

SNOW ACCUMULATION STUDIES IN
THE CENTRAL KARAKORAM, PAKISTAN

Cameron Wake

Wilfrid Laurier University
Waterloo, Ontario

ABSTRACT

Runoff from the Upper Indus Basin depends upon snowfall at high elevations in the northern mountains of Pakistan. However, very little is known concerning the amount of snowfall in the major source area, the high Karakoram, or its spatial and temporal distribution.

Glaciers contain well preserved records of past snowfall and provide a unique tool for retrieving data. An understanding of the seasonality of inputs in the source area can be obtained by interpreting a combination of physical and chemical properties of the snowpack. This in turn can be used to determine the seasonal and annual snowpack stratigraphy.

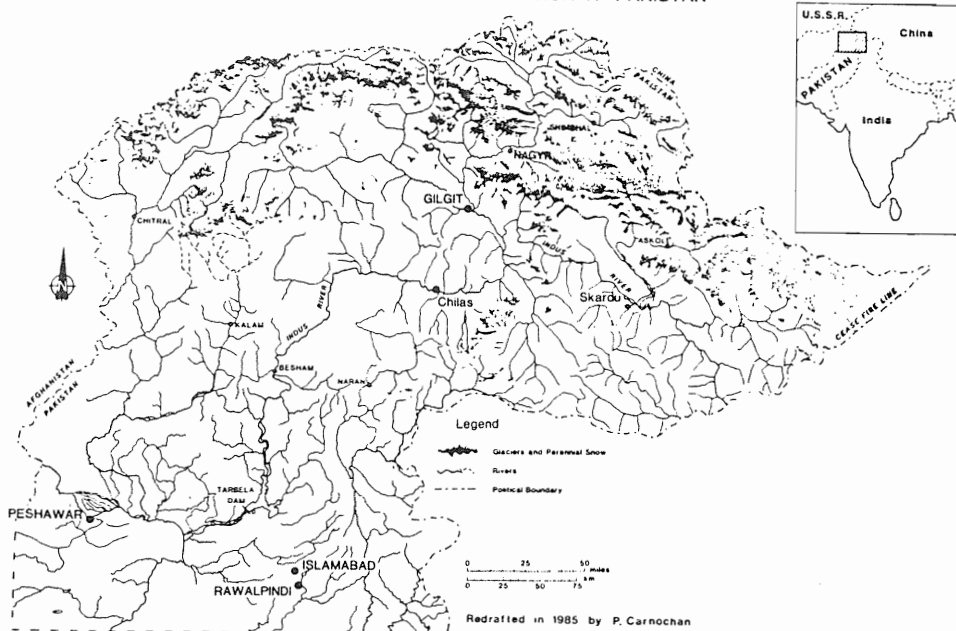
Field work in the Central Karakoram during the summer of 1986 consisted of a physical analysis of, and sample collection from seven snowpits, one snow/firn core and seven fresh snowfall events. Snowmelt samples have been analysed for ion concentrations (including Na, Cl, NO₂, NO₃, PO₄, and SO₄) and oxygen isotope ratios. A limited number of samples were analysed for microparticle concentrations, conductivity and total beta activity.

Preliminary results indicate that both the net annual accumulation and the source of moisture can be confidently determined from the physical and chemical records obtained.

I. INTRODUCTION

Pakistan depends heavily upon the waters of the Upper Indus Basin (U.I.B.) for power generation, irrigation and water supply. Runoff from the U.I.B. depends primarily upon snowfall at higher elevations along the northern portion of the watershed. The tremendous arc of the Karakoram Mountains, which extends over 350 km, holds the greatest concentration of snow and ice in Asia (Hewitt 1985; Figure 1).

FIGURE 1 GLACIERS AND DRAINAGE IN NORTH PAKISTAN



On average, 70-80% of the total annual runoff from the U.I.B. originates as snow and ice melt in the Karakoram. The bulk of this meltwater runoff occurs during the summer season from mid-June to mid-September. The overwhelming role the Karakoram plays in the hydrology of the U.I.B. draws attention to the snow conditions in this mountain range.

Unfortunately very little is known concerning the mountainous headwaters of the Indus. A concerted research effort concentrating on the glacial hydrology of the U.I.B. was begun in 1985 as part of the Snow and Ice Hydrology Project (S.I.H.P.). This is a collaborative project funded jointly by the Canadian International Development Research Centre, the Water and Power Development Authority in Pakistan, and Wilfrid Laurier University, Waterloo, Canada. The main goal of the S.I.H.P. is to improve our understanding of glacial conditions in the U.I.B. with a view towards developing a monitoring and forecasting network so that Pakistan can better manage its limited water resources.

The present study is concerned with moisture input into the region, which represents one crucial component of the hydrological system in the U.I.B. Most of the precipitation in the Karakoram occurs at elevations above 3000 m.a.s.l. (above sea level). It is this precipitation, in the form of snow, that creates the only large moisture surplus for the region. However, precipitation records for the U.I.B. come from weather stations which lie below 3000 m. mostly in semi-arid and valley locations. The stations are in the main towns and subject to the powerful topoclimatic effects of valley wind systems (Butz and Hewitt, 1986).

Fortunately, accumulation zones of glaciers contain well preserved records of snowfall and provide a unique tool for retrieving data. A variety of chemical and physical properties have been used by various investigators in order to identify seasonal and annual snowpack stratigraphy in tropical and temperate glaciers. The locations and techniques employed in these studies are listed in Table 1.

TABLE 1 SUMMARY OF ACCUMULATION STUDIES ON TROPICAL AND TEMPERATE GLACIERS EMPLOYING GLACIOCHEMICAL DATING TECHNIQUES

Reference	SITE CHARACTERISTICS			CHEMICAL PROPERTIES				PHYSICAL PROPERTIES		LOCATION
	elev (m)	depth (m)	acc. rate(m)	stable isotop	ion conc	cond	beta	trit	part	
Miller (1965)	6160	20	1.70						x	Khumbu Glacier, Nepal
Ambach (1976)	?	12	1.80					x	x	Kesselwandferner, Austrian Alps
Oerter (1982)	3150	81.45	0.80	x					x	Veinagferner, Austrian Alps
Jouzel (1977)	4785	16	2.80	x				x		Mont Blanc, French Alps
Oeschger (1977)	4450	55.65	0.32	x		x			x	Monte Rosa, Swiss Alps
Hoerberli (1983)	4450	4	0.52	x		x			x	Monte Rosa, Swiss Alps
Butler (1983)	2720	7	2.00		x					Athabasca Glacier, Canadian Rockies
Mayewski (1984)	4908	17	0.62	x	x			x		Sentik Glacier, Ladakh Himalaya
Lyons (1985)	5670	3	—		x			x		Quelccaya Ice Cap, Peru
Spencer (1985)	2450	3	—		x					Heard Island, South Indian Ocean
Thompson (1979)	5650	15	?	x			x		x	Quelccaya Ice Cap, Peruvian Andes
Thompson (1986)	5650	155.166	?	x		x			x	Quelccaya Ice Cap, Peruvian Andes
Holdsworth (1984)	5340	27	0.39	x			x		x	Mt. Logan, Yukon Territory

These studies suggest that the chemical composition of snow reflects changes in atmospheric chemistry. Therefore chemical records from areas of snow accumulation provide valuable data concerning atmospheric circulation and its seasonal variations. Variations in microparticle concentrations and the stratigraphic position of debris bands and thick ice layers have also proved useful in determining net annual accumulation. (Hattersley-Smith, 1971; Haeberli et al., 1983; Thompson, 1979, 1986). These variations should delineate seasonal and annual snowpack stratigraphy through an understanding of the seasonality of these inputs in the source area.

