central Tien Shan (B) received on the basis of topographical survey from 1890, meteorological observations from 1914 and 1930, geomorphological and dendrochronolgical data from the end of 16th century. Summer air temperature - 5 and melting calculated by it - 6 at average equilibrium line position of Pobeda-Khan Tengry glaciation massif in Central Tien Shan (B).

observed on the North Tien Shan with a tendency of precipitation increase in Central Tien Shan. We think this is caused by weakening of anticyclone intensity because of the global air temperature increase. Such distribution of precipitation has resulted in intensive degradation of glaciers in peripheral mountain regions of the North Tien Shan. The glacier tongues were retreating during the period analysed (Fig. 6), but small shifts of the glacier front did not correlate with the corresponding phase of fluctuations climatic parameters (Aizen, 1987). The short periods with positive mass budget do not affect glacier evolution, they cause only insignificant oscillations of the glacier tongue and changes in the surface level. To the middle 1950-s the glaciers front reached the some level as in 1982. Since that time the glacier front appears to be constant. This is demonstrated by the almost zero integral mass budget for the period from 1960 to 1982 (Aizen, 1987). From 1982 the glaciers front started to retreat. This process was preceded by continuous lowering of its surface starting from 1977 which caused a decrease level by 25—30 m.

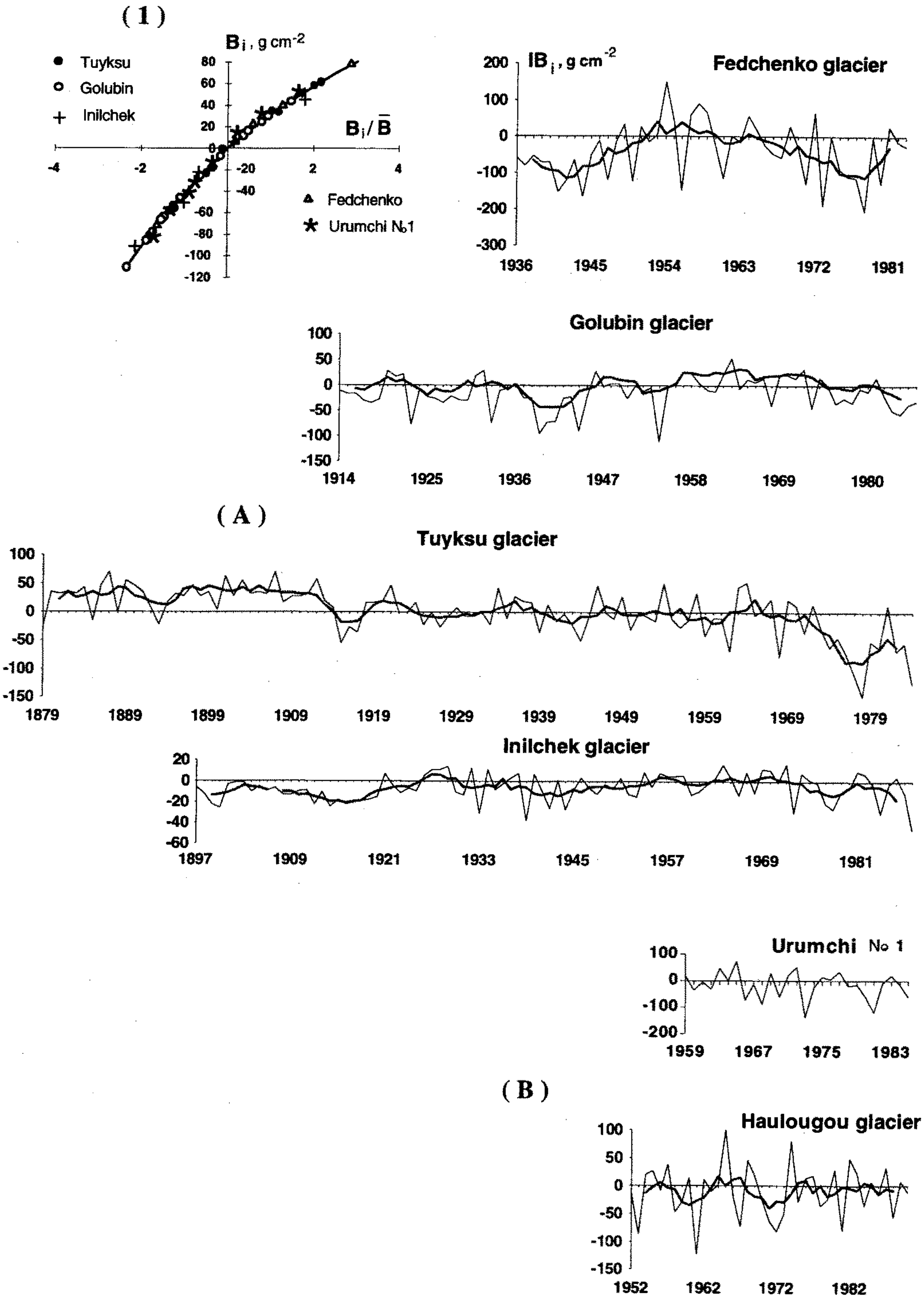

Fig. 7. Correlation between mass balance values (B_i) of five glaciers on the Northern periphery of Central Tien Shan with relative parameters of climate variability (B_i/B) in climate anomaly years (1). Changes of mass balance on the glaciers of North (A) and South (B) periphery of Central Asia.

There was an increase of precipitation in the Central Tien Shan (Fig. 6) at the beginning of the 1800s (about 150 mm at a height of 2800 m) until the beginning of the 1900s (up to 300−350 mm). These fluctuations of the annual precipitation apparently correspond to the change of the predominant forms of circulation - meridional and zonal during the period under consideration. For example, by the middle of the 1800s subcontinental regions of the Central Asia were under the influence of the increased Siberian anticyclone (Vangengeim, 1946), which decreased the possibility of the cyclones influx into the internal regions of high mountain systems. As a result, the volume of precipitation decreased and the conditions of glacial alimentation were worse. At that time there were increase in the number of days with cloudless weather, the summer air temperature, radiation melting, and glacial runoff. This is indicated by the melting curve calculated from summer air temperatures based on the recurrence of warm and cold types of weather (Fig. 6B). The primary meridional forms of circulation are replaced by a zonal one (Vangengeim, 1946 ; Dzerdzeevsky, 1974). The