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STUDY OF GLACIER MELT AND PHYSICS OF GLACIERS

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ABSTRACT

The existence of a Perennial ice mass depends on the interplay of accumulation and ablation processes. Over a time scale of a year or more on a glacier, accumulation processes dominate in the upper reaches and ablation processes dominate in the lower reaches. The terrain factors play an important role in accumulation, ablation and on the time taken for meltwater to travel through the basin. The debris deposits, their extent and thickness also affect the melting of glaciers.

Glaciers act as natural reservoirs, storing precipitation in the winter and releasing it in summer. Glaciers also store water for periods of years, retaining part of the annual precipitation during periods of their growth and releasing extra-water from storage during periods of recession. The routing of meltwater through the glacier system has been a major problem in glacier melt modelling studies. The meltwater enters the watershed storage system which functions in a manner analogous to a reservoir and displaces water in storage which appears subsequently in the channel system. The input varies with change in the rate of ablation and liquid precipitation.

Meso to Macro scale modelling with numerical iteration of the energy and mass fluxes may bring progress in this field. The much improved network of hydrometeorological stations in the inaccessible regions of snow and ice is essential to produce the necessary input data for these large-scale models. The hydrological conditions of

glacier are hard to reveal because of difficulty of organizing longterm observations and big spatial variability of conditions.

1.0 INTRODUCTION

Himalaya constitutes the largest reservior of snow and ice outside the polar regions and supports a multitude of glaciers. Himalayan glaciation is more intense as compared to the Alps and Rockies. It is mainly due to their ultra high altitudes which compensate for its location at low latitudes. More than 15,000 glaciers covering an area of over 50,000 Km drain into the Indian river systems. The principal Himalayan river systems which receives glacier melt contributions with their approximate-mountainous area, glacier area and percentage glaciated catchment are listed in Table 1.

It is estimated that 30-50% of the total annual runoff of almost all the major rivers of India originating in Himalaya is provided by the snow and glacier melt runoff. Reliable estimates of the volume of water contained in the snow and ice pack and the rate of release of the water, are needed for efficient management of water resources which includes flood forecasting, reservoir operation and design of hydraulic structures etc.

Melting of glaciers exposed to atmosphere is greatly influenced by the atmosphere and existing surface conditions. The estimation of the rate of glacier melting requires knowledge of many meteorological, hydrological and surface parameters. Towards the understanding of hydrology of glaciers limited work has been carried out in our country. A comprehensive review of physics of glaciers and their melting has been made in this report.

TABLE - 1

PRINCIPAL GLACIER-FED RIVER SYSTEM OF HIMALAYA(BAHADUR,1986)

No.	Name of River	Major River system	Mountain Area (Km^2)	Glacier Area (Km^2)	Percentage glaciation
1.	INDUS	INDUS	268,842	8790	3.3
2.	JHELUM		33,670	170	5.0
3.	CHENAB		27,195	2944	10.0
4.	RAVI		8,029	206	2.5
5.	SUTLEJ		47,915	1295	2.7
6.	BEAS		14,504	638	4.4
7.	JAMUNA	GANGA	11,655	125	1.1
8.	GANGA		23,051	2312	10.0
9.	RAMGANGA		6,734	3	0.4
10.	KALI		16,317	997	6.1
11.	KARNALI		53,354	1543	2.9
12.	GANDAK		37,814	1845	4.9
13.	KOSI		61,317	1318	2.1
14.	TISTA	BRAHMAPUTRA	12,432	495	4.0
15.	RAIDAK		26,418	195	0.7
16.	MONAS		31,080	528	1.7
17.	SUBANSIRI		18,130	725	4.0
18.	BRAHMAPUTRA		256,928	1080	0.4
19.	DIBANG		12,950	90	0.7
20.	LUHIT		20,720	425	2.1
		TOTAL :	1,001,294	25724	2.6

2.0 REVIEW

2.1 Formation of Glaciers and Their Classification

2.1.1 Glacier sources

Permanent snow and ice is located wherever topographic and climatic factors are suitable for snow to deposite and survive. Glacier ice is derived indirectly from the precipitation of snow or ice crystals from the atmosphere or directly from liquid transformed to ice at the surface. Once on the surface, the characteristics of the snow vary considerably, according to the environment of depsition. In mountainous areas avalanches may be an important means whereby fresh snow is moved from slopes on either side of a glacier to its surface (Vivian, 1975).

Water vapours which freezes on contact with glacier surface forms several types of ice, the most important of which is rime. Rime ice is formed when supercooled water droplets strike a cold solid object and freeze on impact. This ice is whitish in appearance as a result of entrapped air bubbles and is quite firmly attached to the receiving surfaces. Rime accumulates most rapidly in cool and humid conditions on surfaces which are most exposed to the wind (Mellor, 1964).

Superimposed ice is formed when water comes into contact with a cold glacier surface and freezes. The water may be derived from rain or more commonly from melt water from the summer melting of the pervious winter's snowcover. Superimposed ice can only be formed when air temperatures are at or above freezing point.

2.1.2 Areal Extent of Glaciers

At present the aggregate area of the world's glacier is about 14.9 million Km^2, which is about 10 percent of the world's land area.

Of this, about 12.5 million Km^2 is accounted for by the Antartic ice sheet and 1.7 million Km by the Green-land ice sheet (Sugden and John 1976). The remaining 700000 Km^2 of glacier ice is distributed among the other glacierized areas; like ice caps which rarely exceed 10,000 Km^2 in extent mostly in high latitudes, and in several thousands of small glaciers in the upland areas of the world (Table 2). However, more and more details are now forthcoming on the extent of glaciers within specific countries and national and regional inventories and maps are now published, partly as a result of the interest shown in glacier hydrology during the International Hydrological Decade (Lorenzo, 1959; Prest et al, 1968; Meier and Post, 1969; Ommaney, 1969; Ostrem et al, 1973).

It is evident from the Table 2 that, apart from the two ice sheets, glacierization is concentrated in the northern hemisphere, for the most part on the islands of the North Polar basin and on the uplands of the Oceanic peripheries (e.g. Alaska and Scandinavia). Other high lands of the middle and low latitudes, such as Alps, Karakoram and Himalayan ranges, have appreciably ice covered areas. The areal extent of ice on the African continent is negligible. Thus present day ice cover is essentially discountinous and is in no sense balanced between the two hemispheres or between the major landmasses (Hatlersley and Smith, 1974; Ostrem, 1974b).

2.1.3 Glacier Types

The three main glacier types are distinguished by fundamental differences in the way their morphological expression reflects the interaction between glacier ice and topography (Ostrem, 1974b).

Table 2 : World Glacier Extent (Sudgen & John, 1976)

Region	Area (Km^2- approximate)	Sub-totals
South Polar region		
Antarctic ice sheet(excluding shelves)	12,535,000	
Other Antarctic glaciers	50,000	
Subantarctic islands	3,000	
North Polar regions		12,588,000
Greenland ice sheet	1,726,400	
Other Greenland glaciers	76,200	
Canadian Arctic archipelago	153,169	
Iceland	12,173	
Spitsbergen and Nordaustlandet	58,016	
Other Arctic islands	55,658	
North American continent		2,081,616
Alaska	51,476	
Other	25,404	
South American cordillera		76,880
European continent		26,500
Scandinavia	3,810	
Alps	3,600	
Caucasus	1,805	
Other	61	
Asian continent		9,276
Himalaya	33,200	
K'un Lun chains	16,700	
Karakoram and Ghujerab-Khunjerab ranges	16,000	
Other	49,121	
African continent		115,021
Pacific region (including		12
New Zealand		1,015
	Grand total	14,898,320

(a) Ice Sheets and Ice Caps

An ice sheet or ice cap is superimposed on the underlying topography which largely submerges the direction of flow on the ice reflects the size and shape of the glacier rather than the shape of the ground. The difference between an ice sheet and an ice cap is normally accepted as being one of the scale with the dividing line somewhere around 50,000 Km^2 (Armstrong, et al., 1973). Thus, an ice mass covering Antarctica, northern America or the British Isles would be termed an ice sheet, whereas an ice mass over Wales, the Grampian mountains of scotland would be termed an ice cap. The main feature of ice sheet is that highest temperature occurs at the base. In addition, shear-stress is greatest at the base.

These form where precipitation input to land scape system is too great to be dealt with by glacial discharge trough throughs, so that output can not compensate. In other words, precipitation is highly effective once the threshold has been crossed and glacier is growing. The glacier system, once initiated, overflows existing topographic irreqularities and acts as a greater and greater reservoir.

An ice domes builds up so that it is situated approximately symmetrically over the land area involved. Some times, as in the east Antarctica, the summits of ice domes may exceed 4,200 m in altitude and lie over topographic rises. The convex surface slope of an ice dome forms in response to the basic flow characteristics of ice. Assuming that there is adequate snow, ice mill build up until the shear stresses are sufficient to deform the ice efficiently. Since shear stresses are influenced by a combination of ice thickness and surface slope, the thinner the ice the steeper the surface slope needs to be/to maintain flow. Conversley, the thicker the ice the less the surface slope needs to be. For these reasons, the surface of the ice dome is gently sloping in the centre where the ice is thickest, and steepens progressively as the ice becomes thinner towards the margin. Manyy attempts have been made to describe the profile of such dome surfaces. This has usually been achieved by assuming that ice behaves as a perfect plastic material, and by comparing a theoretical curve with reality, in some cases parabola fits well (Nye, 1952a), while in other cases more complicated curves are required (Weertman, 1961; Haefeli, 1961). Nye's solution is particularly useful as an first approximation because of its simplicity. On a horizontal bed the altitude of the ice surface at any given point inland from a known margin can be found from the formula

$$h = \sqrt{2 h_o s}$$

where h = ice altitude in m, h_o = 11m, and S is the horizontal distance from the margin in meters. From camparison with other theoretical profiles and real profiles, it seems that Nye's parabola slightly overestimates the slope (and thus the altitude) near the centre of an ice sheet.

(b) Ice shelves

An ice shelf is a floating ice cap or part of an ice sheet which deforms under its over weight. It can be regarded as a slab of ice being squeezed between two surfaces the atmosphere and the ocean. Ice shelves comprise unique models ideal for developing principles of ice creep, since basal friction can be largely ignored. Rates of ice movement may be between 0.8 and 2.6 Km per year (Swithinbank and Zumberg, 1965). Ice shelves are most common in the Antaractic where they comprise some

7% of the total ice area and 30% of the length of coastline. They require mean annual air temperature below zero, and relatively high precipitation. Relief, however, a significant control, for ice shelves require either broad coastal embayments or groups of island in order to reamin safely anchored.

The main morphological characteristics of Antaractic ice shelves are described by Swithinbank & Zumberg (1965). The surface of an ice shelf is approximately horizontal, but may have gentle undulations. Crevasses are found mainly near the margins or locally grounded areas. The landward margin can be indentified by the start of uphill slope of the inland ice sheet, and by "stand cracks" produced as the ice shelf rises and falls with tide. The seaward margin is a vertical cliff, normally about 30 m high. It is limited to this height by the plastic properties of ice, just as the depth of a crevasse is.

Ice shelves are nourished largely by snow accumulation on their flat upper surfaces, but there may be varying amounts of ice supplied from land glaciers and occasionally by bottom freezing. With the exception of small ice shelves, surface accumulations is highest near the seaward edge and decreases rapidly inland in a similar fashion to that on ice dome. Under certain circumstances bottom freezing may add ice to the bottom (Swithinbank, 1970). Ice wastage is by Calving and by bottom melting (Thomos and Coslett, 1970). Calving may produce massive tabular ice-bergs many tens of kilometers in length.

(c) Glaciers constrained by topography

The glaciers constrained by topography is strongly influenced both in its form and its direction of flow by the shape of the ground.

These are subdivided into component elements on the basis of morphology:

Ice fields

An icefield can be regarded as an approximately level area of ice which is distinguished from an ice cap because its surface does not achieve the characteristic domelike shape, and because flow is strongly influenced by the underlying topography. Icefields form wherever the topograhy is sufficiently high or gentle for ice to accumulate before going on to flow in restricted valley glaciers. A large example occurs in the Pacific mountains ranges of Western North America in the St. Elias mountains area. Smaller icefields occur in the less dissected parts of the World's mountain chains.