The goal of this research is to delineate annual snowpack stratigraphy and determine the rate of snow accumulation in the Biafo Glacier Basin. This is done by interpreting the variation with depth of the physical characteristics (density, position of ice layers and debris band, and microparticle concentrations) and chemical content (sodium, chloride, nitrate, sulfate, oxygen isotopes, and total β activity) of the snowpack, at several locations over a range of elevations. This information will then be used to determine the temporal, spatial and altitudinal variations of annual snow accumulation. From such data it is possible to:

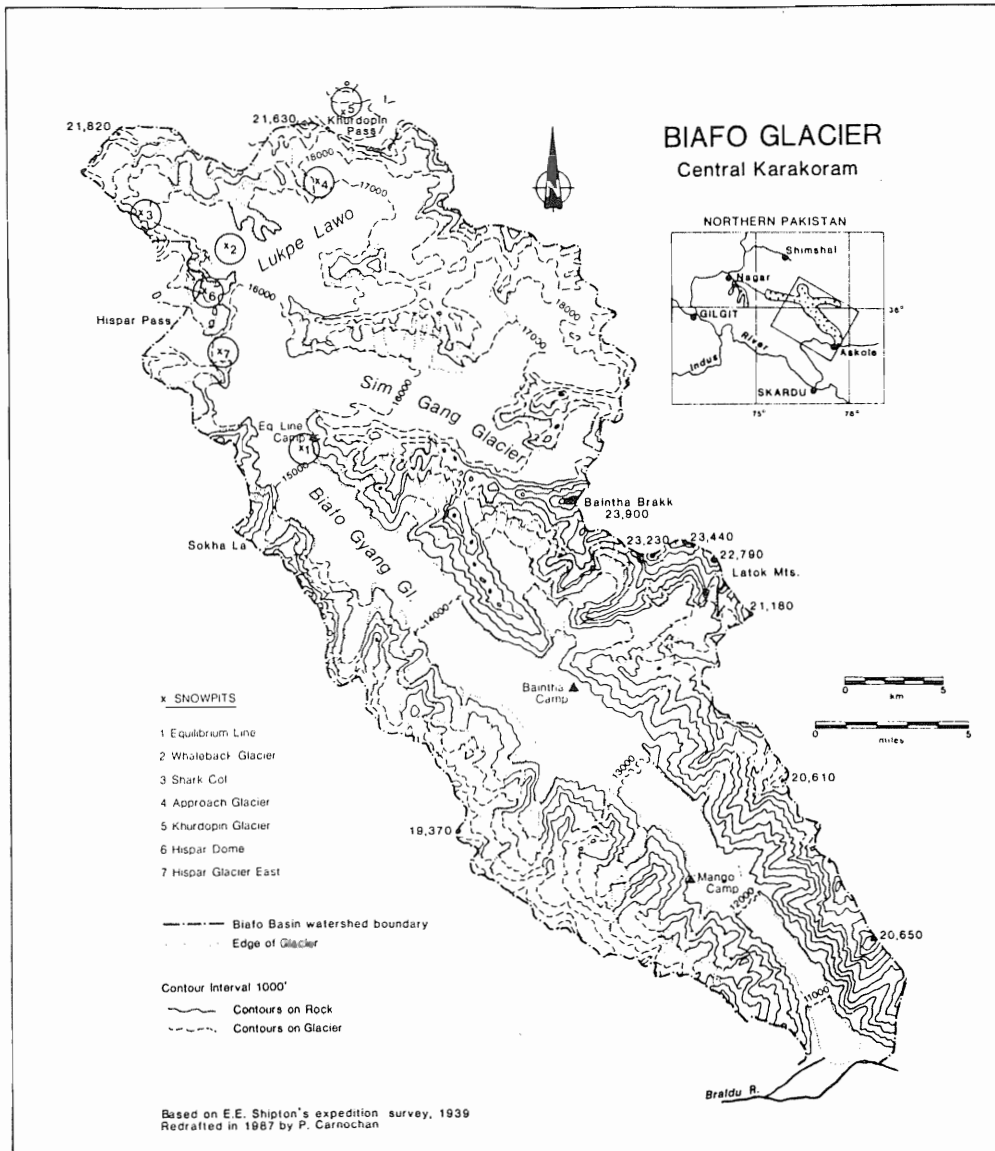
1. provide a first estimate of annual moisture input to the Biafo Glacier Basin;
2. determine the short-term response of the hydrological system to variations in summer and winter snow accumulation;
3. provide basic data concerning moisture input to calibrate hydrological forecasting models (Ferguson, 1986; Quick, pers. comm.);
4. determine the provenance of moisture and its seasonal variation and;

5. provide an input, from a data-sparse area, to a global data set developed to improve our understanding of atmospheric circulation and the Earth's geochemical cycle.