summer favours transport of cyclones into the internal regions of Tien Shan - cloudiness and precipitation increase, radiation melting decreases, and the summer temperature decreases. It appears that this period is more favourable for glacier m to 5−7 km over the last 100−150 years (Bondarev, 1971). Contemporaneously the positive difference in glacier area and relatively low velocity of ice flow (15−45 m/year) indicates to a long response of glaciers in this system to climatic changes. The beginning of contemporary degradation in the Pobeda - Khan-Tengry massif glaciation should be considered to be the end of the so-called "Little Glacial Age."

3. Southern Periphery

The quantitative development of mass-energy exchange components of two different climatic regions in south periphery of Central Asia has been demonstrated by analysis of two years field work and long term data from Chinese meteorological stations. The first one is Gongga massif of South-East Tibet, windward slope of the summer monsoon. The second is Xixibangma massif, leeward of the summer monsoon, on the northern slope of the Himalayas. The glaciers in these high-mountain regions exist because of climatic features connected with the monsoon circulation, Tibetan anticyclone, and the subtropical jet stream. The intensity of solar radiation during summer-autumn monsoon for ablation zone of South-East Tibet is one sixth of that theoretically possible for these latitudes. Here, in the ablation zone (3400 m), the average insolation intensity during the summer -autumn monsoon is 6.9 MJ $m^{-2}d^{-1}$ (Aizen and Aizen, 1993). Theoretically, the radiation amount to which this region could be exposed is great (25°−30° north latitude ; $I = 0.0836$ MJ m^{-2} min^{-1} - extra-territorial value). Without atmospheric effects (reflection and absorption from clouds, …) during the summer it equals 7315 MJ m^{-2} (40 MJ $m^{-2}d^{-1}$ (Alisov *et al.*, 1952)). Correspondingly for the north slope of the Himalayas it is half of that theoretically possible (Aizen and Aizen, 1993). In the Southeast Tibet the continuous rising of air masses, increasing condensation processes and turbulent heat exchange at the same time, causes large precipitation during summer monsoon. On the north slopes of Himalayas the continuous katabatic winds don't yield such significant precipitation, but they sustain the evaporation processes, even in ablation zone, reducing the glacier melting (Table 2). The melt is less intensive because of specific atmospheric conditions on the high altitudes and large values of reflected radiation (albedo changes from 77 to 99%). Here the high values of albedo remain for almost all the summer owing to frequent snowfalls (about one every 3−4 days), generally at night. The high permeability of the atmosphere is shown by small or zero values of long-wave atmospheric radiation, that yields low values of total heat balance. During September, 1991 the mean was only 1.6 $MJm^{-2}d^{-1}$ (Aizen and Aizen, 1993).

