The Assessment of Snow Accumulation, Precipitation and Runoff over the Karakoram Glacier System from Satellite Images

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Many applied and fundamental scientific tasks related to water and energy exchange in high mountain areas have been solved only on the level of glacierisation considered as a whole. Where direct measurements are absent or non-representative (Krenke 1975; 1982), the isopleths of snow-line height on glaciers, combined with information about summer air temperature, have been used as a tool to estimate snow accumulation on the glaciers, glacier runoff, and even total precipitation and runoff in the high mountainous areas. The main problem is collecting snow-line data.

We estimated the position of the snow line over the Karakoram Mountains from images from the Russian satellite Resource - F1- taken in August 1980. Cameras of the KFA-200 (resolution 20m) and KFA-1000 (still better resolution) types were used. The orography of the Karakoram Range, and the boundaries and morphological types of 3,590 glaciers could be defined more accurately on the basis of the transformed images. This provided the possibility of estimating the total area and volume of glaciation in the Karakoram. The total glacier area was put at 12,130sq.km. For volume estimation, the glaciers were divided into three groups: 1) dendritic and complex valley glaciers with lengths of more than 15km: 2) complex and simple valley glaciers with lengths between five and 15km; 3) the simple valley, cirque, and slope glaciers shorter than five km. From the experience of glacier thickness measurements in the Tien-Shan, Pamirs, and on the Batura Glacier in the Karakoram Mountains (Professional papers 1980), the average thickness of the first group was assumed to be equal to 300m, the second one - 70m, and the third - 35m. As a result the total volume was estimated to be 2,160 cubic kilometres. This corresponds approximately to 1,900 cubic kilometres of water storage in Karakoram glaciers.

To assess the accumulation on the glaciers, on the basis of its equality to ablation at the equilibrium line, the firn line on the 743 glaciers was determined from the images. Most of these glaciers are valley glaciers. The difference between the firn line and equilibrium line, which can occur because of superimposed ice below the firn line in cold, dry areas or because of surviving firn below the equilibrium line in warm, wet ones, was neglected, such differences being never more than 150m. This assumption of the identity of equilibrium and firn lines leads to some overestimating of accumulation in the wet and underestimating in the dry areas. Still, these errors are no more than 10 per cent. The firn line is easily recognisable on the images in the visible bands because of the higher firn albedo and brightness compared to the wet melted ice covered by dust and coarser mineral particles.

The altitude of the firn line on the glaciers was estimated using analytical photo-triangulation on overlapping images, giving due consideration to the satellite trajectory. River mouths were used as basis points. Their altitude could be estimated to within 10m from good topographic maps. The resulting error in the estimation of firn line height from the space images appears to be no less than 55m. A map of firn line heights was plotted for the whole Karakoram Range (Fig. 1), with the intervals between isolines equal to 200m. The interpolation between the 743 points was subjectively carried out on the basis of topographical features. Before the interpolation, in cases of significant local variability in firn line height between adjacent glaciers, a smoothing process was carried out by means of averaging.

The firn line on the glaciers on the south-western slope of the Racaposhi Massif is the lowest (4,200masl); the highest one (6,000masl) is in the north-east, on the Central Rimo Glacier. Along the western and south-western slopes of the Karambar, Rakaposhi, Kudjut, and Masnerbrom ranges, the snow-line decreases very rapidly from north-east to south-west. The isolines become very close to each other. The firn line is relatively low along the valleys of the Karambar, Hunza, Indus, and Shyok rivers that open to the south-west. This permits us to conclude that the main precipitation over the Karakoram Range is generated by the Mediterranian cyclones coming from the west. The influence of monsoon precipitation is reflected in isolines features only in the eastern Karakoram Range, but even there they play a secondary role.

At the equilibrium line height, by definition the accumulation is equal to ablation. This latter, and thus both parameters, we estimated from the summer air temperature. To determine this temperature at the equilibrium line altitude (ELA), we plotted a summer air temparature map (Fig. 1) for an altitude of 3,500masl which is close to the glacierised zone and corresponds to the altitude of the highest meteorological station in the area. We managed with only six stations. The corresponding data are given in Table 1. The period over which averaging was carried out ranges from 60 to 70 years. No significant warming after 1960 was found in Central Asia, in contrast to global temperatures. Thus we could use this period as conformable to the needs of the above described ELA estimation. The average long-term June-August air temperatures were estimated for each station and then adjusted to the common height of 3,500masl using the vertical air temperature gradient 0.7°C per 100m taken for the nearby Pamirs. This assumption could be considered only as a first approximation. All Karakoram stations are too far from each other in the horizontal direction to be used in vertical gradient calculations.