II. GEOGRAPHY AND METEOROLOGY

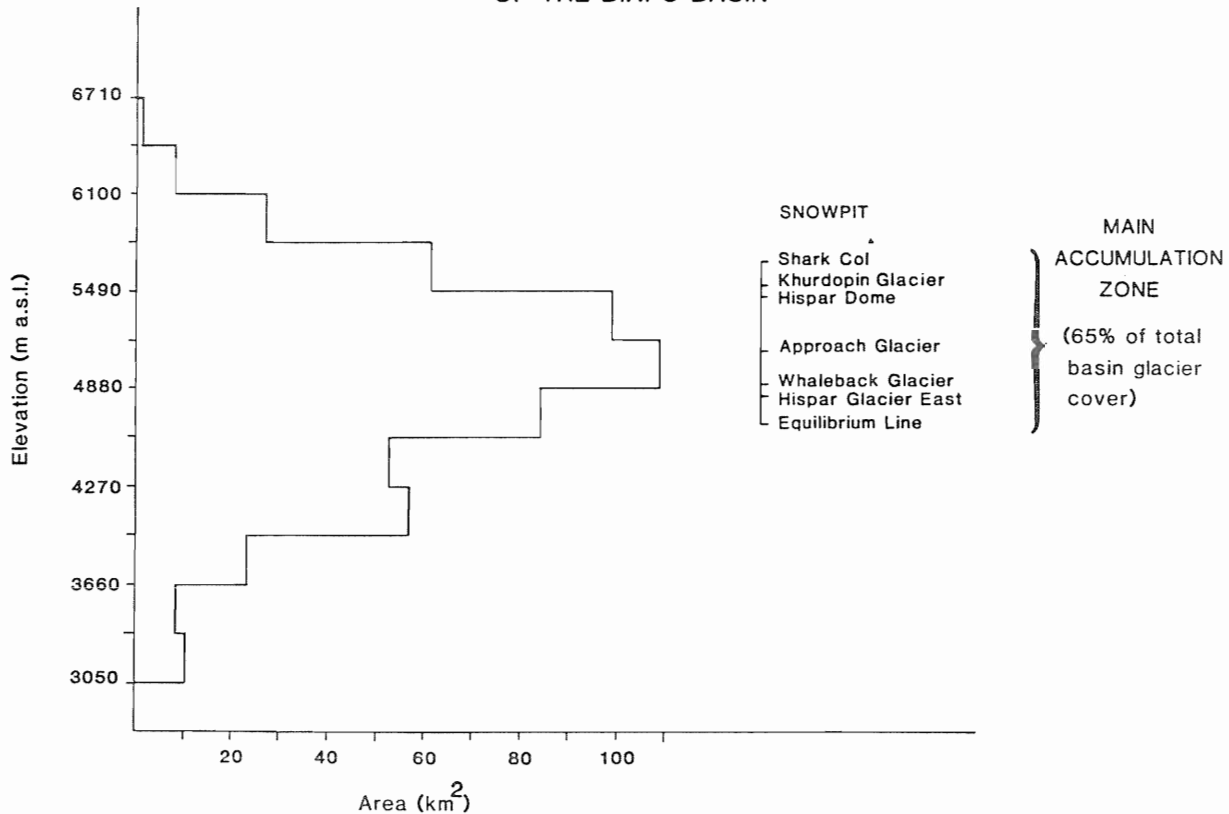
Following the reconnaissance during the 1985 field season in the Biafo Glacier Basin of the Central Karakoram, this area was chosen as a suitable location to dig a series of snow pits over a range of elevations and geographical locations. There exists a reliable base map of the Biafo-Hispar glacial region produced by the Royal Geographical Society (Mott, 1959). A topographic map of the Biafo basin, redrawn from the original Shipton survey maps, appears in Figure 2.

FIGURE 2



The Biafo Glacier Basin covers an area of 852 square kilometers, of which 64% is covered by glaciers or perennial snow and ice. The Biafo Glacier itself is a large valley glacier stretching over 59 kilometers. Figure 3 illustrates the area-altitude relationships for the glacier-covered area of the Biafo basin.

FIGURE 3 HYPSONETRIC CURVE FOR CONNECTED GLACIATED AREA OF THE BIAFO BASIN



The glacier flows from a broad, gently sloping, accumulation zone consisting predominantly of two separate accumulation basins, the Sim Gang Glacier and Lukpe Lawo (also referred to as "Snow Lake"). The glacierized portion of the accumulation zone covers an area of 370 square kilometers and accounts for over two-thirds of the total basin glacier cover. The accumulation basins are separated by steep narrow ridges which occupy a small percentage of the area in the accumulation zone. While avalanches of snow and ice descend onto the Sim Gang Glacier and Lukpe Lawo from the surrounding steep walls, the area represented by elevations at which avalanching predominates accounts for less than 15% of the accumulation zone. The majority of precipitation falls onto the broad, relatively low elevation surface of the accumulation zone, which accounts for an overwhelming 85% of the area in the accumulation zone and 2/3 of total basin glacier cover (Figure 3). Direct precipitation is the predominant form of nourishment. In this regard, the Biafo Glacier appears to be the exception in relation to other valley glaciers in the Central Karakoram. However, because of the minimal influence of avalanches on overall accumulation rates, the basin does provide an ideal location for determining the altitudinal, lateral and temporal distribution of snow accumulation.

Both the winter and summer seasons in the Central Karakoram are dominated by the Westerly air masses, from which the bulk of snowfall is derived (Boucher, 1975). The winter season represents the period of maximum precipitation. Influences from monsoonal air masses are thought to be negligible. However, snowfall during the late summer of 1985 in the Lukpe Lawo region seems to have been an example of an influx of monsoonal storms. Following the

1959 German expedition to the Batura Mustagh, Finsterwalder (1960) reported that the Summer Indian Monsoon crossed the Himalayas in early July and resulted in heavy precipitation in the Hunza River Valley and raging snowstorms at higher elevations.

The Batura Investigation Group (1979) recorded a net winter accumulation for 1973 of 1,034 mm water equivalent (w.e.) at 4840 m on the Batura Glacier, which lies approximately 125 km to the north-west of, and across the Hunza River Valley from, the Biafo Glacier accumulation area. A stratigraphic analyses of ice layers in two different crevasse walls at 5000 m revealed an apparent net annual accumulation rate ranging from 1030-1250 mm w.e. over the previous 5 years.

Density measurements and a description of the stratigraphy in two 4 m snowpits dug during a pilot study in the accumulation area of the Biafo Glacier suggests an annual net accumulation of 1700 mm w.e. at the 5000 m level for 1983-84. It was not possible to accurately separate summer and winter precipitation (Hewitt et al., 1986).

III. METHODOLOGY AND PROCEDURE

Field Work

During the 1986 summer field season two Canadians and four high-altitude porters spent 2 months in the accumulation zone of the Biafo Glacier Basin. The snow accumulation field work was extremely successful due both to very good weather which dominated during our time above snowline, and to the combined effort of all party members.

Overall we examined and collected samples from 7 snowpits 5-10 m deep, 15 m of a snow/firn core drilled from the bottom of a 5 m pit and collected fresh snow samples throughout the summer (Table 2).

TABLE 2 SUMMARY OF SNOW PITS DUG IN THE CENTRAL KARAKORAM, SUMMER 1986

PIT NO.*	DATE	LOCATION	ELEVATION	ASPECT	DEPTH(m)
1	06/15	EQUILIBRIUM LINE	4650m/15,250'	SW	2.25
2	06/21	WHALEBACK GLACIER	3900m/16,070'	SE	4.50
3	06/26-27	SHARK COL	5660m/18,570'	E-W	7.75
4	07/4	APPROACH GLACIER	5100m/16,730'	E	4.80
5	07/5-6	KHURDOPIN GLACIER	5520m/18,100'	W	10.00
6	07/27-28	SNOW DOME (pit)	5450m/17,880'	-	5.00
		SNOW DOME (core)	" "	-	15.00
7	07/31	HISPAR GLACIER	4830m/15,840'	E	4.80

*pit numbers refer to location of snowpits on Figure 1.