Now there are many investigations of the circulation-synoptic processes in the Himalayas and Tibet, especially of southern Himalayan slopes. According to Nakajima *et al.*, (1976), Yasunari (1976) and Ramage (1976) on the southern slopes of Himalayas, the moisture source is the Indian ocean. The moisture is brought to Nepal by monsoon formed in the south and even in the southern hemisphere. The wave activity on the northern Pakistan and northern coast of Bay of Bengal is also important. Its development is intensive during the summer Tibetan anticyclone weakening. However, aerological investigations of the cyclones trajectory (Lautensach, 1949) have shown that, in summer, the intrusions of extratropical cyclones, *i.e.* the cyclones belonging to the extratropical western wind zone, are registered

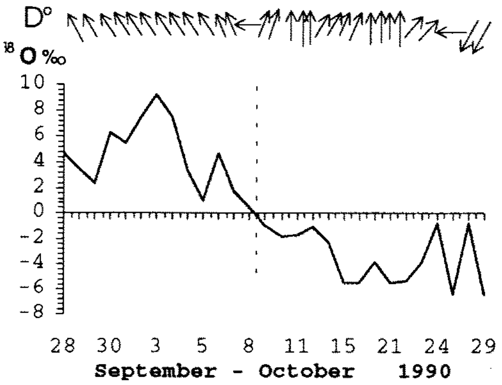

Fig. 8. Variation of the daily wind direction and deviation from the average value of stable oxygen δ^{18}O concentration in atmosphere precipitation during the expedition investigation on Hailougou glacier at 3400 m in sub period of summer monsoon.

Table 5. Average statistics parameters of meteorological characters during summer monsoon period of the expeditionary observation in 1990 at 3400 m height, Hailougou glacier.

	extratropical cyclone				tropical monsoon			
	ave.	max.	min.	σ	ave.	max.	min.	σ
D	101	239	57	24.1	180	260	38	47.7
V	1.3	2.5	0.5	0.4	1.0	2.3	0.5	0.4
P	664	667	662	2.2	682	714	666	8.6
Q	6.6	12.3	3.4		7.0	15.2	3.0	
B_Σ	4.6	8.7	2.6		3.3	10.7	1.4	
T_a	5.8	19.6	2.5	3.4	4.7	23.3	-0.2	4.0
q	86.4	98.1	41.4	12.3	87.9	104.0	33.5	12.5

D, deg. - wind direction; V, ms^{-1}- wind speed; P, mb- atmosphere pressure; Q, MJ m^{-2} day^{-1}- incoming short wave radiation; B_Σ, MJ m^{-2} day^{-1}- total radiation balance; T_a, °C- air temperature; q, % - air relative humidity.

over investigated territory. According to Bluthgen (1972, 1973) and Khromov (1968) in the summer the zone of western equatorial winds advances to the north due to the difference of the northern and southern hemisphere warming. The extratropical frontal zone moves from south to the north and back in summer and results in a double maximum of precipitation during summer (see Fig. 2B). At the same time, due to Nakajima *et al.* (1976), Yasunari (1976) and Ramage (1976) in winter season the snowfalls caused by moving of subtropical jet stream from temperate latitudes are registered. The subtropical jet stream strengthening during winter rounds the blockading systems of Tibet and brings precipitation into high mountain regions (see Fig. 2B). According to Nakajima *et al.* (1976) and Wushiki (1977) the main precipitation on the southern slopes of Himalayas is connected with air masses from South, i. e. from Indian ocean. Yasunari (1976) wrote that in high mountain regions moist air masses come from the direction of upper troposphere west streams.