Valley glaciers

A Valley glacier flows in a rock valley and is over looked by rock cliffs. It may originate in an icefield or a circque. Such glaciers flow in valleys radiating from the main massif on which they form, and commonly display a dendritic pattern simpler than but similar to those if river valleys. Regardless of their different sizes, tributary glaciers tend to join the main glacier with their more or less conformable to one another. Exceptionally, valley glaciers may be 120 Km in length like the Hubbard glacier in North Western America, but lengths of 10-30 Km are more common. As is the case with rivers, there is a tendency for the size of a glacier to reflect its importance in the hierarchy of the drainage basin. A test on the 40 glaciers in the east Greenland revealed a close correlation between the width of valley glaciers and the number of tributaries, measured in this case as the number of cirque collecting grounds. It is common to find valley glaciers debounching from steep mountain valleys into adjacent low lands. Free from the constraints of the valley the glacier snouts broaden out to form piedmont lobes. Some prominent characteristics of valley glaciers are in figure 1 . The attitudinal range of a valley glacier in relation to its size and thus the bedrock slope is likely to be steep. In some situations, valley glaciers form where ice sheets and ice caps can not build because land surface gradients are too steep for an ice sheet profile to be maintained. Also, in most cases, net accumulation increases with above the equilibrium line. The valley sides are relatively ice free and serve to provide the glacier with a veneer of frost-shattered rock debris. Valley glaciers tend to form where precipitation is effective, as long as other variables also favour their growth (Miller, 1973).

Cirque glaciers

A circue glacier is a small ice mass generally wide in relation to its length and characteristically occupying an armchair-shaped bedrock hollow. The glacier may be confined to part or the whole of hollow or simply comprise the arctuate head of the valley glacier. In the latter case, cirques are well filled with ice and snow and the cirque part of the glacier may merge imperceptibly into the large valley glacier, as is common in areas of intense glacierization. Special features of cirque glaciers arise from the fact that much accumulation is obtained from drifting snow from surrounding slopes, neighbouring summits and plateaux. The mean annual wind velocities do not exceed 3 m/s. However, snow drift nourishment of glaciers is some times difficult to distinguish from avalanche nourishment. This makes cirque glaciers more vigorous than adjacent large glaciers at a similar altitude.

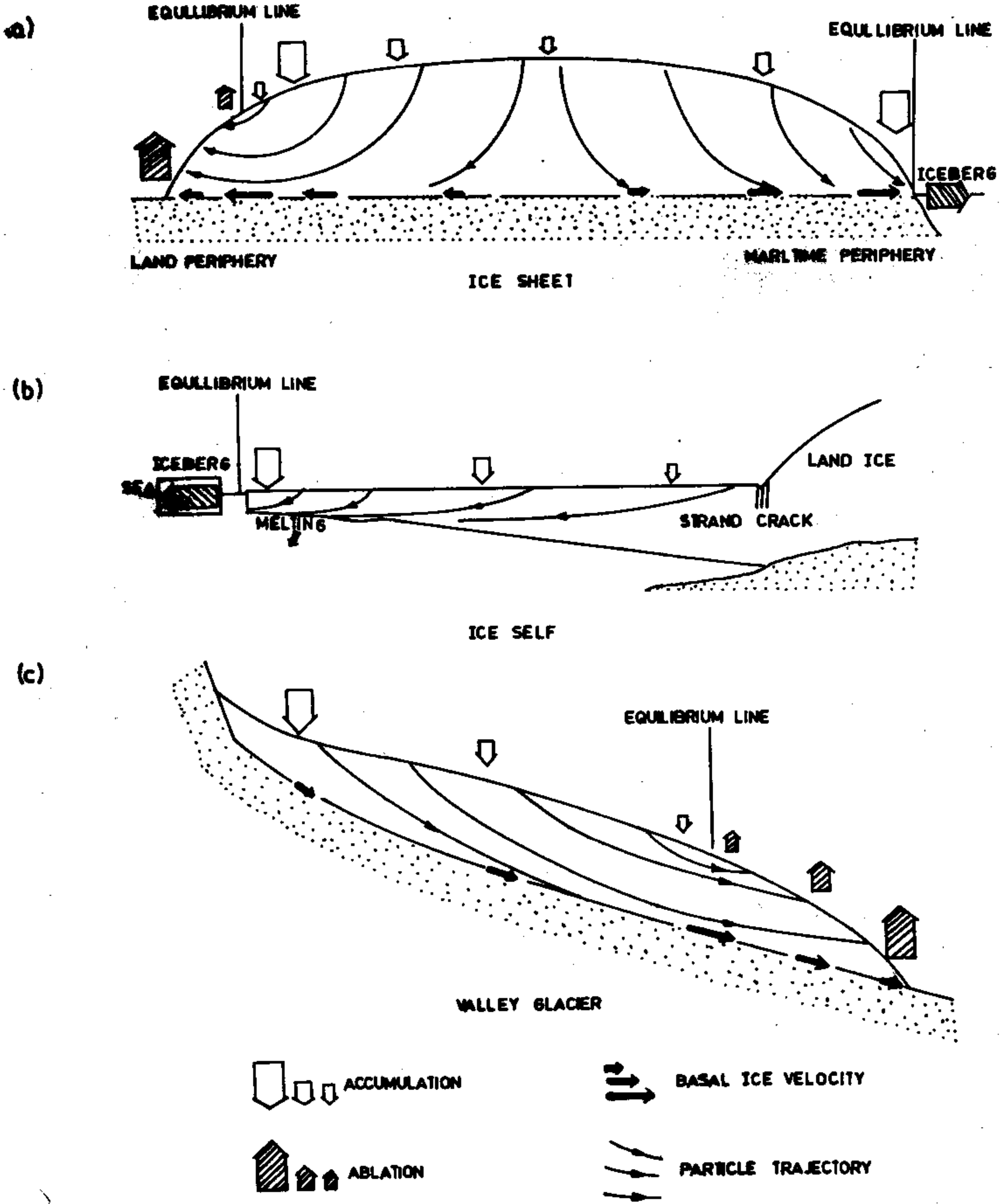

Fig.1: Model ot (a) an ice sheet, (b) an ice shelf and (c) a valley glacier showing distribution of snow input and output and related flow characteristics. Basal slipping is assumed to occur in models (a) and (c) and is at a maximum in the vicinity of the equilibrium line.

There are some other small glaciers found in mountains. Some glaciers cling pre-cariously to small hollows on steep valley sides. Others are thin masses of snow and ice adhering to mountain sides. Such glaciers have been discussed in detail in relation to the Alps by Galibert (1965) and Rothlisberger(1974).

2.1.4 Factors influencing the distribution of snow and ice

The following variables mainly affect the glacier distribution:

(a) Precipitation:

Precipitation is a potent control over glacierization (Wilson, 1970). A high altitude west coast envirenment in the northern hemisphere is often suitable for extremely heavy precipitation (more than 3000 mm water equivalent); however, much of this may be glaciologically ineffective if it falls as rainfall during the summer season or if prolonged periods of clear skies enable solar radiation to achieve higher melting rates at the snow or ice surface. Ostrem et al (1967), Adams (1966) and many other glaciologists have shown that except in high latitudes summer precipitation in the form of rainfall contributes little to glacier mass. Most of it remains unfrozen, and with the exception of that which is converted into superimposed ice, it does not enter the glacier system store in solid form; during rapid throughput either on or beneath the glacier surface it combines with melt water and is discharged at the snout.

The term "nivometric coefficient" has been coined as an index of snowfall effectiveness (Tricard, 1969), being the ratio snowfall (in water equivalent) to total annual precipitation. A coeff-icient of 1.0 implies a precipitation entirely of snow. Consideration of the nivometric coefficient and net annual precipitation together provide a rough guide to the likelihood of glacierization. The regions with nivometric coefficient approaching unity and high precipitation, provide the best conditions for nourishing glaciers. Examples are the high Alps (above 3000 m) and the mountains of South-West Greenland. Regions with nivometric coefficients approaching unity with low annual precipitation totals are less suitable for glacier growth but more suitable for prolonged glacier survival. The Antartic ice sheet is an example. Areas with a medium nivometric coefficient and high precipitation such as higher parts of some mid latitude martime mountain massifs are more marginal from the glacierization point of view, but many nevertheless contain extensive ice fields and vigorous valley glacier.

(b) Temperature:

A close relationship has been found between regional temperature characteristics and glacierization. There have been many attempts to generalize concerning the thermal requisites of glaciers, but few of them have been successful. Peltier (1950) defined one characteristic of his glacial morphogenetic regions as those experiencing a range of average annual taemperatures between-6.7^{o}C & -17.8^{o}C. However, many glaciers exist in environmental conditions outside of these limits, and Paterson (1969) stressed that glaciers can exist in unlikely of climate situations as a result of local pecularities of Glaciers exist in localities where mean annual air temperatures are subzero; in parts of Antarctica mean annual air temperatures may be below-50^{o}C(Loewe, 1970), whereas the mean is - 23 C at Camp Century, Greenland. Few glaciologists consider mean annual temperature as a significant climatic parameter as far as glaciers are concerned as long as air temperatures are below zero snow can accumulate, and as long as air temperature is above zero ablation will occur. Even the annual amplitude of temperature appears in significant, although fluctuations around freezing point can exert some control over glacier regimes (Loewe, 1970).

"Mean summer temperature" preferably measured for ablation season is of much greater importance from a glaciological point of view (Orvig, 1951). Glaciers can exist in an approximately steady state where the number of degree days is well above 1000, but only if accumulation ratio are high. To conceptualize, the lower the solar radiation, the greater are chances of glacier survival,with the over riding proviso that solid precipitation must be adequate for glacierization.

(c) Latitude

The world's high latitude areas are favourable for the existence of glaciers because of fundamental control exerted on solar radiation (Hattersley & Smith, 1974). These areas receive relatively low amounts of annual radiation and experience prolonged winters with more or less unbroken subzero temperatures. It is no coincidence that the North Polar Ocean basin is covered with pack ice and South Polar landmass is burried beneath a continental ice sheet.

As early as 1912 Paschinger showed how the regional snowline is influenced by latitude at a world scale. Also at a continental scale a good correlation between latitude and glacier altitude is expected. Meier (1960) has deomonstrated the relationship between mean galcier altitudes and latitude along the North American Cordillera and Hastenrath (1971) has shown how the modern snowline gradually falls about 2300 m along the crest of the South America Cordillera between latitudes 24° S and 33° S. However, at a regional scale it is less easy to demonstrate the effect of latutude on snowline or glacier distribution, and one has to search for the influence of other variables in combination.

(d) Altitude

Altitude exterts fundamental control over climatic parameters and hence on glacier distribution (Ostrem, 1974 b). Flint (1971) summarized the local influence of altitude thus : "No one who examines the present day distribution of glaciers can fail to realize that glaciers are related to highlands. Without high and extensive mountains some of them situated in the paths of moist winds, extensive glacierization can not occur". Drewry (1972) illustrated this point by reference to Antarctica, the glacierization of which was probably initiated by a phase of upland glaciation in the Transantarctic mountains.

Evans (1969) referred to the altitudinal zonation for mountainous areas as glacial/nival, sub nival, alpine and subalpine. It was pointed out that this is a middle and low latitude viewpoint, and it must not be forgotten that glacierization occurs down to sea level in high latitudes, but only a high altitude in low latitudes. For example, the only three glacierized mountain peaks of Mexico have summit altitudes above 5000m (Lorenzo, 1969). Overall, as pointed out by Klute (1921) and many others since, the regional snowline falls irregularly from the tropics towards poles, and this irrequalrity is again due to the interaction of many environmental variables. However, glacierization can not occur unless there are upland areas above certain ciritical altitudes, these altitudes varying partly in relation to latitude.

(e) Relief

Surface relief is an important morphological variable which exters its own specific control over glacierization at a number of different scales. At a continental scale, for example, an expanding Antarctic ice sheet would have its dimensions limited simply because beyond the edge of the continental slope the bed rock base is too far beneath sea level for the ice to remain in contact (Hollin, 1962). At more local scales, individual plateau glaciers and mountain valley glaciers have obvious links with topotgraphy which are indicated by the descriptive names used. Manley (1955) have shown that how the breadth of an individual summit determines whether or not a glacier can be supported, and this concept has also been used by a number of authors in evolving the idea of "glaciation level". Ostrem (1964,66) followed Ahlmann (1937) in using the Partsch-Bruckner method of defining this limit, recognizing the part played by relief in determining glacier distribution. In short, glaciers must have suitable topographic situations to exist.