The seven snowpits cover an elevation range of 1010 m, from 4650 to 5660 m (Figure 3). This elevation band accounts for almost 85% of the glacier-covered area in the accumulation zone. The elevation of each snowpit in relation to the hypsometric curve is shown in Figure 3. The geographic locations of the snowpits are illustrated in Figure 2. All but the Khurdopin Glacier site lie within the Biafo Glacier basin. The importance of selecting suitable sites at which to collect data cannot be underestimated. Obtaining a 'representative' sample in an area that covers 370 km², over a range of elevations from 4500 to 6500 m is, to say the least, difficult. The problem is compounded by the remote nature of the study area. Most of one's time is spent transporting logistical and scientific equipment to and from the various study sites, and therefore the number of snowpits that can be investigated is limited.

There exists strong local variation in accumulation due to aspect, local relief and the effects of avalanches. Sample sites can therefore not be chosen at random. Rather, it is through a familiarity of the physical processes that govern snow accumulation, and a familiarity with the region in general, that study sites are chosen in areas that are free from the redistribution of snow by avalanches and little affected by the local redistribution of snow by the wind.

In order to compare data among pit locations, some sites should have one common physical characteristic, such as elevation, aspect or location. The snowpack should, aside from the surface, remain below freezing for the entire year to limit the percolation of meltwater. Other things being equal, the higher the altitude the snowpit/core site is, the cooler the temperature. This increases the potential for preserving a chemical record unaltered due to meltwater percolation. Flow geometry is not a critical factor in the study of shallow snowpits.

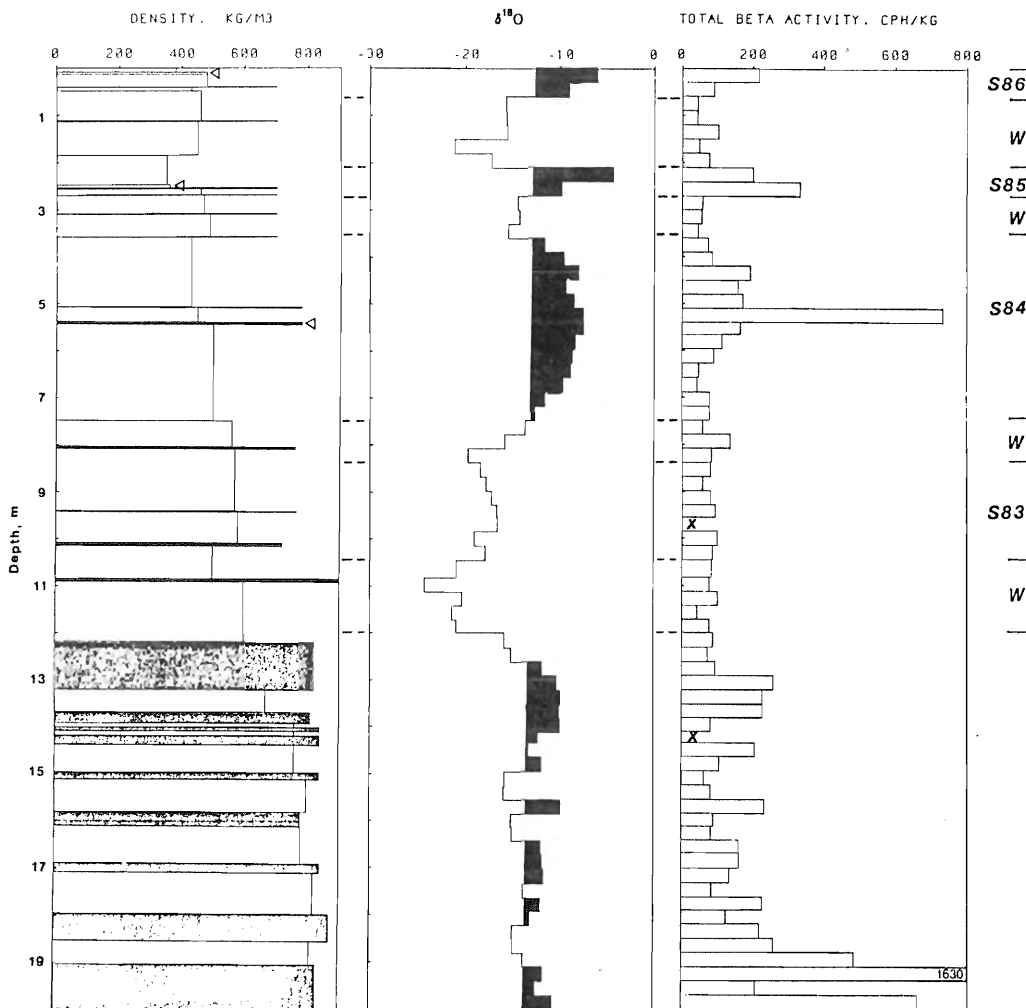
Snowpits were dug in 2 m intervals in order to simplify the sampling procedure. Snowpack stratigraphy was delineated on the basis of layer thickness, hardness, and colour, crystal size and shape, and the position of ice layers and debris bands. Density and stratigraphic layer was measured using a 24 cm stainless steel tube with a cross-sectional area of 41.7 cm². Following collection, the cored snow sample was placed in a nylon stuff sack and weighed with a 2 kg Homs spring scale. In six of the snowpits, snowmelt samples were collected over 15 cm intervals. The complete interval of 15 cm was sampled in order to obtain a complete chemical record of the snowpack. A 30 cm sample interval was used at the 'Whaleback Glacier' pit. Extreme care was exercised at all times during sample collection to ensure the collection of uncontaminated samples. All sample bottles were rinsed 3 times, allowed to stand full for at least 24 hours, and rinsed another three times with distilled, double deionized water that had subsequently been treated with a Milli-Q water purification system. A non-particulating clean suit in combination with an over the shoulder hood, particle mask and plastic gloves were worn by the sample collector. The north-facing snow-pit wall to be sampled was first scraped back approximately 30 cm with a lexan shovel and subsequently scraped clean with a plexiglass scraper. Samples were collected in 125 ml polypropylene wide-mouth containers with polyethylene caps. Following melting, the samples were transferred into 2-20 ml linear polyethylene (LPE) scintillation vials.

At the Hispar Dome site a 15 m core was recovered from the bottom of the 5 m pit by hand drilling with a PICO auger. Before drilling the core, samples were collected from the snowpit wall using the methods described above. Subsequently, thirty centimetre samples from the wall of the 5 m pit were scraped into ziplock bags using lexan scrapers. Thirty centimetre sections of the 15 m core were placed directly into ziplock bags after the stratigraphy had been analysed and the density measured. The resulting 68 samples in the ziplock bags were then melted and transferred into 20 ml scintillation vials and 500 ml LPE containers for isotope and total beta activity analyses, respectively. The core was not sampled for ion analysis. The borehole remained dry throughout the entire drilling period. Temperature at the bottom of the borehole (20 m) was -4.5°C. Fresh snow samples were collected as soon as possible following the termination of a precipitation event. The sample site was approached from a downwind direction in order to avoid contamination of the fresh snow. The sampler wore a gortex suit, non-particulating clean hood, particle mask and plastic gloves. Samples were collected in 125 ml cups and were subsequently processed in the same manner as the samples obtained from snowpit walls. Special care was taken to sample the complete depth of the fresh snow without sampling any of the underlying snow. In a few cases, 125 ml containers were opened and placed upwind from our camp at the beginning of the snow/rainfall event and left open for the duration of the storm. These were collected shortly after the end of the storm. Overall 23 samples of fresh snow/rainfall were collected from seven different precipitation events. One suite of seven fresh snow samples were collected from the same storm over an elevation range from 5105 m to 5450 m. All sample containers shipped to Canada had their tops taped shut prior to their departure from the field to prevent leakage.