With the aim of establishing of atmosphere precipitation genesis on the Hailougou and Xixibangma glaciers, meteorological and isotopic observations were made. According to our data recorded in South-east Tibet the monsoon period may be divided into two sub-period : 1) to October 8 and 2) from 9 to 24 October (Aizen and Aizen, 1993). The wind direction changed on 8 of October from SE to SW and S (see Fig. 8). This was accompanied by some changes in other meteorological phenomena (Table 5) although temperature remained constant.

The data obtained by measurement of the stable oxygen isotope (δ^{18}O) content in the atmospheric precipitation at 2940 and 5000 m heights on the Hailougou glacier (Fig. 9B) during summer monsoon show that there are also two sub periods : up to October 9 and after it. Figure 8 presents a curve showing the deviation of the concentration of oxygen isotope in precipitation from the average value during the observation period $\Delta\delta^{18}$O (the average value is 15. 9‰). Under equal temperature conditions the more negative the δ^{18}O and δD values correspond to the further disposition of atmospheric moisture source. Difference between the oxygen isotope concentrations gave us the possibility to suggest that before October 9 the glaciers at heights of 2940 and 3400 m was supplied by precipitation coming with air masses from the Atlantic Ocean. After October 9 it was the monsoon originating in the Indian or Pacific Oceans which served as the sources of precipitation.

The changes of δD concentration suggested that in course of second sub period of summer monsoons evaporation from the land surface into air currents was the prevailing process. This fact confirms the point that the moisture-bearing air currents developing over the Indian and Pacific Oceans move at relatively low altitudes and have more restricted distribution than the Atlantic air masses transported and depositing moisture at comparatively higher altitudes. As a result the west moisture-bearing air currents are subjected to incoming of evaporated moisture a lesser extent. This is confirmed by deuterium isotopic measurement data (Fig. 9B).