In highly dissected alpine uplands, glaciers are restricted to the valleys; nourished largely by avalanches, such glaciers often reach lower altitudes than usual. Steep slopes often remain ice free, even well above the glacierization limit. Such conditions are typical of glaciers in the Pamirs and other mountainous areas of Central Asia. In the similar vein, Andrews et al (1970) have drawn attention to the "Snow fence" effect of the jagged mountain in trapping snow and allowing cirque glaciers to develop at lower than normal altitudes.

At the local or small scale, the control of relief is exerted in circumstances where microclimatic factors can induce the growth of perennial snowfields or ice masses. Topographic irreqularities, particularly hollow, on hill slopes may induce drifting snow to lodge and accumulate (John, 1974). Young (1972) has shown that how snow accumulation within a river basin catchment varies according to the geometric properties of the surface. A "roughness index" was defined which is of great potential value for larger scales also.

Highly dissected topography with precipitous slopes can some times inhibit glacierization. However, if snowfall is ample and effective, and particularly if conditions are favourable for accumulation of rime ice, glacier can still maintain themselves apparently without difficulty. Also, tropical glaciers can exist on slopes of more than 40°C a high proportion of total precipitation is in the form of hail or rime ice (Tricard, 1969).

(f) Aspect

The orientation of the ground surface with respect to incoming solar radiation is particularly important at the local scale. Chorlton and Lister (1971) in their studies of Norwegian glacierization found that aspect exerted relatively little control over glacierization at the regional and larger scales. However, slope orientation or aspect experts a profound control particularly in marginal upland situations where the regional snowline is not far below the mountain crests (Evans, 1969). In such areas the snowline may vary through a vertical range of several hundred meters according to slope aspect. This is because aspect affects the surface receipt of both solar radiation and precipitation, particularly in middle and high latitudes. Steep north facing slopes receive the least direct radiation in the northern hemisphere. Thus, combined with the fact that north east facing slopes are the lee side slopes in areas of prevailing south westerly winds, explains why the cirque glaciers and firm fields of many upland areas are preferentially oriented towards the north-east.

2.1.5 Transformation of snow to glacier ice

The winter snowfall on a glacier surface is partly removed by melting and runoff in the summer. The newly fallen snow consists of hexagonal crystals whose form depends on the conditions of their formation. The crystal shapes are rounded by the temperature-dependent diagenetic processes such as evaporation, condensation and diffusion processes.

Within several days the snow consists of a loose aggregate of rounded grains whose physical properties change rapidly with time and increasing load. The term 'snow' is usually restricted to mataerial which has undergone little modification since it fell. The material in the intermediate stage of transformation is called 'firn'. This term is generally applied to snow which has survived at least a summer melt season and has begum its transformation. It consists of loosely consolidated, randomly oriented ice crystals with inter-connecting air passages and a density generally greater than 0.4 gm/cc. As the new firn is subsequently burried by additional years of snow and firn accumulation it is compacted and metamorphosed. Transformation of firn to ice takes place by a variety of processes whose overall effect is to increase the crystal size and eliminate the air passages (Paterson, 1969). When consolidation has proceeded sufficiently to isolate the air into separate bubbles the firn becomes "glacier ice". The glacier ice is impermeable to air and water. The air may be present only as bubbles. This change to ice takes place at densities of between 0.80 gm/cc. Further compression of air bubbles increases the density of ice until it approaches the pure value of around 0.90 gm/cc.

Meltwater accelerates packing by lubricating the grains, and permits very close packing because the surface tension of water film tends to pull the grains together. Thus, the maximum, density which can be obtained by packing is higher in a meltwater area than in a dry snow zone.

The transformation of snow to ice at a given place can be shown by a graph of density versus depth. Two such curves, smoothed to some extent, have been shown in Fig.2. Langway (1967) made the measurements at site 2 (latitude 77°N, longitude 56°W) in Greenland. The data for upper Seward Glacier in the Canada were obtained by Sharp (1951). Sitae 2 was near the dry snow line; their location was in the seaked zone of a temperate glacier. It has been observed that the transformation is much more rapid in the soaked zone than in the dry zone. Firn becomes ice (density 0.85 g/cc) at a depth of 13 meter on a Seward Glacier but not until a depth of 80m at site 2. In the percolation zone the transition depth would be probably between 35 and 75 meter. This difference is even more striking if expressed in terms of time by using the known rate of snow accumulation in each area. Snow is transformed to ice in 3 to 5 years on upper Seward Glacier : more than 100 years are needed at site 2. Bader (1960), Anderson and Benson (1963), and Costes (1963) have als derived formulae relating density and depth in a dry snow zone.

2.1.6 Flow of glaciers

To slide the glaciers on their beds, two processes are understood very important (Weertman, 1957). The first process is that ice can flow around bumps or obstacles on the bed because stress concentrations enhance shear strain rates. The other process is regelation : increased pressure on the upstream side of an obstacle causes the ice there to melt. The melt water flows around the obstacle and refre-eezes to fromregelation ice (Fig.3). The ice refreezes in the zone of decreased pressure in the lee of the obstacle. The processe operates best when the latent heat released by freezing can be transferred from downstream side. Thus it is found most effective when obstacles are less then 1m in size (Weertman, 1957). However, it is important that this figure of 1 m size is not taken too literally to mean that the process does not operate over 1 m in size. So long as the heat released by freezing can be dissipated, for example by meltwater flow in a real glacier, there is no reason to expect any upper limit. The existence of such a process of pressure melting has been confirmed by observations under active glaciers (Kamb and La-chapelle, Gow, 1970).

Weertman (1964) considered how a water layer might affect basal slip and concluded that a layer of water only a few millimeter thick could ,increase sliding velocity by 40-100%. This is because the water lifts the ice and decreases the area of friction between rock and ice. Lliboutry (1965,68,70) stressed the role of water filled cavities between ice and bedrock. Although these cavities occur in lee of the obstacle due to ice movement, it was considered that water may partly control their size and areal extent. Basal water pressure is regarded as an independent variable related to the amount of water made available by, for example, melting or rainfall. When there is sufficient water with a hydrostatic pressure

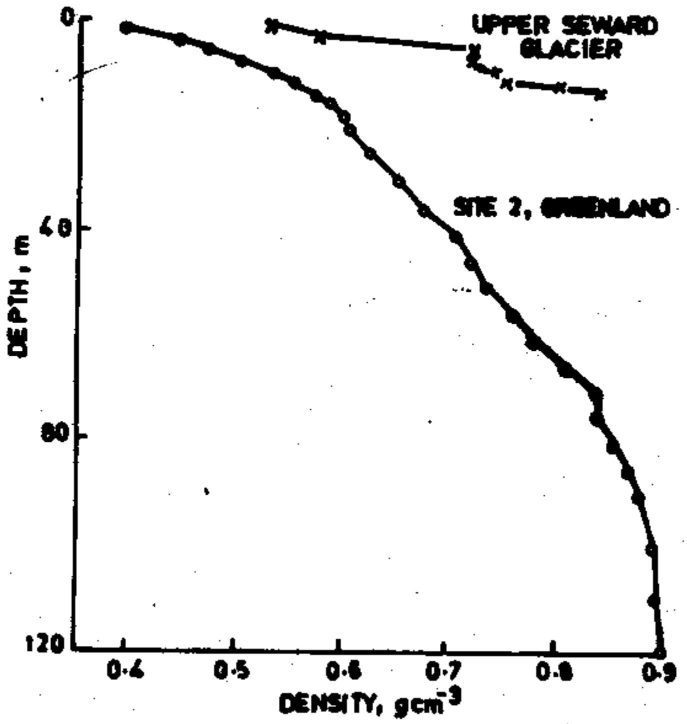

Fig.2: Variation of firn density with depth in a temperate glacier and in the Greenland ice sheet. From Sharp (1951) and Langway(1967).

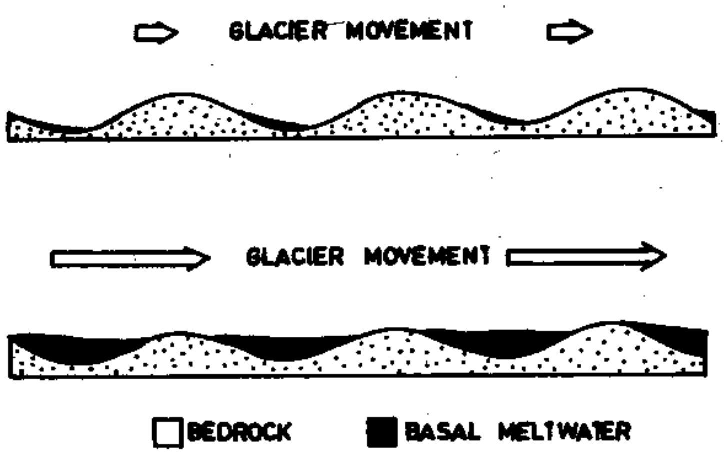

Fig.3: Diagram to illustrate how the growth of water-field cavities reduçes the area of contact between the glacier and its bed and thus reduces friction.

greater than that of ice (as will be the case if head of the water extends to the glacier surface or sufficiently far up glacier), the cavities will grow. A positive feedback mechanism may then come into play with larger water filled cavities leading to higher sliding velocities which then lend to still larger cavities. In all these various theories the bedrock is assumed to be impermeable. If it is not, then part of this water would be absorbed into the rock and this will reduce the sliding velocity of the glacier (Weertman, 1966; Doulton, 1972).

Some glaciers have been observed after many decades of quiescence of sluggish behaviour to suddently flow at very high speed (Post, 1960); these are called surging glaciers. Frequent uneven advances of these glacier tongue, having no direct relation to variation of external conditions, are the pricularity of such glaciers (Doglushin et al, 1963).

Normal glacier flow ranges from a few centimeter to a meter or so per day. Surging glaciers may attain speeds of meters per hour and ice near the termini may advance kilometers in a few months. It is reasonably well established that glacier surges are caused by some abrupt decoupling of glacier from its bed and that this decoupling is relataed to some unusual amount or condition of water. However, the actual process is yet to be explained (Meier and Post, 1969 ; Weertman, 1969).

2.2 Melting and Drainage Characteristics

2.2.1 Melting processes

The atmosphere above the glacier surface is known as the main source for supplying the energy for glaciers melting. However, very small contribution may also be received from the bed of a glacier. The glacier surface can gain or lose heat by several mechanisms. Their net result, at any given time and place, will be to change the surface tempereature if it is below $0^{o}C$ or to cause melting if the surface is already at $0^{o}C$ and is receiving heat. The basic processes of heat exchange are the same whether the surface is at its melting point or not.

The balance of energy fluxes passing through any surface is described by the following equation:

$$Q_m = Q_n + Q_h + Q_e + Q_g + Q_p - \frac{du}{dt} \qquad \text{... (1)}$$

where the subscripts n, h, e, g, p respectively refer to net radiation balance, sensible heat flux, latent heat flux, ground heat flux and flux of heat from rain. Q_m is the energy associated with flux of melt water and $\frac{du}{dt}$ denotes the rate of change of internal energy per unit area of the glacier.

For glacial melt studies the net radiation is a measure of the energy available at the glacier surface. This can be described as follows:

$$Q_n = (1-\alpha)\ G + L_i - L_o \qquad \text{....(2)}$$

where G is the global radiation, α is albedo and L_i and L_o are the incoming and outgoing longwave radiation, respectively Global radiation comprises of the direct solar radiation and diffuse radiation and generally falls within the wavelength range 0.3 μ to 3.0 μ. It has maximum intensity in the visible spectrum at about 0.5 μ wave length. The reflective properties are expressed in terms of albedo of the surface. An albedo of α implies that a friction $(1-\alpha)$ of the incident radiation is absorbed. The albedo of any surface is not strictly constant. It varies with elevationof sun and amount of cloud. Typical values of albedo for shortwave radiation are about 0.7 - 0.9 for fresh snow, 0.4-0.6 for firn, and 0.2 - 0.4 for glacier ice (Paterson, 1969). Baker et al(1982)also computed shortwave balance with typical albedo values 0.8, 0.6, 0.4 for snow, firn & ice respectively. The major sources of incoming longwave radiation are sky and surrounding terrain. In the atmosphere water vapour, carbon dioxide and ozone also emit longwave radiation. The longwave radiation are generally included in the wave length range 6.8 μ to 100 μ, containing its maximum intensity at around 11 μ. In the surface energy balance studies particularly over ice surface, it is not necessary to separate transmitted from the emitted radiation, and it is possible to consider the atmosphere as a grey body with an effective emissivity ε_a (Paltridge and Platt, 1976). Thus incoming longwave radiation is generally computed through the following equation:

$$Li = \varepsilon_a \, \sigma \, Ta^4 \qquad \ldots.(3)$$

where $\sigma (5.67 \times 10^{-11} KW/m^2k^4)$ is the Stefen - Boltzman constant and T_a is the temperature (k) of the air near the surface.