An important consideration in all snow chemistry work is sample alteration due to natural postdepositional processes. The main natural source of change is melting in situ. Maximum recorded temperatures at 4900 m and 5450 m were as high as 20°C and 11°C, respectively. These high daytime temperatures and the existence of several ice layers, comprising from 6%-10% of the water equivalent of the snowpits, indicates that melting events do occur. Experiments in the field and laboratory in Norway (Johannessen and Henriksen, 1978) indicate that impurities in the snowpack tend to become concentrated in the initial meltwater; 50-80% of the impurities leave the snowpack in the first 30% of the melt. Davies and others (1982) suggest that elution of some ions from snow and firn occurs more rapidly than others. Their ranking, from high to low mobility, is: SO₄ > NO₃ > Na and Cl. Therefore the fractionation process responsible for the removal of impurities from the snowpack is also a differential process; some ions are removed more quickly than others.

Interpretation of glaciochemical data recovered from snowpits in which there has been melting, such as in the Karakoram, relies upon the assumption that the redistribution of ions due to meltwater percolation is limited to the uppermost stratigraphic layers. This seems reasonable for two reasons. First, the snowpack temperature drops below freezing slightly beneath the air-snow interface. Meltwater percolating down through the snowpack would theoretically freeze and form an ice layer upon encountering a below-freezing layer. The presence of ice layers in the top 50 cm of all the snowpits analysed suggests that this is the case. The ice layers thus formed would then act as a barrier to further meltwater percolation. Meltwater percolation would therefore result in local mixing, but, considering the amount of snowfall each year, an insufficient redistribution of ions to disrupt the seasonal chemical signatures. Second, the chemical record from a snowpack in which there has been significant meltwater percolation resulting in the downward migration of ions should: (1) appear "washed out" and show very little variation of chemical concentrations in the "summer" portion of the snowpack (Mayewski and others, 1981); and (2) display distinct chemical concentration maxima in or bordering ice layers. This is not the case for any of the chemical records recovered from the snowpits in the Central Karakoram. (Figures 4A and B).

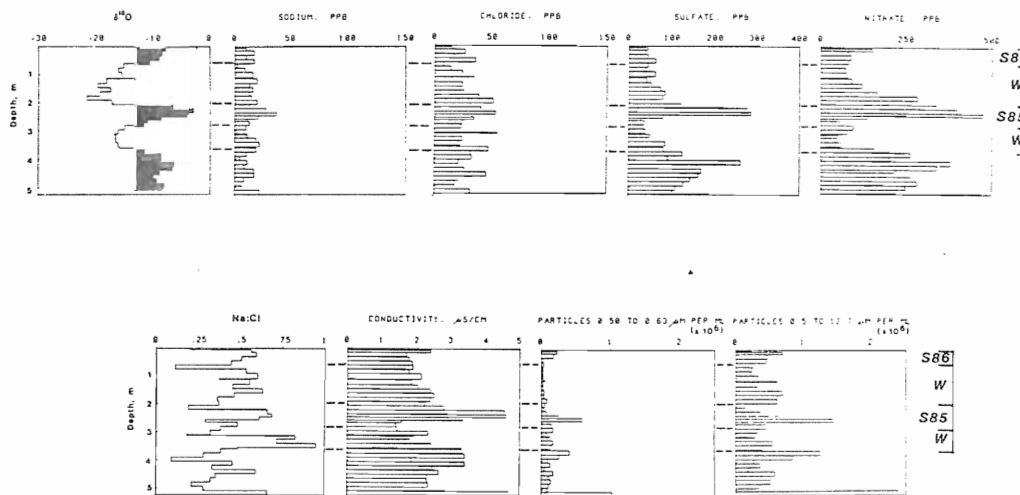
FIGURE 4 (A) HISPARDOME 5450 m



◁ Open triangles identify dirty horizons

FIGURE 4 (B)

HISPAR DOME 5450 m



Laboratory Work

Over 375 snowmelt samples were collected and returned for chemical and microparticle analyses. Anions (Chloride, sulfate and nitrate) have been analysed using a Dionex model 2010 ion chromatograph with an AS-4 column and 0.0021M NaHCO_3 /0.0017M Na_2CO_3 eluent and a computer-driven autosampler. The detection limit for chloride, sulfate and nitrate is 0.5, 1.0 and 1.0 ppb (parts per billion). Ten percent of these samples were analysed in duplicate using two separately drawn samples. This procedure resulted in a relative deviation from the mean of 2% for chloride, nitrate and sulfate. (Precision calculations after Skoog and West, 1976). Sodium was analysed by stabilized-temperature furnace atomic absorption spectrometry using a Perkin Elmer model 2280 equipped with a model 400 furnace. The detection limit for sodium, using this method, is 0.5 ppb. All of these samples were analysed in duplicate, resulting in a relative deviation from the mean of 3%. Both the anion and cation analyses were completed using the facilities of the Glacier Research Group at the University of New Hampshire. The oxygen isotope analyses were done by the Geophysical Isotope Laboratory at the University of Copenhagen using gas source mass spectrometry. All of these samples were analysed twice. The relative deviation from the mean was 0.3%.

Forty samples from Hispar Dome 5 m snowpit were also analysed for total conductivity and for microparticle concentrations in 15 different size ranges using a Coulter Counter TA II, by the Institute of Polar Studies at Ohio State University. Sixty-eight samples from the combined 5 m snowpit and 15 m snow/firn core from the Hispar Dome were analysed for oxygen isotope ratios and total β activity. The total β activity was measured at the University of New Hampshire, using a Canberra model 2404 alpha-beta-gamma counter with a 12-hour counting period for 47 cm Whatman SA-1 cation filters through which 500 ml samples had been gravity filtered twice.

As described below, two sets of field blanks, one set of transfer duplicates and one set of laboratory blanks were made to determine if snow samples collected in the Karakoram had been contaminated in any way during the sampling, transport and laboratory analysis phases of this project. Before leaving for the field, six 20 ml scintillation vials were filled with Milli-Q water at the University of New Hampshire. The blanks, termed 'UNH blanks', travelled with all of the other samples and remained sealed throughout the field program. As previously mentioned, samples were collected in 125 ml cups and, after melting, were transferred into 2-20 ml scintillation vials. Seven 'transfer blanks' were made by transferring

Milli-Q water directly into the scintillation vials in the field at the same time the snow samples were transferred. The results appear in Table 3.