The summer air temperature at a fixed height increase in the inner part of the mountains, from south-west to north-east, varies from 12.5 to 16.5°C. The horizontal air temperature gradient is about 2.4°C per 100km, five times as large as the south-north gradient over the whole Asian continent. This reflects the increased continentality towards Tibet and the increase of the mountain's heating effect with the increase of range height up to 8,000masl in the Central Karakoram Range. The isotherms follow the directions of the Karakoram and Kunlun ranges surrounding the dry northwestern Tibetan provinces.

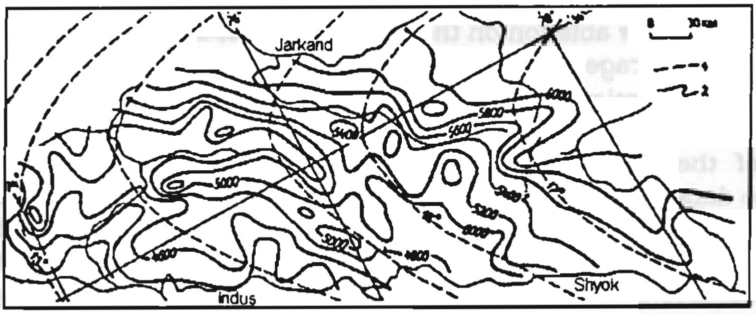

Figure 1: The Height of the Firn Line on Glaciers and the Summer Air Temperature at a Height of 3,500masl

1 - Summer isotherms at a height of 3,500masl in °C.;

2 – Isolines in masl

Table 1. Meteorological Stations in the Area

Station	Altitude masl	Coordinates		Average summer air temperature °C	Data collection period
		N. lat	E. long.		
Dras	3,068	34°26′	75°46′	15.9	1896-1940
Leh	3,541	34°10′	77°40′	16.2	1882-1960
Skardu	2,288	35°17′	75°45′	22.4	1897-1947
Gilgit	1,290	35°48′	74°00′	27.5	1893-1947
Srinagar	1,586	34°05′	74°50′	23.0	1893-1960
Murri	2,186	33°55′	73°23′	19.8	1881-1940

The air temperature at the ELA was estimated by superimposing the summer air temperature map over the firn-line height map. The summer temperature was estimated for every grid point and for every big glacier. The same vertical air temperature gradient was used. The cooling effect of the glacier ice was taken into account by introducing the marginal temperature jump, which depends on the size of the glacier (Khodakov 1978):

$$\lg Ti = 0.28 \lg Xo - 0.07 \qquad (1)$$

where,
Ti is the air temperature jump between the average summer temperatures at a height of 2m over the rock and ice surfaces at the same altitude above sea level, and Xo - is the length of the glacier.

The jump is equal to 1.0°C for glaciers shorter than five km; 1.5°C for glaciers having a length of between five and 15km; 2.0°C for glacier lengths between 15 and 30km, and 2.5°C for glaciers longer than 30km. The size distribution of the glaciers with their estimated firn-line heights in each cell of the grid was taken into account.

For each grid point, the total yearly ablation was estimated according to a previously published formulae (Krenke 1982):

$$A = 1.33(Ts + 9.66)^{2.85} \qquad (2)$$

where,
A is the yearly ablation on the glacier surface, and
Ts is the average summer air temperature.
The accumulation is assumed to be equal to ablation.

The map of the accumulation at the ELA was plotted for the whole area according to data for each grid point (Fig. 2). The interval between the isolines is equal to 500mm, which is approximately two times more than the errors from the above-described calculations. The theoretical error after the addition of all the particular errors under the square root is in the range of 16-31 per cent in relative and of 150-400mm in absolute figures. The experimental error was estimated only on the basis of the field measurements on the Batura Muztagh Glacier (Hunza River basin) carried out by Chinese scientists in 1974-1975. Their observations are in the range of from 1,030-1,250mm; ours in the range of from 1,000-1,450mm. Thus experimental error is in the range of from -3 to +15 per cent.

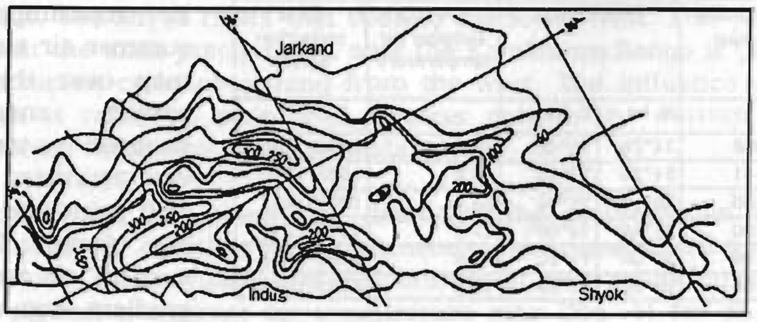

Figure 2: Snow Accumulation at the Height of the Firn Line on Glaciers (cm of water equivalent)

The accumulation decreases from 3,500mm in the region of the Rakaposhi Massif to less than 500mm in the vicinity of the Tibetan border. Typical features of the accumulation map are the isoline loops reflecting the interaction of air streams with the complex high mountain topography. The existence of such streams is confirmed too by the vegetation patterns on the images.

