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A STUDY OF SLOPEWASH PROCESSES IN THE  
CONTINUOUS PERMAFROST ZONE, BANKS ISLAND, WESTERN CANADIAN ARCTIC

by

Antoni G. Lewkowitz

Submitted to the School of Graduate Studies in Partial  
Fulfilment of the Doctor of Philosophy Degree in Geography.

University of Ottawa  
Ottawa, April 1981

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The present thesis describes three years of fieldwork undertaken on sloopwash processes, one year of which was presented as an M.A. thesis (Lewkowicz, 1978).

"Hill. Yes, that was it. But it is a hasty word for a thing that has stood here ever since this part of the world was shaped."

J.R.R. Tolkien (1954). The Two Towers.

CHAPTER 1

INTRODUCTION

## Background

The goal of this thesis is to partially redress the imbalance that exists between geomorphological research concerning azonal and "unique" aspects of the permafrost environment. The subject of study is a group of processes known collectively as slopewash, embodying (1) surface wash, the downslope transport of weathered material over the ground surface by running water, and (2) subsurface wash, the set of processes associated with water movement and sediment transport within the regolith (Young, 1972, pp. 62-70).

Slopewash processes were studied in the continuous permafrost zone, in the vicinity of the Thomsen River, Banks Island, Northwest Territories (Figures 1 and 2). The study area is located within the lowlands of central Banks Island (Figure 2) and is underlain by unlithified and poorly lithified Cretaceous rocks. It was not glaciated during the last "classical" Wisconsin glaciation, but was covered by ice during at least two pre-Wisconsin advances (Vincent, 1978a). Some slopes in the area, unaffected by meltwater discharges or lacustrine inundation associated with the last ice sheet, may have been subject to uninterrupted slopewash action since pre-Wisconsin times.

## Aims

The aims of the study were twofold.

First, the geomorphological aim was to assess the frequency and magnitude of slopewash processes in a permafrost area. The values obtained were to be compared with others from nonpermafrost regions. Suggestions have been made in the literature, that permafrost increases the probability of surface wash generation by acting as an impermeable subsurface barrier.