TABLE 3 NET ANNUAL ACCUMULATION AT HISPAR DOME (5450 m)

SEASON	SNOWPACK DEPTH(m)	W.E. (m)	% OF ANNUAL ACC. (m)	NET ANNUAL ACC. (m)
S86	0-0.70	0.32+	27%	1.20
W85-86	0.70-2.05	0.88	73%	
S85	2.05-2.70	0.22	34%	0.65
W84-85	2.70-3.60	0.43	66%	
S84	3.60-7.50	1.95	78%	2.51
W83-84	7.50-8.40	0.56	22%	
S83	8.40-10.50	1.10	52%	2.12
W82-83	10.50-12.20	1.02	48%	

Six 'transfer duplicates' were made by transferring snow samples into a third 20 ml scintillation vial to test for ion contamination. The analytical results of this set of duplicates is listed in Table 4.

TABLE 4 VARIATION IN ANNUAL DISCHARGE OF THE UPPER INDUS BASIN

STATION*	MEAN ANNUAL DISCHARGE (1000 A-ft.)	PERCENTAGE OF MEAN ANNUAL DISCHARGE																			
		1966	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81	82	83	84	85
3 Indus R at Katchura	25,210	—	—	—	—	88	86	82	131	85	94	91	99	120	81	121	107	94	118	115	98
6 Gilgit R at Alam Br	16,610	122	127	—	119	114	104	85	122	77	77	81	90	114	100	93	—	74	87	106	—
7 Indus R at Portab Br	43,160	104	110	102	100	101	100	88	140	87	105	100	97	115	90	96	102	87	100	104	94

*Station Number refers to Figure 2.1.

The results shown in Tables 3 and 4 indicate that, on the whole, the transfer process and the transport of samples introduced very little contamination.

One out of every ten samples analysed by ion chromatography and atomic absorption spectrometry was a sample of Milli-Q water. Throughout the laboratory procedure, these Milli-Q samples showed no signs of contamination indicating that the laboratory procedure did not contaminate the samples.

IV. PRELIMINARY RESULTS

Although not fully analysed yet, sufficient work has been done with the Hispar Dome records from the combined 5 m snowpit and 15 m snow/firn core to allow for a preliminary interpretation. Similar and straightforward seasonal variations are evident in most of the time-series profiles (Figure 4, A & B). The seasonal stratigraphy for the snowpack at the Hispar Dome is delineated on the right-hand side of Figure 4. The net annual accumulation at Hispar Dome is listed in Table 3. Distinct trends in the oxygen isotope profile have been used as reference horizons to define summer and winter strata. The oxygen isotope profile shows strong seasonal variation for the first three summers. The record of ion concentrations, microparticle concentrations and conductivity extends only to 5.25 m below the surface. The sulfate and nitrate profiles are characterized by strong, broad peaks which occur in mid to late summer. The chloride and sodium profiles show only a weak seasonal variation. Peaks tend to occur in mid to late summer, however solitary peaks are also evident in the winter portion of the profile. Peaks in the conductivity profile also occur in summer, but are generally narrower, and not nearly as strong as the sulfate and nitrate peaks. Peak concentrations in both the small (0.50-0.63µm) microparticles occur early in the summer; the small size particles display stronger peaks and a more distinct seasonal variation.

The total β activity profile displays distinct peaks in early to mid summer, for the first three summers. The peak values in the total β activity profile coincide with horizons containing visible dirt. The correlation coefficient between total β activity and total microparticle concentration is 0.72 at the 0.001 significance level. Unfortunately, the microparticle data stops at 5.25 m below the top of the snow pit, and therefore a concrete relationship between the microparticle concentrations and total β activity cannot be established. Only the physical, total β activity and oxygen isotope records are available below 5.25 m. The summer strata in 1984 is clearly identified by a sharp peak in the total β activity and a strong, broad peak in the oxygen isotope profiles. The interpretation of the chemical profiles below this is uncertain. The very weak peak in the oxygen isotope profile extending from 8.4 to 10.5 m can be interpreted as summer strata bordered above and below by weak winter troughs. This interpretation results in net annual accumulation of 2.51 and 2.12 m water equivalent (w.e.) for 1983-84 and 1982-83, respectively. On the other hand the trough extending from 7.5 to 12.2 m could represent one winter's worth of precipitation, resulting in net annual accumulation of 4.63 m w.e. for 1983-84. This value seems very high when compared to the net annual accumulation of 1.20 and 0.65 m w.e. for 1984-85 and 1985-86. The second interpretation is supported by the lack of any peak in the total β activity profiles. However, there is one sample missing from the β activity profile, and there is a conspicuous lack of a visible dirty horizon at this level in the snowpack. On the basis of consistency two annual layers are assigned to the snowpack from 3.6 to 12.2 m.

From 7 to 12 m below the surface, the seasonal variation in both the $\delta^{18}\text{O}$ and total β activity records from the core is reduced (Figure 4). Both records appear smoothed. Mass exchange by vapour diffusion can reduce the amplitude of seasonal variations. This is limited when the annual accumulation is greater than 0.25 m w.e. (Dansgaard et al., 1973), such as in the Central Karakoram. However, melting events do occur. Therefore the 'smoothing' of the chemical records in the core below 7 m is probably due to both the percolation of meltwater, predominantly within summer strata combined with compaction, and perhaps distortion, of the record with depth. Molecular diffusion in ice is a very slow process and would therefore not seriously effect shallow cores of this nature (Patterson, 1980).

A distinct transition occurs in the physical characteristics of the core at 12.2 m (Figure 4). An abrupt contact marks the change from fine grained granular snow with a density of 590 kg/m³, to a clear, icy layer containing enclosed air bubbles, with a density of 820 kg/m³. Below 12.2 m the density profile shows a distinct pattern of alternating high density, clear, icy layers (represented by the stippled pattern) and lower density, white granular layers. These characteristics suggest that meltwater has played a role in the relatively shallow transition of firn to ice. Perhaps this portion of the core represents a period of accelerated melting. Other studies in the wet snow zone of glaciers have identified abrupt firn ice transitions at similar depths. Sharp (1951) recognized a sharp transition of firn to ice at a depth of 13 m on the Upper Seward Glacier. Vallon et. al. (1976) found an abrupt firn ice transition at 32 m in the Vallee Blanche in the Alps. Meltwater speeds up the transformation process by increasing the settling rate and sintering of snow crystals (Patterson, 1981).