Heat is lost from the glacier surface by outgoing longwave radiation. The outgoing longwave radiation from a glacier surface is calculated by -

$$L_o = \epsilon_i \sigma T_i^4 \qquad \ldots(4)$$

where ϵ_i is the emissivity of ice surface and T_i is the absolute temperature of the ice surface (k). Kuhn (1979) considered the value of ϵ_i equal to 0.98, for glacier studies.

The transfer of sensible heat fluxes, Q_h and latent heat fluxes, Q_e is considered secondary in comparison to radiation. The turbulent exchange processes occuring in 2-3 meter of the atmosphere above the glacier surface are responsible for sensible heat and latent heat transfers. The downward temperature and water vapour pressure gradient directly transfer the heat and moisture from the air on to the glacier surface respectively. Transfer of water vapour results in condensation or evaporation at the surface and thus in libration or absorption of latent heat. Condensation of one gram of water vapour on a glacier surface librates enough heat to melt about 7.5 gm of ice. The turbulence fluxes are expressed in the following way (Schmidt, 1925; Prandtl, 1932):

$$Q_h = C_p \rho K_h \frac{\partial T}{\partial Z} \quad \text{.... (5)}$$

and

$$Q_e = L_v \rho K_e \frac{\partial q}{\partial z} \quad \text{.... (6)}$$

where ρ is air density, C_p is the specific heat at constant pressure, Lv is the latent heat of vapourization and K_h and K_e are turbulent transfer coefficients or eddy diffusivities for sensible heat and water vapour respectively. These equations have developed on the assumption that vertical fluxes are constant with height. The K_h and K_e are computed from wind profiles. The procedure has been outlined by many investigatiors (Priestley, 1959; Kraus, 1972; Brown, 1974; Schwerdtfiger, 1976; Anderson, 1976).

Bulk aerodynamic formulae are also used to compute the sensible and latent heat transfer to the glacier surface. These employ bulk transfer coefficients derived from flux gradient relationships and semi-empirical profiles forms of wind, temperature and humidity. These usually take the form:

$$Q_h = D_h U_z (T_a - T_i) \qquad \ldots.(7)$$

$$Q_e = D_e U_z (e_a - e_i) \qquad \ldots.(8)$$

where, D_h and D_e are the bulk transfer coefficients for sensible heat transfer and latent heat transfer respectively. U_z is the wind speed at a reference height (1-2 meter above glacier surface. e_a and e_i are vapour pressure at reference height and ice surface respectively. Previous investigators have also shown that over alpine glaciers $Q_e < Q_h$ by an order of magnitude (Lachapelle, 1959; Ambach and Hoinkes, 1963; Hoinkes 1964; Streten and Wendler, 1968). Ahlmann (1948) has pointed out that in maritime subpolar latitudes the turbulent heat transfer is much more significant for the glacier heat budget than radiation. Later on, Lambeth (1951) confirmed this to be ture and found rain to produce the greatest ablation. But precipitation is difficult to measure and therefore uncertain in windy and mountainous regions.

The transfer of heat from the underlying ice surface, Q_g, is described by

$$Q_g = - K_t \frac{\partial T}{\partial Z} \qquad \ldots.(9)$$

where K_t is the thermal conductivity and $\frac{\partial T}{\partial Z}$ is the temperature gradient at the ice surface. The thermal conductivity of ice has been observed approximately four times than that of water. The surface will gain or lose heat according as temperature increases or decreases with depth. In the early part of summer the surface will be warmer than the layer immediately below it and heat will be conducted away from the surface. Infact heat will flow downwards as a result of the pressure melting gradient. But as this gradient is only about 0.7°C per 1000m, so the heat flow is negligible.

The heat transferred to the ice by the rain water is the difference between its energy content before falling on the glacier and its energy content on reaching thermal equilibrium within the pack. The heat supplied by rain to each square centimenter of surface per second is given by

$$Q_p = C_w P_r (T_r - T_i) \quad \text{....(10)}$$

where C_w is the specific heat of water P_r is the rate of precipitation in g Cm^{-2} S^{-1}, and T_r and T_i are the temperature of rain and ice surface respectively. If the surface is at melting point, rain only transfers a small amount of heat. It is assumed that about 10 cm of rain would have to fall in a day to produce the same amount of heat as does longwave radiation. If the surface temperature is below freezing point the rain will freeze and each cubic centimeter will release 80 cal to the ice. In such cases rain will be significant source of heat.

2.2.2 Induced melting

Dusting of snow and ice surfaces by a dark material is known to cause an increase in their melting. It is considered that the dusting of such surfaces change the intensity of melting mainly because of change in albedo due to dust within the layers (Avsiuk, 1953). Natural examples of intensive glacier melting under the influence of dusting of their surfaces with proclastic material can be found in volcanic regions of Alaska, Iceland and South America. One of the major application of dusting the snow and ice surfaces could be to increase the quantity of usable water supplies and alter the timing of flow. Other applications may be more important such as, opening water distribution system in the spring for irrigation and opening of winter feeding areas for wildlife. During the years of after world war II a number of experiments on arctic ice by means of surface dusting were carried out in U.S.S.R. The amount of additional melt water yield depends on the natural pollution of the ice and snow surfaces, on the extent of dusting, on weather conditions prior to and after the experiment, and also on the size and position of the experimental area of glacier.

Research on surface - blackening materials began with studies of reflective properties of various materials over a styrofoam base (Megahan, 1968). Carbon black and powdered charcoal at the rates of $24 g/m^2$ increased shortwave radiatin absorption by 6 to 10 times. Coal (0.5 - 2 mm diameter) and various size of Coke (from less than 0.18 mm to 2 mm diameter) at the rates up to 127 g/m^2 increased shortwave absorption as much as 2 to 3 times and gave an approximately linear increase in effectiveness with increasing rate of application. Megahan et al (1970) conducted studies similar to Styrofoam on snow at Rocky Mountain Front Range in Northern Colorado, U.S.A. These tests were conducted over periods of only one day. Net radiation to the snow was increased an average of 80 percent with a range in increase for the various treatments from 36 to 122 percent. The carbon blacks gave greater increases than fluid coke; the smaller particle sized fluid coke was more effective than the larger. Meiman (1972) reported the treatments to know effects over periods of several weeks or longer. Plots 2x2m were established at an elevation of 3750 m in June. Fifty five plots gave the opportunity to evaluate ten different treatments each replicated five times. After two days, the ablation on treated plots was from 1.7 to 2.4 times greater than that on control plots. Fluid coke applied at the rate of 127 g/m^2 had a significantly lower ablation rate than carbon blacks at 36 and 48 g/m^2. During the two to six days after treatment, the increase in ablation ranged from 1.6 to 1.8 times the ablation on the control. Average ablation was increased 1.4 to 1.6 times during the 6th and 9th day after treatment. The artificial ice dusting is more effective during the first several days. This happens because the fine dust is gradually washed away from the areas by the melt water.

Kotlyakov and Dolgushin (1972) reported about 100 experiments performed each for 20 hours. It was found that the particles of the blackening substances not only absorb the solar energy and use it for melting but also destroy the upper porous ice horizon and moisten it. This leads to an additional decrease of environmental albedo where particles are placed and causes an additional increase in melting. Some 'radiation brush' (pitting of the surface) appears on the dusted surface. It decreases the albedo value 10-12%, and that causes a further increase of melting by 25-30%. The "radiation brush" periodically is destroyed and reformed again, causing albedo and melting variations to have a cyclic character. The experiments showed that temperatures of 0°C or a little lower are the most suitable for the artificial increase of melting. At a higher temperature and especially at lower temperatures the increases in melting are not so great.

Infrequent summer snowfalls on glaciers have a negative effect on the artificial increase of glacier melting. However, snow on the blackened surface disappears faster than it would from a natural one. The maximum increase of melting by the particles of the blackening substance would be achieved if the particles placed on the snow and ice stop sinking to the melting thickness in the immediate range of the surface. In this case the activity will be not only more powerful but constant also. It was observed that at the same rates of dusting, the increase of melting on clean firn in the accumulation areas of glacier is more than that on the polluted glacier tongues. A theoretical study of ice surface dusting influence on melting intensity was also carried out by Zotikov and Moiseeva (1972).

The most efficient rate of snow dusting, when each gram of Coal dust gives the highest quantity of melt water, was found 5 g/m^2 (Kotlyakov and Dolgushin, 1972). Such a rate of dusting gives the optimum distribution of the blackening substance on the snow surface. The highest efficiency of melting was achieved using coal dust particles 0.2 mm in size. It was shown that at a 2 g/m^2 rate of blackening the increase of melting is the same as at 25 g/m^2 with crushed slag 0.4 mm in diameter.

2.2.3 Infiltration of melt water

The speed and direction that the meltwater takes from the snow or ice surface to the outlet of a glacier is an essential parameter in many hydrological analysis of glacier basins. The relative amount of water passing through rapidly or slowly, or in temporary storage in a glacier has been a matter or controversy (Paterson, 1964, 65; Meier, 1965). The importance of resolving this problem is critical in regard not only to glacier hydrology but also to glacier dynamics because of important role of water film or water packets at the bed in controlling the rate of glacier sliding.(Elliston, 1963; Lliboutry, 1968; Weertman, 1969).

Water on the surface infiltrates a porous medium (snow or firn) or runs into discrete openings in a Karstlike medium (ice). In the firn basins a snowpack before the beginning of melting falls clearly into two parts according to its water infiltration properties, there is at the top a pack of almost homogeneous fine-grained snow with greater water permeability and small density. Deeper the snow gives place to a dense firn with interlayers and lenses of ice whose amount of thickness increase with depth. After beginning of melting, with the advent of meltwater in the snow in the vertical profile of the pack affected by infiltration, a clear differentiation in its structure at once begins manifesting itself. The pack starts separating into wash-out horizon in the upper part and a wash in horizon in the lower part. In the former horizon infiltration loss of matter takes place while in the latter infiltration accumulation occurs. The two horizons are displacd downward affecting deeper layers, as the surface melts and sinks. Throughout the whole periods of melting continous redistribution of water storage take place in the pack. However, the differentation mentioned remains in the pack upto the end of melting and subsequent winter freezing.

A considerable ice accumulation does not occur in the wash-in horizon as infiltration processess in the snow thickness take place. When meltwater attains the boundary between snowpack and last years firn and the rate of infiltration front movement sharply decreases, meltwater begins to accumulate at that level and ice starts to grow intensively. Gradually a pack forms which has a layered structure: interlayers and lenses of ice alternate with firn lenses. In case of plenty meltwater, the cold storage fails to freeze it, a part of water flows down along the impermeable horizons and other parts penetrate deeper. Its maximum accumulation occurs in deeper layered horizons (Fig.4).