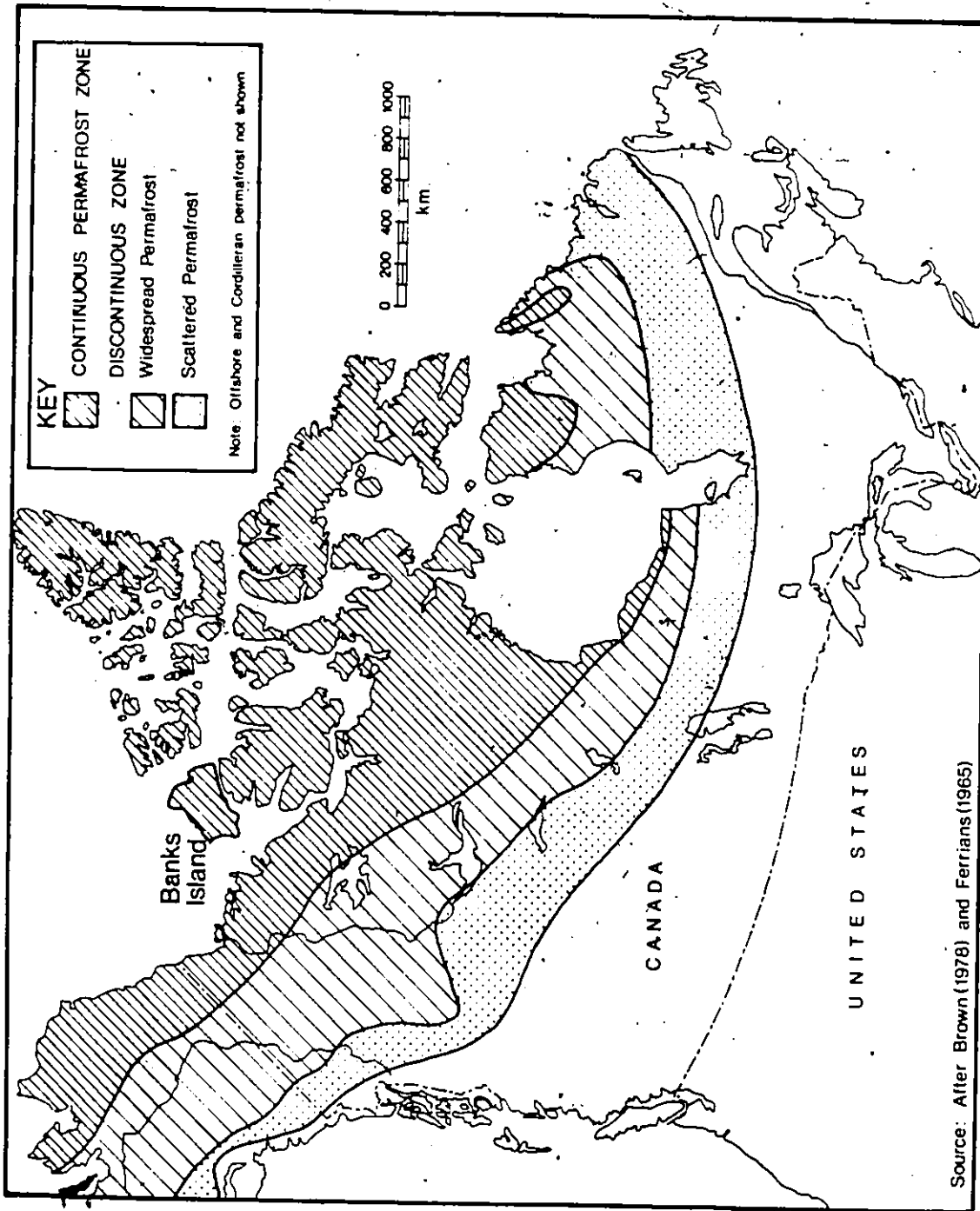


FIGURE 1 : LOCATION OF BANKS ISLAND WITHIN THE ZONE OF CONTINUOUS PERMAFROST

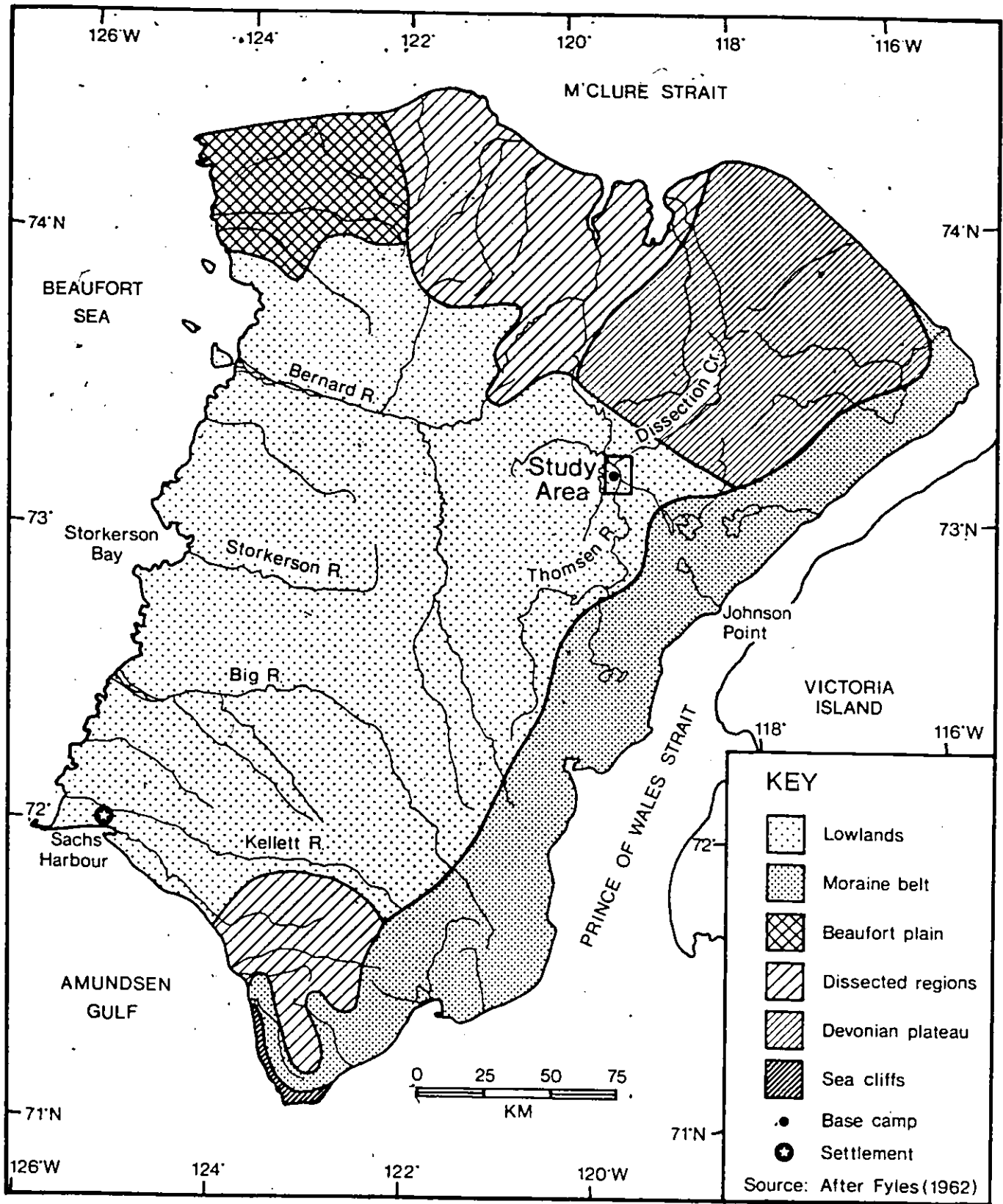


FIGURE 2 : LOCATION OF THE STUDY AREA WITHIN THE PHYSIOGRAPHIC REGIONS OF BANKS ISLAND

Furthermore, with higher surface runoff coefficients and the limited protection furnished by a sparse vegetation cover, slopewash becomes a particularly effective erosional agent in an Arctic tundra environment (e.g. French, 1976b, pp. 141-142; Washburn, 1980, p. 244). While this claim appears reasonable, there is a severe paucity of quantitative data either to confirm or deny its veracity (French, 1976b, p. 141). One objective of the present study was to supply the data needed to evaluate this hypothesis.

Second, the hydrological aim was to examine the slope hydrological cycle in a permafrost area. This objective was closely linked to the geomorphic investigation since both required evaluation of the spatial distribution and movement of water on slopes. In addition, the hydrological aim included the assessment of the applicability of models of stream runoff generation in nonpermafrost regions, to a permafrost area.

#### The importance of slopewash processes

Slopewash processes are considered important in humid temperate, hot arid and semiarid climatic zones. They are studied from four main scientific viewpoints.

Engineers are concerned with the ability of surface wash to cause accelerated erosion during land clearance, for construction (e.g. Diseker and Richardson, 1962; Weber and Reed, 1976; Younkin, 1973). The emphasis of their studies is on problem-solving rather than theory development and few attempts have been made to generalise from specific cases.

Slopewash processes are central to the work of agricultural engineers, and the problem of soil loss from agricultural land. In the U.S.A., the Universal Soil Loss Equation (USLE) has been developed

(Wischmeier, 1959; 1960; 1962; Wischmeier and Smith, 1958; 1965) and used to predict rates of soil loss. Unfortunately, the USLE is not truly universal since it cannot be used for steep slopes, in areas outside the conterminous U.S.A., or in areas where snowmelt-induced erosion is important (Wischmeier, 1976). At present, therefore, the USLE cannot be used to predict soil loss rates in permafrost regions.