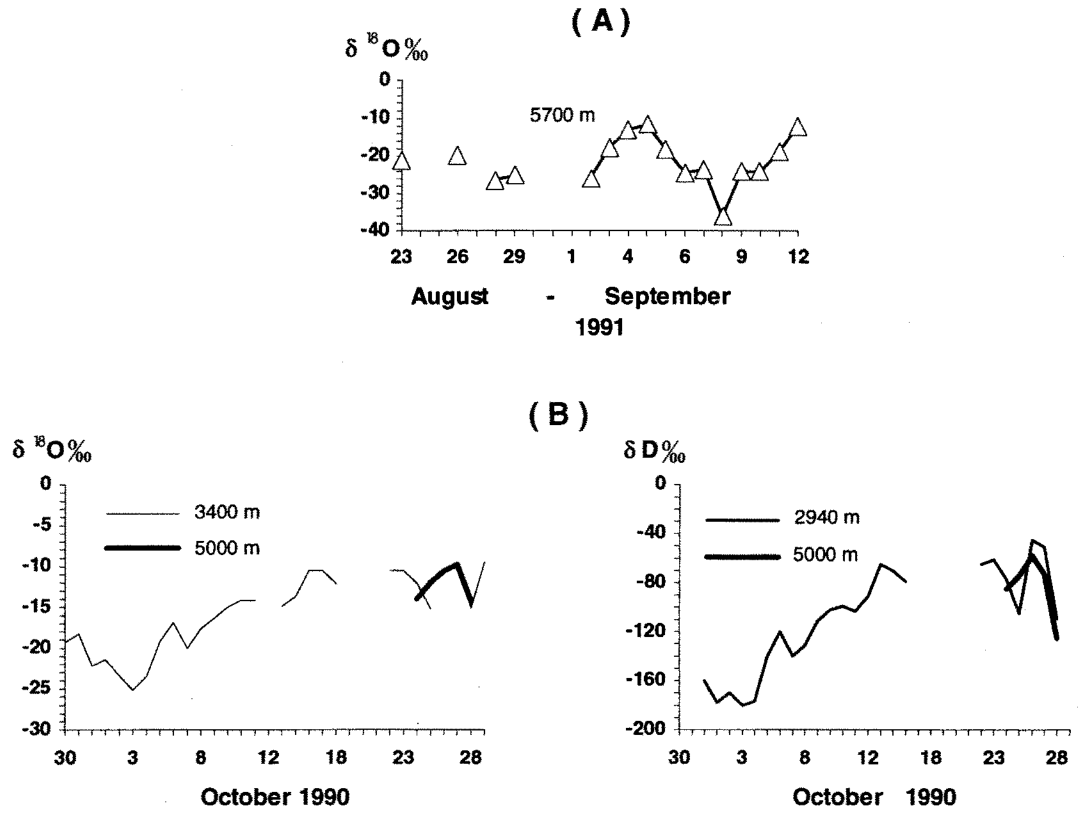

Fig. 9. Distribution of stable $\delta^{18}O$ (1) and δD (2) isotope in atmosphere precipitation during summer monsoon on Xixibangma (A) and Hailougou (B) glaciers.

In course of expeditionary observations on Xixibangma glacier such distinct periods of time with different $\delta^{18}O$ concentration in atmospheric precipitation (Fig. 8A) were not observed. However, two different days can be analysed : 8 and 12 September. Under the same daily air temperature (4.1 and 4.3 °C) the values of $\delta^{18}O$ concentration are great different (correspondingly $-36.2‰$ and $-11.9‰$). During most of this time the predominant wind direction, including the 12 September, was southeast, *i. e.* intrusion of the Indian monsoon from the lower Central part of the Himalayas. On 8 of September only the wind direction was southwest and the value of $\delta^{18}O$ was greatly reduced, indicated in increase of distance from source of moisture forming. Therefore, we hypothesise that the air masses bringing snowfall on 8 September was part of the west streams in the upper troposphere.

Thus the first results of our analyses gave the possibility to suggest that within the Eurasian continent the air masses developing over the Atlantic advances further, and at higher altitudes, than those from the Pacific and Indian Oceans. This effect is partly responsible for the existence of local glaciation on the Central Tibetan plateau higher than 6000 m.

The annual distribution of precipitation is very different on the northern and southern slopes of Himalayas as well as the Southeast Tibet (Fig. 2B). On the southern slope of the Himalayas the quantity of precipitation is three times more than its correspond average value on the northern slope (Fig. 2B, Table 2). On the southern slope of Himalayas, open to summer monsoon influence, two maximum in its annual distribution are observed (Fig. 2B). The first takes place from July till October end. The precipitation is the brought mainly by the southwest monsoon. The second maximum is observed in February-March when the precipitation is associated with activity of subtropical jet stream. According to Ramage (1976) the precipitation caused by west cyclones reaches a maximum in March, and weakens in April and May with weakening of the jet stream and west cyclones. Thus on the southern slopes of Himalayas the precipitation falls about the whole year with a minimum in November, when it drops to 11 mm.

On the northern investigated region of Himalayas the precipitation falls in summer from July till September with a maximum in August. The main moisture content air masses come here through lowering in central part of Himalayas. According to Sino-Nepalese Investigations (1988) the precipitation increase on the southern slope is observed till 2500 m (Fig. 4B), according to Yasunari and Inoue (1978) to 4000 m, and higher the precipitation does not change or even decreases. This is evidence about the altitudinal limits of the southwest monsoon development. The precipitation increase on the northern slope of Himalayas from 4000 m to 6000 m is the result of moisture content in air masses coming through the lower in central part of the Himalayas. The precipitation in May and September in this region could be connected with the subtropical jet stream. In summer the precipitation decrease is caused by intensive development of the tropical monsoon. Thus arrival of Atlantic moisture is more probable in winter, preceding and following monsoon seasons (May, September), especially on the southern Himalayan slope. In summer the main precipitation comes from the Bay of Bengal. However, local collection of moisture is possible because of development of strong evaporation processes. As we suppose, at elevations higher 5500 m the possibility of precipitation forming from the moisture, brought by general west transfer from the Atlantic, increases.