The accumulation patterns reflect the precipitation ones. However, they differ quantitatively because of avalanches and snow-drift concentration over the glacier. Some of the precipitation is liquid and not included in our approximative calculation. In order to estimate the solid precipitation, Khodakov (1978) suggested dividing accumulation by the concentration coefficient, equal to the relationship of the glacier basin area to the glacier area itself.

However, not all the slopes in the glacier basin are free of snow. A significant part of the snow on slopes melts or evaporates in places not affected by avalanches or snowstorms. We have modified the concentration coefficient using the satellite image data. We have estimated that in the vicinity of the firn line this coefficient could be quantified as

$$Cc = 0.8\ (1 + Sf/Sg) \qquad (3)$$

where,

Cc is the concentration coefficient,
Sf is the length of the cross-section of slopes free of snow at the ELA, and
Sg is the length of the cross-section of the glacier at the ELA.

The solid precipitation map was plotted according to the accumulation data corrected by the concentration coefficient. The interval between the isolines is again equal to 500mm (Fig. 3).

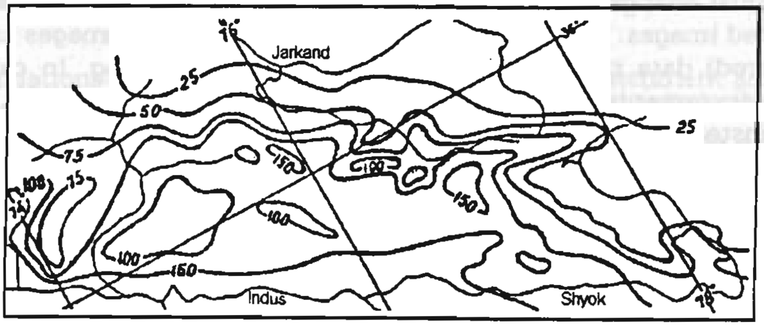

Figure 3: Precipitation at the Height of the Firn Line on Glaciers (in cm)

The total precipitation is equal to the sum of solid and liquid precipitation. The share of the liquid precipitation decreases very rapidly with altitude. Nowhere in the Karakoram Range is the ELA below 4,000m, and the monthly temperature is never below +2°C. Most precipitation occurs in the early spring, not during the warmest time. Due to the seasonal distribution of precipitation and the dependence of the ratio of liquid to solid precipitation on air temperature, the amount of liquid precipitation at the height of the firn line in the Karakoram Range is negligible. Thus, the solid precipitation map is valid for the total precipitation too. The precipitation decreases from south-west to north-east, from 2,000mm on the southern slopes of the Karambar Range to 200-300mm in Tibet.

The glacier runoff can be calculated from the ablation at the snow-line. The average ablation from the total glacier surface in the first approximation (where there is a linear dependence on the altitude) could be assumed to be equal to the ablation at the snow-line. The ablation consists of melting and evaporation. The latter's share in the total ablation, due to condensation in the low parts of the glaciers, hardly exceeds two to three per cent, though its role increases with altitude. According to H. Untersteiner, evaporation on the Chogolungma Glacier is equal to five per cent of the total ablation (Konovalova 1972). The role of evaporation increases to the north of the rivers Sheksgam and Shyok. By analogy with the Pamirs, it can be estimated at 100-150mm. Subtracting this amount from the total ablation and multiplying the result by the glaciers' surface area we arrive at the glacier runoff. The glacier runoff from the whole Karakoram glaciation is assessed to be as much as 17.1 cubic kilometres. Of this 14.3 cubic kilometres is directed to the Indus River and only 2.8 cubic kilometres to the Jarkand River. The average period of mass turnover (water storage divided by ablation) is about 120 years, compared to 80 years for the Pamirs and 100 years for the Tian-Shan. The total runoff from the high mountain belt could be estimated from the total precipitation multiplied by the runoff coefficient, equal to about 0.85.

Future investigations need to include a more accurate runoff assessment for separate years, based on the upper limit of the snowline, and to develop models of the seasonal change of glacier runoff using the transient snow-line positions from repeated images. This will require a better resolution of images and the use of IR (Infrared) data to determine the upper limit of melting. In cases where ground measurements are possible, regional empirical formulae could be developed instead of the global ones we have used in this presentation.

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