Of special interest in the total β activity profile (Figure 4) are the two sharp peaks; one at 5.10-5.40 m (733 dph/kg) and the other at 19.1-19.4 m (1630 dph/kg), below the snow surface. The first peak at 5.1 to 5.4 m is of the same magnitude as the maximum total β activity peak (639 dph/kg) found at 15.4-15.96 m in the core from Sentik Glacier in the Ladakh Himalaya. The peak in the total β activity profile from Sentik Glacier coincides with the 1963 thermo-nuclear test level. (Mayewski et. al., 1984). The peak at 5.1-5.4 m in the Hispar Dome Core is assigned to the summer of 1984. This peak does coincide with a dirty, ice layer suggesting that the total β activity has been concentrated by surface melting and subsequent adsorption of beta-active material onto dust particles (Prantl, 1973; Ambach, 1968).

The peak at 19.1-19.3 to extremely high. There appears to be no comparable peak in the Sentik Glacier core. This peak could be assigned to the Chinese nuclear weapons tests in the early 1970's or the early 1980's. Total β activity measurements of precipitation in Taiwan show a sharp peak in June of 1973, which is attributed to atmospheric nuclear weapon testing at Lop Nor (40 N; 90 W) in mainland China (Weng et. al., 1977). Assigning this peak to increased atmospheric levels of total β activity in 1973, results in a mean annual accumulation rate from 1973 to 1982 of \sim 0.73 m w.e. This is low in relation to the mean annual accumulation rate from 1982-1986 of 1.62 m w.e. If the chronology for the total β activity

peak at 19.1-19.4 m is correct, the data suggests that a portion of the record has been lost, presumably through intensive ablation at the Hispar Dome site. Assuming this period of intensive melting did occur, the resulting downward percolation of meltwater carrying with it radioactive material, provides an appealing mechanism for the concentration of beta active particles resulting in the large total β activity spike at 19.1-19.3 m.

High concentrations in the total β activity of air and precipitation samples were measured in 1981 by Health and Welfare Canada (Dr. Eaton, pers. comm.), presumably due to atmospheric nuclear weapons testing by the Chinese in October of 1980. Assigning the peak at 19.1-19.3 m to the winter of 1980 results in a mean annual accumulation rate of 2.90 m w.e. which is almost comparable to the rate of 2.51 m w.e. in 1983-84, but still much greater than the four year mean of 1.62 m w.e. This question remains unresolved.

The data from all sites for 1985-86 sites show that 30-50% of the net annual accumulation occurs during the summer. However, the 1986 net summer accumulation listed in Table 3 does not include accumulation from a large snowfall event that we know occurred in early August. This suggests that the relative contribution of summer accumulation is possibly greater than 50%. This supports our original feelings, after witnessing large snowfall events during the summer of 1985 and 1986, that snow accumulation during the summer can be considerable.

Net summer accumulation at the Hispar Dome site in 1983 and 1984 was 3 and 6 times net summer accumulation in 1986, respectively. It has been demonstrated that summer snowfall in glacierized basins substantially decreases the rate of melting for short periods of time (Young, 1977). The experience of the Snow and Ice Hydrology Project during 1985 & 1986 at gauging stations in the U.I.B. indicates a dramatic drop in discharge during, and for some time after, extended summer storms, even though these storms are accompanied by heavy rainfall in and below the ablation zones. Summer precipitation from storms that extend over a few or more days is therefore inversely proportional to discharge. The reduction in discharge is most likely due both to the increase in cloud cover accompanying summer storms which decreases the amount of energy reaching the surface and therefore decreases the rate of melting (Wake, 1985), and an increase in the surface albedo, predominantly in the upper portions of the ablation zone. This type of response to precipitation within the U.I.B. demonstrates that melting snow and ice provide the predominant control on the temporal variation of runoff.

With a significant increase in summer accumulation in 1983 and 1984, one might expect a corresponding decrease in annual discharge. The annual variation in discharge for three gauging stations in the U.I.B. are listed in Table 4. Contrary to the expected trend, all three stations show above average discharge in 1984; two of three stations show above average discharge for 1983. Therefore, as the summer of 1983 and 1984 show no reduction in annual runoff, it is unlikely that increased accumulation was caused by storms occurring throughout the summer. Rather, the high levels of summer accumulation were probably a result of an intensified period of precipitation, similar to our experience in August both 1985 & 1986. The sharp decrease in the runoff hydrographs from the gauging stations in the U.I.B. during the summer in 1983 and 1984 are indicative of intensive summer precipitation accompanied by extensive cloud cover. Therefore, data from 1983 through to 1986 suggests that increased summer accumulation is a result of low frequency, high magnitude precipitation events, on the seasonal scale. These storms result in a dramatic short-term decrease in discharge. A more concrete relationship could be established by first, determining the variation in runoff for 5-day periods with respect to the long-term mean, and second, by obtaining a longer term record of summer accumulation. In addition, a more complete chemical analysis of the snowpack would help identify the predominant source of moisture for these summer storms.

Finsterwalder (1960) stated that when the monsoon broke into the Karakoram in the summer of 1959 snowstorms, which resulted in substantial accumulation of snow at higher elevations, were accompanied by very heavy precipitation in the valleys. The summer precipitation records for Karimabad (36 19 N; 74 40 E) in the Hunza Valley appear in Table 5.

TABLE 5

KARIMABAD (2400 m)
SUMMER PRECIPITATION (JUNE-SEPT.)

YEAR	PPT (mm)	DEPARTURE*
1979	57	-18
1980	86	+11
1981	96	+21
1982	51	-24
1983	42	-33
1984	40	-35
1985	23	-52

The 1983 and 1984 summers show below average values for summer precipitation. This reinforces the point that precipitation measurements at valley bottom stations, such as that at Karimabad, provide little indication of conditions at higher elevations.

A rough indication of the strength of the summer monsoon in the north-west quadrant of the subcontinent is provided by the summer rainfall records from Srinagar (34 05 N; 74 50 E). This data originates from the 'Monthly Climatic Data for the World' (NOAA, 1982-1986), and appears in Table 6.

TABLE 6

SRINAGAR (1518 m)

HISPAR DOME (5450 m)

YEAR	SUMMER PPT (mm)	DEPARTURE*	SUMMER ACCUMULATION
1982	158	-33 (mm w.e.)	----
1983	229	+38	1100
1984	200	+9	1950
1985	109	-82	220
1986	142	-49	320

*mean=191 mm

The 1983 and 1984 summers show substantially more precipitation than the summers of 1985 and 1986. For the four summers from 1983-86, the variation in summer precipitation at Srinagar roughly corresponds to the variation in net annual accumulation at the Hispar Dome (Table 6). Concrete conclusions cannot be drawn due to the short time period represented by the data set. However, the data do suggest there is a relation between summer precipitation in Srinagar and summer accumulation in the Central Karakoram.