The altitudinal gradient of precipitation equalled $\gamma(P)=19$ mm per 100 m, recorded by us through simultaneously measured data at Tingri station, and at 5700 m on Xixibangma glacier in the course of expeditionary observations. It corresponds with the value represented in the work of Sino-Nepalese Investigations (1988). The proportion of liquid precipitation in the total was calculated using its variation curves derived from average air temperature (Fig. 5B). These curves were based on observation years on Golubin glacier station, north Tien Shan (Fig. 5A) and differ insignificantly from the corresponding data on the precipitation distribution on the northern slope of Himalayas and Southeast Tibet. The average value of liquid precipitation at 5900 m equals 10% of the July quantity, *i. e.* 8 mm. According to the curves all other precipitation (550 mm) above 5900 m falls as solid.

On the Southeast Tibet the major of precipitation falls during the summer monsoon (Fig. 2B). The spring averages 14–16 days of precipitation. In summer this increases to 17 to 20 days, and occurs as rain showers (long-term data of meteorological stations). In autumn, there are also up to 20 days per month with precipitation. The largest amount of precipitation is observed in June, but a lesser maximum occurs in September. On the Southeast Tibet there is no second winter maximum, the same as observed on southern slope of Himalayas. More strong influence of Tibetan anticyclone and intensive development of Northeast passats over Southeast Asia than over India is the reason of such annual precipitation distribution.

We observed that in diurnal variation of summer monsoon precipitation, the maximum occurs during the afternoon and at night, due to the motion of air masses upwards from the valley throughout the day. It is also important to note that the high-mountain chains of south-eastern Tibet are beyond the zone of maximum precipitation. Therefore, depending on the humidity and the thermal stratification of incoming air masses, the greatest amount of precipitation falls on the windward slopes of peripheral ranges, even before it reaches a summit level (Weischet, 1965). We collected the field data (Aizen and Aizen, 1993) which shows that during the summer-autumn monsoon, the precipitation increases up to an elevation of about 3600 m, and above this level the gradient diminishes. The average altitudinal gradient of annual precipitation calculated on the basis of Chinese meteorological records equals 329 mm per 100 m. The gradient value above 3000 m amounts to 4.9 mm per 100 m (Aizen and Aizen, 1993). The average quantity of precipitation at the elevation of the equilibrium line position (5200 m) was calculated and amounted 2207 mm. The share of liquid precipitation is 23% according to the mean of air temperature and precipitation distribution data (Fig. 5B). Thus the long term quantity of solid yearly precipitation is 1699 mm here.

The average gradient of air temperature, depending on the height, is 0.56 °C for 100 m in Southeast Tibet. The high gradient of air temperature on the Gongga glaciation is associated with continuous advection of warm air masses from the valley to the low levels of the glaciers as well as with the cooling influence of the large nourishment area. As well as this intensive advance, condensation processes are developing especially in the ablation zone (in heat balance its value could reach 20%, Table 2). Taking into account that here advection is a significant process in melt (from 35 till 58%, Table 2) air temperature preferably as the entrance predictor of runoff calculation (Aizen and Aizen, 1993). An average value of the air temperature gradient over the glacier surface equalled 0.55 °C per 100 m during the wet period and 0.34 °C in the dry one was recorded during field work on northern slope of Himalayas.

The principal component which influences the intensity of snow-ice melt and the formation of glacial runoff, is a degree of climate continentally of the mountain regions. Therefore in regions with continental climate it would be preferable to calculate the rate of melting by using solar radiation parameters. In regions with clear monsoon type of climate the calculation of snow-ice melt and runoff is best using air temperature data.