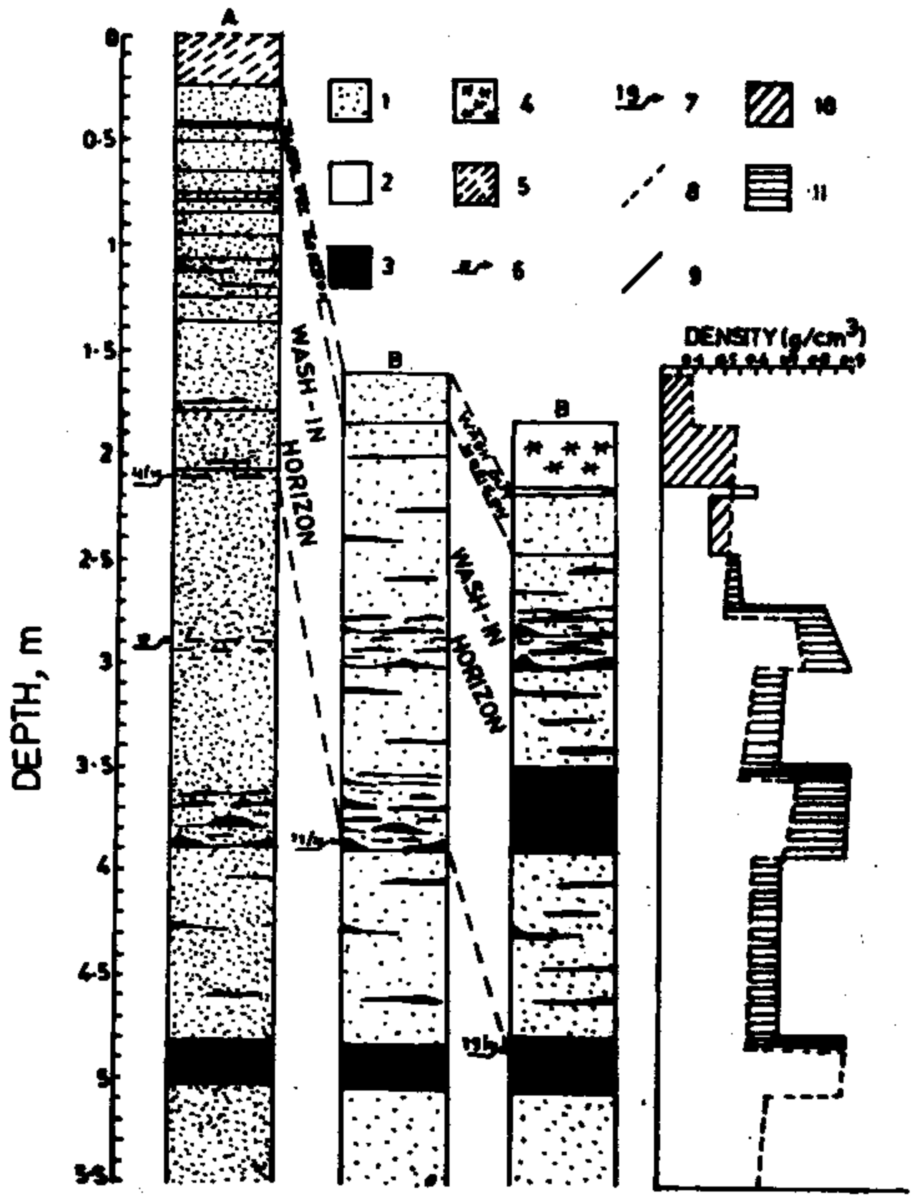

Fig.4: The variation of firn structure on Elbrus in the 1962 summer: A-4 June, B - 13 August, B - 19 October, 1 - the snow of 1961/62 winter, 2-firn, 3-ice, 4-the snow of 1962/63, 5-snow which melted from 18 May to 4 June, 6 - the level of marking by dye, 7-infiltration front level on a given data, 8-the density curve on 13 Aug., 9- the same on 19 October, 10- an area proportional to water to equivalent decrease, 11- the same proportional to water equivalent increase. On the left is a depth scale in meters.

The interior of a glacier is not accessible, therefore observations of water flux except at the surface and outlet are generally impossible. An exception is in the snowpack where flow can be examined to depths of several meters with some difficulty.

A large numbers of experiments have been conducted in the laboratory and in the field, few have properly duplicated the thick stratified snowpack found on many glaciers, and few have treated the movement of water through the complete snow-firn-ice system of the glacier. Colbeck (1971) analysed the one dimensional movement of water through snow according to the theory of two-phase flow in porous media. These results seem to be generally confirmed by the field observations of Sharp (1951). Bazhev (1969) made some observations of meltwater infiltration on Mount Elbrus (Central Caucasus). The results shown in fig. 5 demonstrate that the upper part of a cross-section (about 3 m) underwent infiltration for 20 days and lower part about 2 m for 83 days. The difference in the rate of front infiltration movement were due to the difference of physical and hydrological properties of different parts of the pack. From an average rate along the whole corss-section of 4.5 cm/d in the upper part it smoothly increases with depth reaching its maximum 25cm/d at the bottom of snowpack. At the interface with the firn the rate sharply drops down almost to 2 cm/d. Deeper in the firn the rate curve has the nature of a sinusoid with an applitude decreasing with depth. The curve of the change in the rate of infiltration front movement reflects the influence of two factors: the cold storage and the

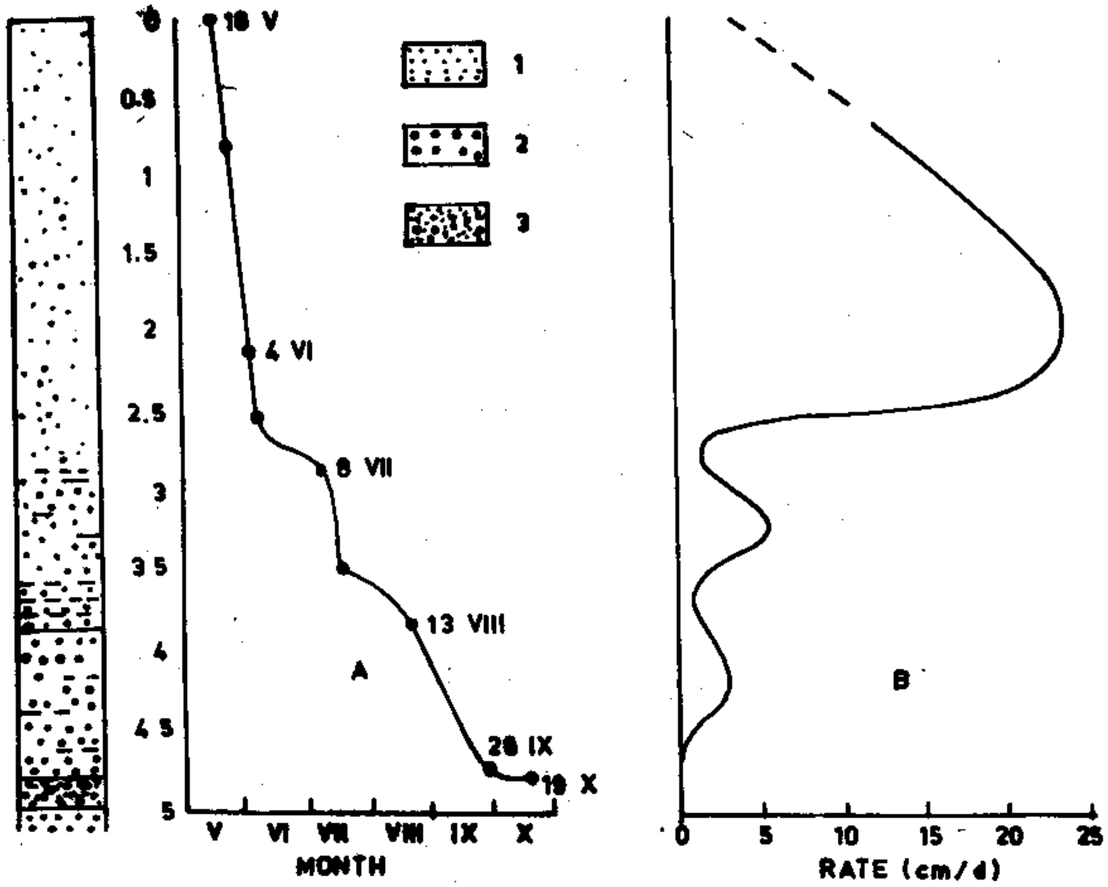

Fig. 5: Infiltration front penetration deep down the firn (A) and the variation of water infiltration ratio with depth (B) on Mount Elbrus in 1962; 1-snow, 2-firn, 3-ice.

water equivalent of the pack. The former slows down the infiltration most in the upper part of the cross-section (because at the top cold storage is more) and its effect diminishes with depth. The latter suppresses the infiltration in the upper layers slightly and in the lower ones to a much greater extent. Along with rapid decrease of the water permeability downward the infiltration rate becomes variable because of occurence of ice layers.

Krimmel et al (1972) made few studies of water infiltration through cores and pits. Dye injectious on the snow surface were used for this purpose. The cores and pits through dye patches indicate that large variations in the percolation rate of the dye through snow and firn can be expected. Rhodamine B was spread over six similar areas of about 0.6 m of snow depth having density about .5 g/cc. Five patches were in the centre of runnels and sixth was between runnels. The underlying one year old firn had a density of about .65 g/cc. After 2,5, 11, 25 and 35 hrs, pits and cores were cut through the snow under the patchers in the runnels. These experiments showed dye penetration of 0.14 m/hr, 0.12 m/hr, 0.73 m/hr and 0.051 m/hr, respectively. The patch between the runnels showed the penetration rate of 0.08 m/hr.

2.2.4 Water storage and drainage system

Liquid water is effectively stored in several locations within an apline glaciers, such as surface snow, firn, pools on the ice surface, crèvasses, englacial pockets, subglacial cavities, the moulin conduit network and basal moraines, while in transit at differing velocities through the various hydrological pathways of internal drainage system. Water storage in glaciers fluctuates over periods of several days with hydrometeorological conditions, resulting in melt-stream discharge having diurnal saw toothed which varies more slowly.

The internal drainage of glaciers has a wide range of significance in different scientific fields. The hydrologists are interested in the rate of internal drainage as a main characteristics of a glacier catchment areas. The glacial geologists and geomorphologists require a knowledge of drainage for interpretations of glacial features or discussions of deglaciation sequences. The basic conditions for the initiation of internal drainage at the ice surface are the existence and collection of superficial run off and possibilities for the water to penatrate down into the ice body and from there to be drained off further (Stenborg, 1969). The movement of meltwater arised from two reservoirs runoff derived from ice-melt in ablation area and water derived from the firn in the accumulation area has a number of pecularities. The main one being that ice-runoff may be divided into two parts, surface and deep flows. Surface runoff consists of water running on the glacier surface in a network of streams, from tiny rivulets to wide streams. The water appears as a result of absorption of energy coming from the atmosphere, and since this absorption occurs over the whole glacier surface, surface water flow is not discrete (in contrast to deep runoff) but presents a picture of continuous flow of water on the glacier surface. In some localities, depending on the relief, the water layers are composed of streams of rivulets, in other localities it may be extremely thin or even in the form of films. Open water on the glacier surface has been disappearing into vertical holes called "moulins".

When crevasses occur in snow and firn covered parts of the glacier, they introduce a channel for water to penetrate deeper into the firn than it could in the absence of the crevasses. This is because ice layers, which are parallel to the surface, prohibit water percolation. The crevasses, by breaking the ice layers, provide vertical path for the water to move down and spread laterally on deeper ice layers.

The crevasses also act as loci of conduits which extend below the crevasses into unknown depth of the plumping system. Meier (1972) described that crevasses extend only to a comparatively shallow depth in a glacier. Normally they are not deeper than a few "tens" of meters. The, crevasses are most important in the surface zones of a glacier. Many studies have been made in which dye tracers were introduced into crevasses and these have shown that water travelling on the surface in the ablation area the ice zone does infact funnel into crevasses.

The deep ice-runoff is formed in channels penetrating the body of glacier. These streams are fed by surface runoff. In the snow and firn part of the glaciers, percolation is found very slow. Flow in this unsaturated porous medium proceed at varying rates because of continual changes in the amount of water saturation at any level. When water has percolated through some meters of the old firn it seems likely, although it never been confirmed, that porous medium type of flow gives way to flow through many small tributary channels. These channels join others to form larger channels so that water finally reaches at the base of the glaciers in a relatively small numbers of fairly large channels. After penetration to the bed of glacier water flows along the bed towards the terminus. It may flow in conduits of appreciable size (Embelton and King, 1968) which are curved in the ice at or near the ice/bed rock interface. Finally water emerges from the terminus in one or several discrete streams. Th morphology and structures of large vertical conduits have been described by Stenborg (1968).

Studies of isotopic and chemical composition of meltwater from Alpine glaciers suggest that ice-melt water drains through conduits in a few hours (Behren et al; 1971; Collins, 1979), whereas water from the firn percolataes slowly being stored in firn acquifer enroute to conduits (Lang et al,1977). Consequently, only a portion of the meltwater produced each day reaches the portal on the same day. Runoff of the remainder as delayed until succeeding days, adding to later daily melt contribution (Martinec, 1970). How much of stored water actually becomes integrated with mortal outflow depends on the state of connection from the firn area to conduits, conduit network development, and on the length of recession period. If favourable conditions promote increased ablation, conduits in the ice will rapidly wider allowing increasing flow, whereas pathways from the accumulation area do not much expand and meltwater in firn leading to maximum water levels in the firn acquifer (Oertev, 1981). As the internal hydrological system enlarges in capacity and extent in summer, the improved efficiency of through flow progressively reduces runoff delay during summers (Elliston, 1963). Sudden termination of ablation by snowfall may interrupt periods of either increasing, steady or already declining flow.