REFERENCES

- Ambach, W., Eisner, H., Elsasser, M., Loschhorn, U., Moser, H., Rauert, W., and Stichler, W. 1976. Deuterium, Tritium and Gross-Beta-Activity on Alpine Glaciers (Otztal Alps). Journal of Glaciology, Vol. 17, No. 77, pp. 383-400.
- Batura Glacier Investigation Group, 1979. The Batura Glacier in the Karakoram Mountains and its Variations. Scientia Sinica, Vol. 22, pp. 958-974.
- Boucher, K. 1975. Global Climate. New York, John Wiley and Sons, Inc.
- Butler, D., Lyons, W., Hassinger, J., and Mayewski, P. 1983. Shallow Core Snow Chemistry of Athabasca Glacier, Alberta. Canadian Journal of Earth Sciences, Vol. 17, No. 2, pp. 278-281.
- Butz, D. and Hewitt, K. 1986. A Note on the Upper Indus Basin Weather Stations, in K. Hewitt (ed), Snow and Ice Hydrology Project Annual Report and Scientific Papers, 1985. Wilfrid Laurier University, pp. 64-76.
- Dansgaard, W. 1964. Stable Isotopes in Precipitation. Tellus, V. 16, pp. 436-468.

- Davies, T.D., Vincent, C.E., and Brimblecombe, P. 1982. Preferential Elution of Strong Acids from a Norwegian Ice Cap. Nature, Vol. 300, pp. 161-163.
- Ferguson, R. 1986. Parametric Modelling of Daily and Seasonal Snowmelt using Snowpack Water Equivalent as well as Snowcovered Area. IAHS Publ. No. 155, pp. 151-165.
- Finsterwalder, R. 1960. German Glaciological and Geological Expedition to the Batura Mustagh and Rakaposhi Range. Journal of Glaciology. Vol. 3, No. 28, pp. 787-788.
- Haeberli, W., Schotterer, U., Wagenbach, D., Haeberli-Schwitler, H., and Bortenschlager, S. 1983. Accumulation Characteristics on a Cold, High Alpine Firn Saddle from a Snow-Pit Study on Colle Gnifetti, Monte Rosa, Swiss Alps. Journal of Glaciology, Vol. 29, No. 102, pp. 260-271.
- Hattersley-Smith, G., Krouse, H.R., and West, K.E. 1975. Oxygen Isotope Analysis in Accumulation Studies on an Ice Cap in Northern Ellesmere Island, NWT. IAHS Publ. No. 104, pp. 123-128.
- Hewitt, K. 1985. Snow and Ice Hydrology in Remote, High Mountain Basins: The Himalayan Sources of the Indus. Snow and Ice Hydrology Project Working Paper No. 1, 28 p.
- Hewitt, K., Wake, C., and Young, G. 1986. Snow Stratigraphy in the Accumulation Zones of the Biafo and Hispar Glaciers, in K. Hewitt (ed), Snow and Ice Hydrology Project Annual Report and Scientific Papers, 1985. Wilfrid Laurier University, pp. 170-179.
- Holdsworth, G., Pourchet, M., Prantl, F.A., and Meyerhof, D.P. 1984. Radioactivity Levels in a Firn Core from the Yukon Territory, Canada. Atmospheric Environment, Vol. 18, No. 2, pp. 461-466.
- Johannessen, M., and Henriksen, A. 1978. Chemistry of Snow Meltwater: Changes in Concentration During Melting. Water Resources Research, Vol. 14, No. 4, pp. 615-619.
- Jouzel, J., Merlivat, L., and Pourchet, M. 1977. Deuterium, Tritium, and Beta Activity in a Snow Core taken on the Summit of Mont Blanc (French Alps). Determination of the Accumulation Rate. Journal of Glaciology, Vol. 18, No. 80, pp. 465-470.
- Lyons, W.B. and Mayewski, P.A. 1983. Nitrate plus Nitrate Concentration in a Himalayan Ice Core. Geophysical Research Letters, Vol. 10, No. 12, pp. 1160-1163.
- Lyons, W.B., Mayewski, P.A., Thompson, L.G., and Allen, B. 1985. The Glaciochemistry of Snow-Pits from Quelccaya Ice Cap, Peru, 1982. Annals of Glaciology, Vol. 7, pp. 84-88.
- Macpherson, D.S. and Krouse, H.R. 1967. O18/O16 Ratios in Snow and Ice of the Hubbard and Kaskawulsh Glaciers. Isotope Techniques in the Hydrological Cycle, American Geophysical Union Monograph No. 11, pp. 180-194.
- Mayewski, P.A., Lyons, W.B., and Ahmad, N. 1981. Reconnaissance Glaciochemistry Studies in the Indian Himalaya. Proceedings of the Eastern Snow Conference, 38th Annual Meeting, pp.45-48.
- Mayewski, P.A., Lyons, W.B., and Ahmad, N. 1983. Chemical Composition of a High Altitude Fresh Snowfall in the Ladakh Himalayas. Geophysical Research Letters, Vol. 10, No. 1, pp. 105-108.
- Mayewski, P.A., Lyons, W.B., Ahmad, N., Smith, G., and Pourchet, M., 1984. Interpretation of the Chemical and Physical Time-Series Retrieved from Sentik Glacier, Ladakh Himalaya, India. Journal of Glaciology, Vol. 30, No. 104, pp. 66-76.
- Miller, M.M. 1965. Tritium in Mt. Everest Ice-Annual Glacier Accumulation and Climatology at Great Equatorial Altitudes. Journal of Geophysical Research, Vol. 70, No. 16, pp. 3885-3888.

- Mott, P.G. 1950. Karakoram Survey, 1939. A New Map. The Geographical Journal, Vol. 116, No. 1-3, pp. 89-95.
- NOAA 1982-1986. Monthly Climatic Data for the World. Vols. 35-39.1 National Climatic Data Centre: Asheville, North Carolina (also available at AES in Downsview, Ontario.)
- Oerter H., Reinwarth O., Rufli, H. 1982 Core Drilling Through a Temperate Alpine Glacier (Vernagtferner, Oetztal Alps) in 1979. Zeitschrift fur Gletscherkunde und Glazialgeologie. Vol. 18, pp. 1-11.
- Oeschger, H., Stauffer, B., Haeberli, W., and Rothlisberger, H. 1978. First Results from Alpine Core Drilling Projects. Zeitschrift fur Gletscherkunde und Glazialgeologie. Vol. 13, pp. 193-208.
- Raina, V.K., Kaul, M.K., and Singh, S. 1977. Mass-balance Studies of Gara Glacier. Journal of Glaciology, Vol. 18, No. 80, pp. 415-423.
- Spencer, M.J., Mayewski, P.A., Lyons, W.B. and Hendy, R.H. 1985. A preliminary assessment of the potential application of glaciochemical investigations on Heard Island, South Indian Ocean. Journal of Glaciology, Vol. 31, No. 109, pp. 233-236.
- Thompson, L.G., Hastenrath, S., Morales Arnao, B. 1979. Climatic Ice Core Records from the Tropical Quelccaya Ice Cap, Science, Vol. 203, pp. 1240-1243.
- Thompson, L.G., Mosely-Thompson, E., Dansgaard, W., and Grootes, P.M. 1986. The Little Ice Age as Recorded in the Stratigraphy of the Tropical Quelccaya Ice Cap. Science, Vol. 234, pp. 361-364.