Hydrologists are concerned with aspects of surface and subsurface wash, because in a typical catchment, these are the processes by which a major portion of the precipitation reaches the stream channel. In recent years, much research effort has been expended in attempts to generalise about the routing of water and the position of runoff source areas within individual drainage basins (e.g. Arnett, 1974; Betson, and Marius, 1969; Dunne and Black, 1970a; 1970b; 1971; Kirkby, 1969; 1978a; Ragan, 1968; Weyman, 1974). As a result of these studies, the simple Horton model (Horton, 1933; 1937; 1945) is no longer accepted as the usual mode of runoff production in humid temperate and other regions. It has been replaced by the variable source area concept (e.g. Dunne and Black, 1970a; 1970b; Freeze, 1974) and by the subsurface routing mechanism (e.g. Hewlett and Hibbert, 1967; Whipkey, 1965). The major problem evident in hydrological theory is that the very complexity of the runoff generative processes prevents the prediction of the specific disposition of source areas without intensive field investigation.

Geomorphologists have conducted numerous studies on natural and accelerated surface erosion resulting from wash processes, rainsplash, or both. These studies have been undertaken in humid temperate (e.g. Gerlach, 1967; 1976; Kirkby, 1967; Young, 1960), tropical (e.g. Lam, 1977;

Rougerie, 1956), and hot semiarid areas (e.g. Emmett, 1970; Leopold et al., 1966; Schumm, 1956; 1964). Few investigators have considered adequately the relationship between sediment movement and the areal distribution of surface runoff. As a result, existing models of sediment processes are based on hydrological assumptions that may be false (Kirkby, 1978b, p. 325). A notable exception to this statement is the work of A. Yair and colleagues in which both runoff and sediment generative processes have been examined (Yair, 1972; 1973; 1974; Yair and Klein, 1973; Yair and Lavee, 1974; 1976).