Nye (1969) pointed out that the weight of the overlying ice would tend to close the channel drainage system, but this tendency is restricted by the water pressure and also, as Weertman (1962) suggested, by the melting that will occur if the water is initially warmer than ice; there is also melting that comes from frictional heating in the water. There is another process also that tends to close the channels; it arises from the forward motion of the ice over the bed and is purely geometrical in nature (Fig.6.). This process, like the action of turning off a value is quite distinct from a pinching-off due to over-burden pressure, for it can occur even when the water pressure, balances the over burden pressure. For example, in Fig.6 , the water pressure could be equal to the ice over burden pressure throughout the closing off process. If the drainage system is to be continuous, as seems likely, there must, at any one time, be sufficient length of channel that is stable against both sorts of closure process. The channels must not only have a sufficiently high water pressure in them, but they must be characterized geometrically by the fact that forward motion of the ice does not readily seal them. In general a channel incised into bed would not be vulnerable to the sort of closure process typified by Fig.6. On the other hand a channel incised upwards into the ice would move with the ice and, as in Fig.6 , it would be liable to become sealed off by meeting a protuberance in the bedrock. One therefore expects that any channel system incised upwards into ice will be constantly changing, with lengths constantly becoming closed-off, while a channel system incised into the rock, once formed, would be much more permanent. Thus erosion of the bed by water to form channel, once started at a particular place, will tend to continue at that place: thus the necessary time is provided for erosion to produce a well-developed channel into bed rock.

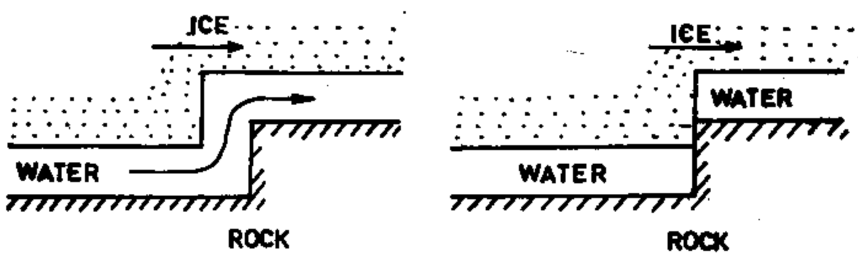

Fig.6: Closure of channel by ice movement.

2.2.5 Delay of runoff

The average basin lag τ is the characteristics of ablation and accumulation areas together. In fact, basin lag is different for the ablation area and accumulation area. It may be expected that runoff from the ice tongue is faster than the runoff from firn in accumulation area. The impermeable ice surface and an inconsiderable roughness of ice channels contribute to fast runoff of meltwater. It is supposed that water which is caught by revasses moves rather fast as well, because during the ablation period an underglacial drainage system could be developed well. In the opposite case the water should be in the bottom of crevasses, but in reality this is observed rarely.

The graph of hourly runoff from the glacier is saw-tooth-shaped, moreover the peaks of discharges are on a rather high base. It is assured that mainly these peaks are formed because of the runoff from the ice tongue and the base flow is formed by the firn melt water and by the water accumulated in moraines and lakes. The verification of these assumptions has been carried out by Golubev (1969). Further, Golubev (1969) has shown that for Lednik Dzhankuat the correlation $W = f(Q)$ is quasilinear, and than $\tau = \frac{W}{Q}$. The accumulation volume W is formed by the water of tongue W_i and the water of the firn W_f i.e.

$$W = W_i + W_f \qquad \ldots.(11)$$

Approximately accumulation volumes can be expressed in such a way that

$$W_i = \tau_i Q_i \qquad \ldots(12)$$

and

$$W_f = \tau_f Q_f \qquad \ldots(13)$$

Where τ_i is an ice tongue lag and τ_f is a firn lag. Q_i and Q_f are water discharges from the ice and the firn respectively. Therefore

$$\tau = \frac{W}{Q} = \frac{W_i + W_f}{Q}$$

$$= \frac{\tau_i Q_i + \tau_f Q_f}{Q}$$

$$= \tau_i \frac{Q_i}{Q} + \tau_f \frac{Q_f}{Q} \qquad \ldots(14)$$

or

$$\tau_f = \frac{\tau - \tau_i \frac{Q_i}{Q}}{\frac{Q_f}{Q}} \qquad \ldots(15)$$

In an experiment it was demonstrated that during the ablation periods of 1965 and 1968, the yield of glacier tongue was nearly 40% of water yield from the whole glacier and outflow of the accumulation area was 60%: $\frac{Q_i}{Q} = 0.4$; $\frac{Q_f}{Q} = 0.6$; $\tau = 2.5$ days; $\tau_i = 0.125$ d. From equation (15) follows that $\tau_f = 4$d. So, the obtained values of τ_i and τ_f demonstrate a very different effect of ice and firn stream flow control. The values of have been obtained for the outlets situated next to the snouts for 1-5 years of observations. A relationship was drawn between τ and S where, S is the active drainage area of the glaciers. For small and medium glaciers the following relationship was discovered

$$\tau = 3.8 \log (S+1) \qquad \ldots(16)$$

where is in days and S is in Km^2.

To obtain a rough quantitative measure for the purpose of comparison a recession coefficient $K(\tau = \frac{1}{k})$ was calculated as follows (Lang, 1969).

$$K = \frac{\Delta q}{\Delta h} \cdot \frac{1}{\bar{q}} \qquad \ldots(17)$$

where Δq is the difference between the monthly means of discharge at 1800 hrs and subsequent 0600 hrs in m^3/s; Δh = 12 hrs; $\bar{q}$ mean monthly discharge in m^3/s. As the falling segments of mean daily variations are not undisturbed dry-weather recession curves, the coeffieient K is only approximately equal to the recession coefficient in the equation $q = q_o e^{-kt}$. The coefficients for the single months of the ablation seasons for the year 1965-68 are shown in Table 3.

Table 3 : Recession Coefficients (hr^{-1})

Year	May	June	July	Aug.	Sept.
1965	0.0108	0.0112	0.0162	0.0181	0.0194
1966	0.0104	0.0114	0.0186	0.0196	0.0217
1967	0.0115	0.0124	0.0159	0.0168	0.0166
1968	0.0142	0.0128	0.0185	0.0177	0.0134
Mean	0.0117	0.0120	0.0173	0.0181	0.0183

Early in the melt season, when the channels are scare, small and irregular, the melt water will take longer to reach the outlet. Later the water courses will be more direct, particularly in the lower parts of the ablation area and therefore, only small time lags will occur. A rising tendency of the recession coefficient K, particularly from June to July confirm the decrease in the runoff delay (Table 3).

A large number of experiments were undertaken on South Cascade Glacier, Washington, during the ablation seasons in attempt to clarify some of the problems of time delay of water through a glacier and storage mechanisms of water within a glacier (Krimmel et al, 1972). Three types of experiments were performed; some dealt with the snowpack only, some dealt with water travel from a snow surface to the glacier to the terminus, and some dealt with the time taken by water flowing freely in open channels on ice or entering the glacier at the sides to travel the terminus.

The presence of glaciers in a basin also delays the time of maximum runoff (Tangborn, et al, 1975; Fountain and Tangborn, 1985). This delay is caused by temporary liquid storage of spring melt water in the glacier, which is released later in the summer, and by the peak meltwater production occuring in mid-summer. It was shown that for unglacierized basins the maximum runoff occurs in May and for glacierized basins maximum runoff occurs later and later as the percentage of glacier cover increases Fig. 7. The delay is caused, in part, by later snow melting with increasing altitude. It has been reported that the time of peak seasonal flow occurrence from an unglacierised basin as a function of basin mean altitude shows this relationship to be linear and of small influence in comparison with the parabolic relationship between time of maximum seasonal runoff and percentage glacier cover Fig. 8. The maximum flow may be delayed as much as a month, if the basin glacier cover changes from 5 to 15%, whereas it is delayed only about 2 more weeks as the glacier cover increases from 50 to 100%. This has an important implications for hydroelectric site selection and for the effects of elimate change on streamflow. If the climate changes, the glacier will respond by changing size, which will result in a change in the time of appearance of the maximum stream flow, especially for basins with small glacier covers.

2.2.6 Modelling of Glacier melt

Derikx and Loijens (1971), Campbell and Rasmussen (1972), and Kraus (1972) have made attempts to develop models to calculate the daily runoff from glacierized basins. The input consisted of selected meteorological parameters, basin characteristics etc., and construction of models are based upon known formulae within glacier meteorology. Attempts have also been made to relate daily runoff to incoming radiation, air temperature, atmospheric moisture, wind velocity and precipitation. The influence of glacier melt and precipitation on the daily runoff was calculated for separate 100 meter elevation bands, and a runoff component was also included

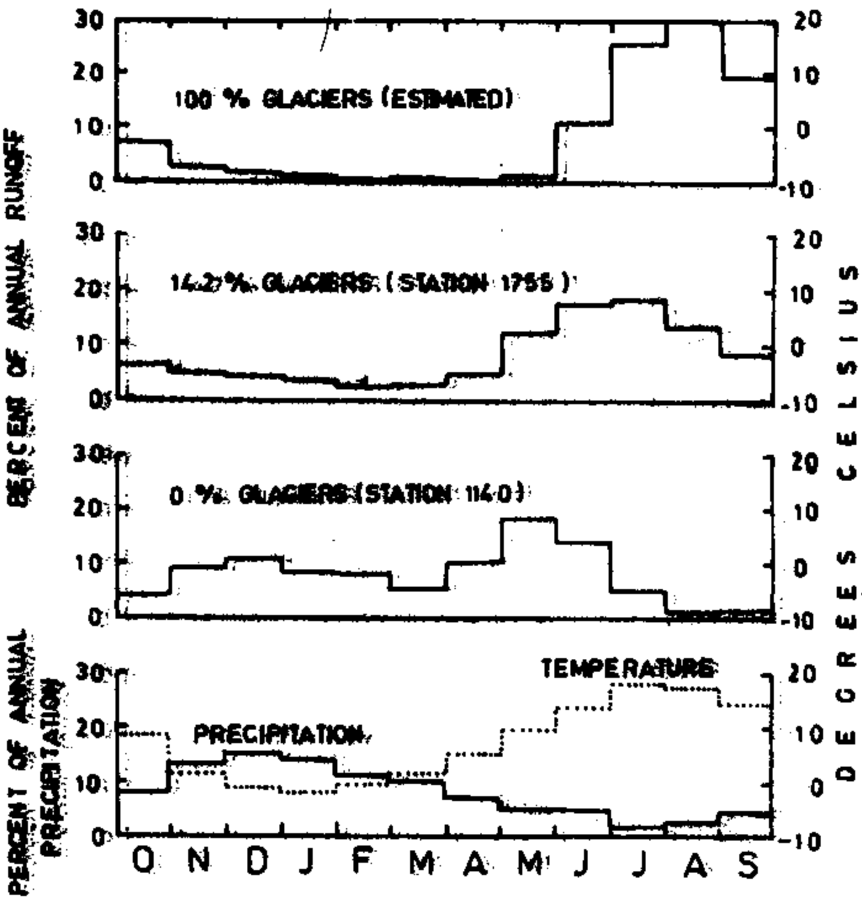

Fig.7: Monthly fraction of the annual specific runoff for basins of various glacier covers. The monthly faraction of precipitation and mean monthly temperature(Snoqualmine Pass, Washington) are included for comparison.

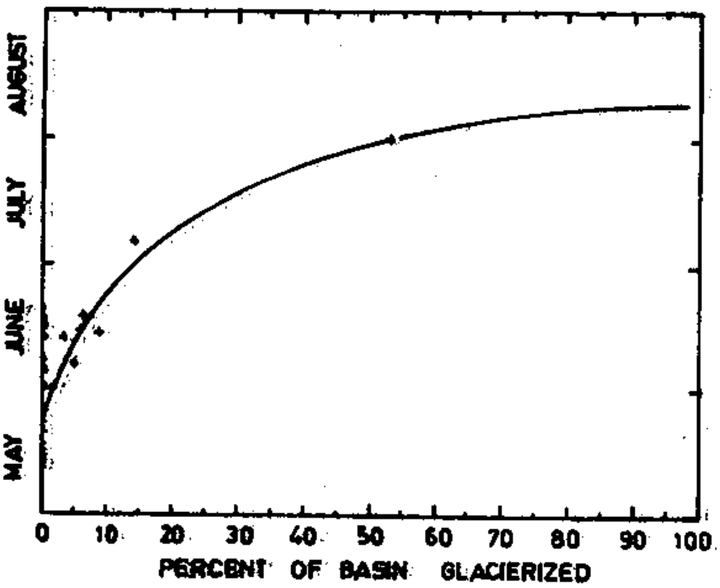

Fig. 8: Timing of peak specific runoff as a function of glacier cover for basins in the North Cascades, Washington.

for non-glacierized part of the basin. Each component of energy balance equation was computed according to known relationships. For example, incoming radiation was calculated for various areas according to slope and orientation and known sun angles throughout the days. Efforts were made to calculate direct (clear sky) and diffuse (cloud cover) radiation separately, based upon recorded hours of sunshine. Corrections were made for energy absorbed by water and ozone in the atmosphere. The albedo for ice was kept constant at 0.40 whereas the albedo of snowcover allowed to undergo a continuous decrease from an original value of 0.80.