Glacial melt on the South East Tibet takes place mainly from advection during the summer monsoon (Table 2). Therefore, the interaction between the diurnal melt and the diurnal air temperature is rather noticeable here equation (7).

Southeast Tibet

$$W_i = \begin{cases} 8.3 \cdot T - 19.0 & \text{when } T > 2.2\,^{\circ}\mathrm{C} \\ 0 & \text{when } T < 2.2\,^{\circ}\mathrm{C} \end{cases} \tag{7}$$

On the glaciers of northern slopes of the Himalayas the solar radiation maintaines melting processes (up to 86%, Table 2). The insignificant role of turbulent heat exchange is a result of continuos katabatic winds destroying the turbulent fluxes. The process of melting is distinguished by features characteristic for the high mountain glaciers of inner Tibet. For North slope of Himalayas we have calculated the melting using two types of empirical equations (8, 9).

North slope of Himalayas

$$W_i = \begin{cases} -7.5 - 1.4 \cdot [Q(1-A)] & (8) \\ -1.52 + 3.2 \cdot T \ \text{when } T > 0.5\,^{\circ}\mathrm{C} & \\ 0 \qquad \text{when } T < 0.5\,^{\circ}\mathrm{C} & (9) \end{cases}$$

On the North Himalayas for estimations evaporated and condensed mass at elevations of 5700 m and 5900 m, the gradient measurements and weighing method were employed during field work 1991. At an elevation of 5700 m, in the course of melting period in daytime, the evaporation prevailed over the condensation occurring during the night. On average the daily value of evaporation was 0.72 mm a day. Maximum values recorded during cloudless weather reached 3.1 mm day^{-1}. The measurements at an elevation of 5900 m give us a mean of sublimation of about 0.02 mm in day. For estimation of total annual evaporation estimated as 66 mm, the daily mean of evaporation (0.72 mm day^{-1}) was multiplied on the duration of the melting period.

The calculated long term yearly precipitation at the elevation of the equilibrium line (5900 m) is 558 mm and its solid quantity amounts 550 mm (Table 2). The long term value of mass balance index for Xixibangma glacier amounted -3.4 gcm^{-2} on the basis of the yearly value of melting (598 mm) calculated by radiation, as more representative for this region, the mean of yearly evaporation and the value of ice formation from melting water (8.0 gcm^{-2}).

The total ablation value during the summer monsoon has been calculated for June it is 550 mm, July 929 mm, August 568 mm and 2050 mm per year. The contribution by condensation is not large, accounting for only 4% for the total. The value of ice forming-according calculations by Golubev (1976) for elevation of the equilibrium line position (5200 m) is -25.0 gcm^{-2}. Tabl 6 gives the summarised results of the calculation of annual mass balance index (I_{bi}) and the level of zero balance ($Hi_{e.L}$) Hailougou glacier over 39 years (1951/52 to 1989/90) whereas Fig. 7 shows the chronoqogical fluctuations of the mass balance of the glacier, which as a total has a negative value (-9.9 gcm^{-2}). The altitude difference between $H_{e.L}$ and $Hi_{e.L}$ reaches 450 m in some years. For the calculation of annual elevations of the equilibrium line position ($Hi_{e.L}$) the equation (10) is used :

$$Hi_{e.L} = H_{e.L} - I_{bi}(H_{e.L}) / \gamma(I_b) \tag{10}$$

where $I_{bi}(H_{e.L})$ is annual mass balance index at elevation of long term average equilibrium line position, calculated by the equation (6). Supposing that long term average mass balance index at $H_{e.L}$ has zero value the mean of altitudinal mass balance index gradient $\gamma(I_b)$ is calculated as $I_b(H_o)/(H_o - H_{e.L})$.

Table 6. The main indexes of mass balance Hailougou glacier.