Ostrem (1972) tried a model on two glaciers situated in completely different environments: Peyto Glacier in the Rocky Mountains (maritime). In such computations four meteorological parameters, air temperature, atmospheric moisture, wind velocity and precipitation were used. It has been reported that the daily computed values proved to be within ± 21 percent of the obseraved values for Peyto Glacier, whereas they were closer to the observed ones for Berendon Glacier. For most of the melt season the daily basin runoff can be calculated with an average accuracy of 10-20 percent provided basic meteorological observations are available for one station within or close to the basin (ostrem, 1972). It was also mentioned that daily variations in meteorological parameters and daily discharge from a glacierized basin, may reveal certain correlations. A warm period will, in general, cause a rise in water discharge, as will also a rainy period. In maritime catchments a period of humid and mild winds will promote melting and also cause a rise in runoff (Paterson, 1966; Ostrem, 1966).

Krenke (1971) reported that the relationship of air temperature and glacier melting was reanalysed together with Khodakov (1966). On the basis of experimental data from different glacier regions the following relation was emerged:

$$M = (T + 9.5)^3 \qquad \ldots (18)$$

where M is the total melt for the ablation season from the glacier surface expressed in millimeter and T is the mean temperature at height of 2 m over the glacier surface for summer period (June, July and August). Standard deviation was about 20%.

Khodakov (1971) investigated the dependence of summer ablation of the glaciers on mean summer (July, August) air temperature at a height of two meters above the surface. Analysing 93 pairs of values by the least squares method the following formula was obtained

$$M = 0.096\ (T + 10)^{2.93} \qquad \ldots (19)$$

where M is in g/cm^2 . Practically it coincides with the earlier formula using a cubic parabola derived with less data (Khodakov, 1965). Neverthless, the increase of data caused an increase of mean square deviation of calculated values compared with values measured up to $\pm$ 89 (g/cm^2). Consequently, another variable was introduced in the dependence, the shortwave radiation balance for three summer months, RS (k Cal/cm^2). Taking into consideration the dissimilarity of the variables and the complexity of the calculations, the squares method was preferred and the following formula was derived by way of selection (Khodakov, 1978):

$$M = (T + 1.3\sqrt{Rs} + 4.0)^3 \qquad \text{.... (20)}$$

on the basis of field data it is assumed that

$$Rs = 0.32\ R \qquad \text{... (21)}$$

where R is the total radiation for a summer at the stations situated in non-glacierized areas, varying over the territory of the USSR from 32 to 62 K Cal/Cm^2. The empirical coefficient of 0.32 accounts for the albedo of melting firn (0.6), the increase of cloudiness and increase of total radiation of the clear sky with height as well as exposure averaged for the glaciers. This modification has been carried out for 50 groups of glaciers over the territory of USSR.

From studies made by Roald (1971) concerning the correlation between different climatological parameters and the runoff from the basin on a daily basis, Lundquist (1982) selected the following Climatological parameters for glacial melt computation: precipitation, air temperature, wind speed and humidity. In the first efforts, however, a poor correlation was found between wind speed and humidity and the obsered runoff. In the final computations the use of only precipitation and temperature measured at two sites was made and it had very much resemblance with simple degree-day method. The theory that air temperature is a good lumped measure of most of the parameters of importance for ice melt on a basin scales, was also cnfirmed. Because of an obvious seasonal variation in the value of degree-day factor, a simple cosine transformation was introduced and final ice melt equation was given as

$$M = C_x \; (1-\cos \; (2 \pi d/365)^n \; (T-T_o) \qquad ...(22)$$

where

M = meltwater volume (mm/time step)

C_x = degree day factor (mm/time step).

d = number of the day counted from 1st January

T = actual air temperature

T_o = air temperature at which melting begins

n = exponential with value around 2.

As the snowcover shrinked on the glacier surface, the degree day factor for snow was substituted by one for ice (of the order of two to three times larger) to represent the different albedo of ice.

An empirical relation between specific icemelt to mean daily temperature was derived by Young (1980). It has the form:

$$M_r = 1.56 + 5.338 \; T_m \qquad ...(23)$$

where

M_r = specific melt rate (mm d^{-1})

T_m = mean daily temperature (^oC)

Specific melt per day was calculated for snow free areas of ice and firn for points on a square grid (grid interval) = 100m) using the environmental lapse rate to determine temperature at each point and then applying the above described equation. Melt in firn area was estimated as 85% of ice melt for the same elevation (based on melt measurements at stake locations). Moreover, it was also assumed that all water derived from ice melt passed the stream gauge the same day. Only half of the melt from the firn area existed the basin on the same day as it was produced (the other half was added to half the calculated firn melt of the next day). It is considered that these assumptions are very reasonable for a glacier the size of Peyto Glacier, Aleberta ; however, the lags might have to be modified for much larger glaciers. Ostrem (1972) made an statistical analysis in which temperature (T), precipitation (P) and wind velocity (V) were used as independent variables. After a long series of experiments to combine variable (sums, products, exponential forms, such as T^n etc. it became clear that the best result was obtained when certain products of temperature and precipitation, another variable would be the product of temperature and wind speed, and in most cases, the temperature itself would be the third variables. The general form of the resulting formula was

$$Q = K + K\ (TV) + K_2 .\ (TP) + K_3 .\ T \qquad ....(24)$$

This formula would, for most glaciers, express the discharge (Q) for the next day. However, if each of the meteorological parameters are calculated as running means for two or three days (depending upon the size of the glacier) the formula makes it possible to predict water discharge approximately two days in advance. A check of the formula was made by applying it to previous years observations. Discrepencies between calculated and observed discharges were in general less than 5 percent. For single events, however, the difference could be as large as 15 percent.

It is difficult to explain physical processes that are connected to various parts of the resulting formula. Many of the 'independent' variables are infact intercorrelated. Correlation analysis made on the daily values have shown, for examle, that air temperature and humidity have a correlation coefficient of 0.6 - 0.8.

Collins (1985, 87) analysed the records of discharge from gauges on rivers draining from glacierized basins in the Swiss Alps for the period 1922-1983 together with climatic data in order to describe climatic variation and to determine the impact on total annual flows. It was reported that mean May-Septaember air temperature and annual discharge show considerable yet markedly parallel year-to-year variations at all stations. A fall in ten-year mean temperature of 1° C produced a reduction of about 25% in ten year mean discharge. High levels of explanations of variance of discharge were obtained using multiple regression on mean ablation season air temperature and winter precipitation accumulation, the degree of fit being greater for basins with larger percentage glacierization. Relationships between precipitation variables, such as total annual input and winter precipitation with runoff were reported weak.

Ablation under a debris layer:

Most large glaciers in the Himalayas are covered with debris in their ablation area (Moribayashi and Higuchi, 1977; Fujii and Higuchi, 1977). Zalikhanov (1969) also reported that a considerable part (26%) of the Lednik Bashil's tongue is covered with morainal material. The larger part of this area has morainal cover of 10-15 cm thickness and more. It is essential therefore to understand the effect of debris cover on ablation of glacier ice for predicting the water supply from mountain glaciers.

However, it has been postulated that there are two difficulties in estimating the mean value of ablation under a debris layer extending over a wide area. One is the difficulty of determining directly the thermal resistance of the layer in the field which is one of the essential parameters for ablation estimation. The other related to the fact that detailed information on meteorological variables is necessary for the estimation. In regard to former difficulty, Nakawo and Young (1981) put forward a method by which a reasonable estimate of the thermal resistance of debris layer can be obtained in the field. This method was successfully employed in analysing the experimental data (Nakawo and Young, 1982). Further, Nakawo and Takahashi (1982) simplified the model developed by Nakawo and Young (1982) in order to overcome the latter difficulty for its pracatical use in the field.

Zalikhanov (1969) measured that under the layer of moraine of 2-3 cm thickness, 342 mm of ice melted; under the layer of 10-12 cm thickness only 261 mm ice melted; and under the layer of 20-25 cm, 189 mm ice melted.

Vohra (1980) has also indicated the possibility that thick layer on ice surface contribute significant quantities of water to the water discharges.

Further Nakawo and Takahashi (1982) reported that a thin debris layer accelerates the ablation rate of the underlying ice, whereas a thick layer retards it. There is a critical thickness h_c at which the ablation rate for debris covered glacier ice is the same as for debris-free ice. Since it is convenient to express h in terms of a thermal resistance R, R corresponding to h_c has been discussed by Nakawo and Takahashi (1982). The value of R_c was determined by the following equation:

$$R_c = \frac{G\ (\ \ _o -\ \ )}{K_o\ (\ (K_o + K_r)\ T_a\ -G(1-\ \ _o)} \qquad ...(25)$$

where G is the global radiation flux (Wm^{-2}), is the albedo of debris layer, is the albedo of debris-free layer, K is the degree day factor for debris-free ice ($W\ m^{-2}\ deg^{-1}$), K_r = 4 $(273)^3$ = 4.615 $Wm^{-2}\ K^{-1}$, and T_a is the mean air temperature ^oC.

The value of R_c can be determined directly by an experiment in which ablation rates are measured under debris layers with different thickness (different thermal resistance). The comparison of the value of R_c estimated by equation (25) and value from direct method has been represented in Fig.9. The agreement between the two estimates is by no means excellent, however, equation (25) can give a reasonable estimate of the order of magnitude of R_c in spite of many assumptions and simplifications in deriving the equation.

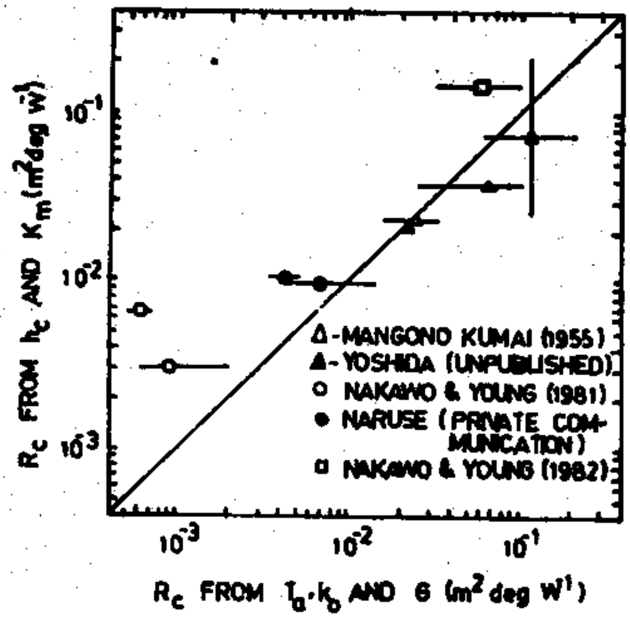

Fig.9: Comparison of R_c estimated from T_a,k_o and G with values from h_c and k_m. The horizontal bar corresponds to the standard deviation or range of each datum. K_m was not measured in Yoshida's experiment. The verical bar represents a range of R_c corresponding to the range of K_m for his experiment. K_m for Magono and Kumai' data was obtained from a value of T_s through the procedure given by Nakawo and Young (1982). k_o for all the data was calculated by $k_o = C_o/T_a$.