Years	$Hi_{e.l.}$ m. a.s.l.	A_{cc} g cm^{-2}	W g cm^{-2}	$B_{e.l.}$ g cm^{-2}
1952	5248	193.9	207.0	-13.1
1953	5517	196.3	283.4	-87.1
1954	5127	211.8	191.7	20.1
1955	5104	187.5	161.2	26.3
1956	5231	206.0	214.6	-8.6
1957	4780	292.1	176.5	37.7
1958	5371	183.0	229.9	-46.9
1959	5303	209.2	237.5	-28.3
1960	5149	214.0	199.3	14.7
1961	5647	175.8	298.6	-122.8
1962	5158	195.6	184.0	11.6
1963	5236	181.9	191.7	-9.8
1964	5192	193.9	191.7	2.2
1965	4839	207.0	107.7	99.3
1966	5265	189.1	207.0	-17.9
1967	5465	187.5	260.4	-72.9
1968	5035	198.9	153.5	-45.4
1969	5137	186.0	168.8	17.2
1970	5284	183.9	207.0	-23.1
1971	5429	182.1	245.2	-63.1
1972	5498	186.1	268.1	-82.0
1973	5384	179.3	229.9	-50.6
1974	4910	202.7	123.0	79.7
1975	5305	178.1	207.0	-28.9
1976	5152	197.3	184.0	13.3
1977	5133	202.3	184.0	18.3
1978	5332	201.1	237.5	-36.4
1979	5288	198.1	222.2	-24.1
1980	5097	197.2	168.8	28.4
1981	5496	179.1	260.4	-81.3
1982	5026	193.8	145.9	47.9
1983	5119	183.9	161.2	22.4
1984	5333	201.0	237.5	-36.6
1985	5174	206.0	199.4	7.1
1986	5255	191.8	207.0	-15.2
1987	5083	208.8	176.5	32.3
1988	5405	196.4	252.8	-56.4
1989	5172	199.3	191.7	7.6
1990	5244	214.9	226.9	-12.0
ave.	5222	195.2	205.1	-9.9
	178	18.9	42.3	48.9

$Hi_{e.l.}$ is annual zero balance level; A_{cc} is average-weigh annual value of accumulation on the elevation of equilibrium line position (5200 m); W is average-weigh annual value of ablation on the elevation of equilibrium line position (5200 m); $B_{e.l.}$ is average-weigh value of the annual mass balance at the elevation of equilibrium line position (5200 m).

4. Conclusions

It was established that precipitation decrease from west to east and from north to south the increase with altitude, and displacement of their time maximum from winter to summer, depend on interaction between Siberian anticyclone and Western cyclone air masses during the year. The Atlantic, Indian and Pacific Oceans are the moisture sources on the Eurasian continent as demonstrated with the aid of glaciometeorological and isotopic measurements (^{18}O and D) of the atmospheric water at different altitudes of the Central Tien Shan, S-E Tibet and N. Himalayas glaciers. In the regions with monsoon type of alimentation (the Tibet, Himalayas) we have found moisture from all three oceans, and on the Tien Shan the Atlantic is the only moisture source.

It was established, by investigations on North Tien Shan, that the changes in climate parameters are synchronous with those in mass budget. The glacier tongue was retreating and small shifts of the glacier front did not correspond with the phase of mass budget fluctuations. The decrease in precipitation quantity on the West and North Tien Shan periphery is observed simultaneously with tendency of precipitation growth in Central Tien Shan.

The existence of glaciers in this high mountain regions of South periphery of Central Asia is determined by the intensity of monsoon development. These include the decrease of insolation intensity (on the S-E Tibet approximately six-folds) as a result of reflection and absorption because of continuous cloudiness, and high values of albedo after summer snowfalls (on the northern slope of Himalayas not less than 58%).

The glaciation systems represented in this discussion have a negative mass balance on the average, which testifies to the general trend towards degradation of glaciation of Central Asia. The highest values of negative balance are typical for the glacier systems on Northern periphery of Central Asia, especially for the glaciers of the Central Pamir, whose glaciation conditions are determined, to a considerable extent, by the influence of Atlantic air mass. It was shown that the outlying mountain areas of Central Asia respond first to variations in the general circulation of atmosphere.

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