The following relation has been given by Nakawo and Takahashi (1982):

$$\frac{r}{r_o} = \frac{C}{C_o} = \frac{K}{K_o} = K^* \qquad \ldots(26)$$

where r is the ablation rate under debris layer (ms^{-1}), r_o is the ablation rate under debris-free layer, C is the conduction heat flux through debris layer (Wm^{-2}) and C_O is the heat used for melting ice per unit time for debris-free ice (Wm^{-2}). Therefore, one can estimate r or C if K* is known. It was shown that

$$K^* = \frac{1 + G^*}{1 + G^*R^*} \qquad \ldots(27)$$

where $G^* = G\,(\alpha_o - \alpha)/C_o$

and $R^* = R/R_c$

for a large R* in comparison with 1/G*, i.e. for at hick debris layer

$$K^* = 1/R^* \qquad \ldots(28)$$

It is suggested accordingly that K* is inversly proportional to R*. Zhang and Bai (1980) proposed an empirical relation between K* and h;

$$K^* = \frac{1}{h^b} \qquad \ldots(29)$$

~~it~~ was found that the value of b increased, approaching unity as C_o decreased i.e. $\frac{1}{G^*}$ decreased. Since h is approximately proportional to R*, their results can be considered as in agreement with equations (27) and (28).

Figures (10 & 11) demonstrate a plot of K* against R* for particular values of G* which are derived from experimental or field conditions as cited. The equation (27) is represented by the solid lines in this figure. Agreement between the theoretical predictions and the field data is quite good, although there is some scatter in the data.

There are other data available on K* and R* (Ostrem, 1959; Loomis, 1970; Moribayashi & Higuchi, 1972), however, no data on G* are given. Nevertheless, they are plotted in Figure 11 in which the dependence of K* on R* given by equation (27) is also shown for various values of G*. The data of each set of observations are compatible with a predicted curve corresponding to one of the values for G*, and the variety of the results obtained by the different authors can be explained in terms of G*.

Nilsson and Sunblad (1975) also proposed a model consisting of a system of reserviors. They tried to determine the model parameters as well as meltwater input by statistical methods. Later on Baker et al (1982) also presented a reservior model, but the model parameters and inparticular the melt water input have been determined and melting processes within and on the glacier. The glacier was divided into three different areas, each of which displays its own characteristic storage and discharge behaviour. Each area was represented by a so called 'linear reservoir'. In this model, however, rainfall during the model period was not taken into account.

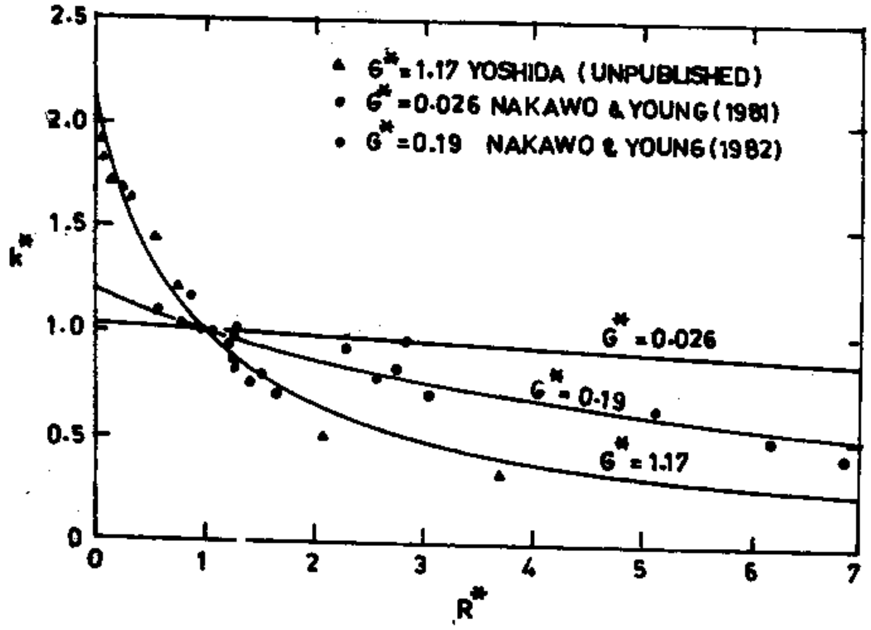

Fig.10: K*VS.R*. Solid lines represent predictions for particular values of G* corresponding to each experiment. R* for Yoshida's data was calculated by $R^* \approx h/h_c$, i.e. assuming K_m constant for every thickness.

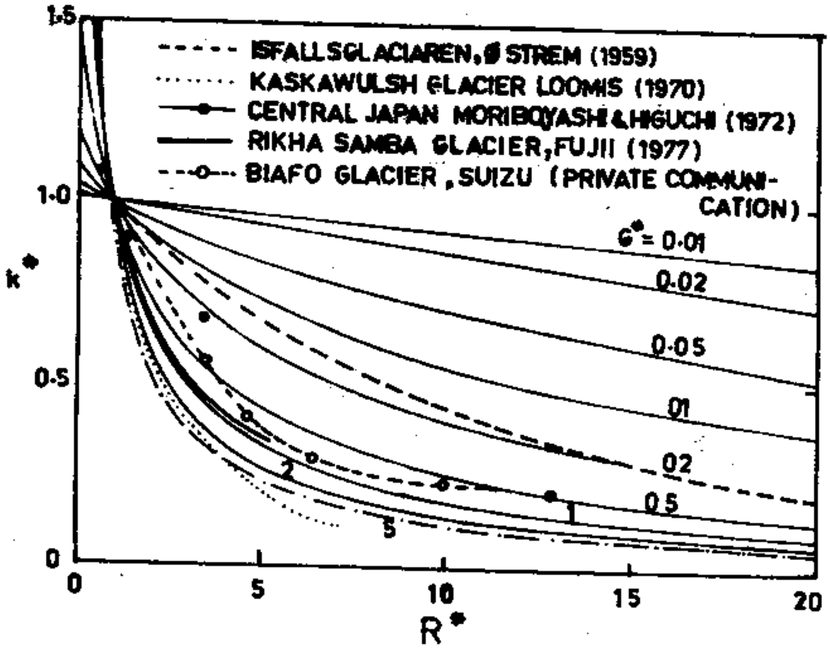

Fig.11: K*VS.R*. Predictions are shown with fine solid lines for various values of G*. R* for all the experimental data was calculated by $R^* \approx h/h$ assuming k_m to be independent on h.

2.3 Indian Studies

In India the glacier studies were initiated in 1947 under the guidence of Dr. J.E. Church, a meteorologist from Agriculture Experiment Station, Reno, Nevada, USA and then President of International Commission on Snow and Glaciers. Presently, several organizations such as Geological Survey of India, Snow and Avalanche Study Establishment, National Institute of Hydrology, India Meteorological Department, Wadia Institute of Himalayan Geology, Central Water Commission, Physical Research Laboratory, Space Application Centre, Defence Terrain Research Laboratory, Survey of India, and Jawahar Lal Nehru University have been involved in the glacier studies. GSI is monitoring a number of Himalayan Glaciers for the period more than a decade. HILTECH is also taking an active part in exploring the field of glacier hydrology and had organized an expedition to Gangotri Glacier in 1971 (Vohra et al, 1971). DST has also organised several multidiscip-linary expeditions to Chhota Shigri Glacier (H.P.) conti-nuously since 1985.

Most of the studies carried out in our country are concerning with glacial geomorphology, glacial dynamics, glacial flow characteristics etc. Very little efforts have been made to study the glaciers from water resources point of view.

The investigation of melt water runoff of any glacierized area involves a complex pattern of studies, as this runoff is controlled by numerous factors. The problem becomes more complex when applied to glacierized areas in the Himalaya, not only because of factors which affect melting surface, exposure, altitude, area distribution and albedo, but also for the fact that very little meteorological information is available. In most cases no climatological station exists in the glacierized area or nearby and wherever, it does exist, it deal only with recording of the precipitation in the form of rain and very seldom of the snowfall in winter. Consequently, not even a single model for the glacial melt-runoff simulation or forecasting has been developed.

The studies to measure the glacier melt water by CSI were initiated in Gara Glacier basin in 1973 (Vohra et al, 1974). Bhakra Management Board (B.M.B.) established a gauging station - a weir across the valley - at an altitude of 4400 m.a.s.l. about 2 Km downstream of the Gara glacier snout in 1974. Further B.M.B. and GSI made joint efforts to collect the discharge in the years of 1975 and 1976 (Raina et.al, 1977). In 1974 during the same expedition IMD installed a meteorological observatory close to the snout of the glacier at an altitude of 4700 m.a.s.l. . The meteorological data including maximum and minimum temperature, wind velocity, global radiation, albedo, precipitation were collected covering the period of 30-45 days between late July to early September.

During the 1986 expedition to the Chhota Shigri Glacier, NIH measured the discharge of melt stream for the duration 19th August to 12th September, 1986. The gauging site was located at 3831 ma.s.l. which was about 1.25 Km away from snout. The maximum discharge was recorded in the month of August. The details of the study have been reported by Singh (1987). IMD observed the meteorological parameters for the expedition duration.

Recently NIH, CWC and JNU participated in the July/Aug. 1987 expedition to the same glacier for the hydrological studies. Various techniques such as floats, current-meter and salt dilution were attempted to measure the discharge at the same site which was selected in 1986 expedition by NIH. The stage-discharge relationship was established for the gauging site. The hourly observations were also carried out for 48 hrs continuously to study the diurnal variation in the melt run-off. Data on snow density, depth of seasonal snow in accumulation zone was collected by NIH. The suspended sediment samples to estimate the sediment transport were collected by NIH, CWC and JNU. Other required data for hydrological modelling such as solar radiation, air temperature, wind velocity, albedo, sun shine duration etc. were collected by the IMD team in the accumulation zone and at the base camp.

Nijampurker and Bhandari (1977) dated the glaciers with environmental isotopes. The possibility to study the receding behaviour by isotopic techniques was suggested. A review of isotopic technique for snow and glacier hydrology has been made by Bahadur (1983), and Jain and Navada (1983). Nijampurkar and Bhandari (1983) has also reported the investigations of Himalayan glaciers using radioactive and stable isotopes. It was reported that study of radioactive tracers can provide quantitative estimates in understanding behaviour of glaciers in the past. In the 1987 expedition to Chhota Shigri Glacier, samples of snow and ice were also collected by Nijampurker.

Dhanju (1987) has carried out the study of reflectance parameters of snow and ice features using field radiometer in the visible and near infrared range of the electromagnetic spectrum. A sequential change in the reflectance pattern from the snow to glacier ice has been found. It is also reported that debris play an important role in modifying the values of reflectance. Such experiments were again conducted by Dhanju for the 1987 expedition to Chhota Shigri Glacier. In the Himalayas the only attempt for augmenting ice and snowmelt by coal dusting was made by Raina et al (1976).

3.0 REMARKS

The accumulation of ice mass on most glaciers is primarily due to snowfall, but a few glaciers are also nourished by avalanches, wind blown snow or by refrozen melt water. International Hydrological Decade (IHD:1965-1974) progam created help in the understanding of glacier distribution and fluctuations, seasonal snow line, and meso scale weathers etc. A little progress has also been made in a global integration and analysis of these data.

The physics of radiation and heat exchange between the atmosphere and glacier surface and the relative importance of the various components of the energy of balance are to be understood for debris covered and debris free glacier surfaces. A thorough understanding of the relation of heat and ice balance at glacier surface requires data on melt at much shorter time intervals (hourly for instance). But the inaccessibility of several areas in glacier region and difficulties inherent in continuous winter time measurement over the complete glacier area, have seriously restricted the study of glacier hydrology. Efforts are needed to increase the meteorological and hydrological observations for longer duration.

Computations of glacier discharge require an understanding of two separate processes melt water production on the glacier surface and drainage processes within the glacier Several attempts have been made in the past to describe the meltwater flow within the glaciers, but knowledge of internal drainage is still very limited. One reason for this may be the practical difficulties involved, as the main parts of drainage courses can be only studied indirectly. The major runoff delay facators appear to be the time of season, thickness of ice, and the distance of travel. More experiments are needed to make a definite conclusion regarding delaying processes. However, a little success has been achieved in measuring delay time by spreading relatively large amounts of slightly diluted dye over the considerable area of glacier surface with a sprinkling can.

The possibility of increasing glacier melting by means of blackening galcier surface to get additional amounts of melt water has been investigated by various investigations. Hardly any study has been conducted in our country considering the water resources aspect. The feasibility of artificial melting to cope with paucity of water in certain seasons has to tested using large basins.

The numerical models are also to be developed to simulate and forecast the glacier melt contribution in the glacier fed rivers. No such model is available in the literature for Himalayan glaciers. The selected glaciers are to be monitored for the complete melt season collecting meteorological and hydrological data on micro scale. Some index methods are also yet to be evolved for glacial melt computation.

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