In addition to the field investigations of slopewash processes, experiments have been undertaken in the laboratory (e.g. Bryan, 1974b; 1976; 1979; De Ploey, 1977; Savař, 1975; 1977). The main advantages of the latter are (1) the ability of the investigator to control input variables and thus to examine separately the influence of each factor, and (2) through this control, to simulate events of low frequency and high magnitude without the use of long observation periods. The main disadvantage of laboratory research is that natural systems usually have to be transformed and simplified to make them manageable. This problem does not invalidate the laboratory results themselves, but it does create difficulties if the experimental results are to be applied to the natural environment.

In view of the recognition given to slopewash processes in other climatic zones, it is surprising that they have been studied so little in permafrost environments where they might be expected to be equally important. It cannot be claimed that this paucity of studies derives from a lack of awareness of the significance of slopewash processes. As early as the nineteenth century, it was noted that on Baffin Island and Greenland, larger than expected volumes of surface wash and streamflow were generated

during spring thaw. These were attributed to the impermeability of subsurface frost which also contributed to runoff as it melted (Tarr, 1897). Numerous other authors mention in passing the efficacy of slopewash processes in permafrost areas (e.g. Boneš, 1973; Czepe, 1965; Czudek and Demek, 1973; Dutkiewicz, 1967; Dylík, 1972; French, 1976b, pp. 141-145; Jenness, 1952; Journaux, 1976; Malurie, 1960; McCann et al., 1972; Pissart, 1967; Rudberg, 1969; Ryden, 1977; St-Onge, 1965; 1969; Washburn, 1947, p. 84; 1980, p. 144; Wedel et al., 1978) but none of them examines the processes in any detail.

In the permafrost environment, hydrological studies either consider limited aspects of the slope hydrological system (e.g. flow routing (Ballantyne, 1978) and snow distribution (e.g. Woo and Marsh, 1978)), or involve water balances of small drainage basins with considerable black-boxing of the internal processes (e.g. Marsh and Woo, 1979). Recent geomorphological work has ignored the slopewash system per se in permafrost areas and has concentrated on water movement on slopes within the context of nivation (e.g. Bird, 1967, pp. 226-229; St-Onge, 1969). Very few quantitative measurements have been made of slopewash sensu stricto, exceptions being Jahn (1961) on Spitsbergen and Wilkinson and Bunting (1975) on Devon Island. Only in the latter paper are both the hydrological and geomorphological aspects of slopewash considered, yet it is clear that these dual elements must be examined if the system is to be understood.

There are many similarities between slopewash processes in permafrost and nonpermafrost environments. At the level of mechanics, the process by which a particle is entrained in surface wash is the same in all climatic zones. At a larger scale, however, the origin of surface wash, its frequency of occurrence and magnitude, may be very different. These

differences and the difficulty of reproducing them in a laboratory are the justification for a field study in such a remote area.

#### Use of runoff plots

Data collection involved the measurement of several processes, in the belief that the effects of each on surface and subsurface water movement and sediment discharge could be identified. In theory, such an understanding allows predictions of the frequency and magnitude of slopewash processes in other similar areas from limited data sets. This approach was chosen in preference to that of studying slope form. Although wash-dominated slope segments have been discussed for permafrost areas (Jahn, 1972), direct field observation of the processes operating on such slopes has not been undertaken. In view of problems with equifinality, process measurement remains the obvious solution to determining the importance of slopewash in the permafrost zone.

The complexity of the slopewash system precluded intensive data collection over large areas. As a result, four sites, each having a runoff plot, were selected in the study area to the west of the Thomsen River (see Chapter 2, pp. 42-56). Three of the sites were examined during three successive years, while the fourth was studied for only the two later years.

Site selection involved three considerations. First, sites had to be reasonably accessible to the base camp to enable easy and regular monitoring. Second, the design of the runoff monitoring equipment meant that it could be used only on slopes with angles greater than  $3^\circ$ , and on those where there existed sufficient level ground ( $0.4 \text{ m}^2$ ) to seal the collector to the surface. The latter prevented measurements on areas possessing high concentrations of earth hummocks. Third, the sites needed

to be representative of the Thomsen River area. Since interfluves form a high proportion of the slopes in the catchment (approximately 30 - 40%), an interfluve location without a thick snow accumulation was chosen for one plot. The remaining three plots were located on slopes at points where large snowbanks develop during the winter. It was hypothesised that these were the locations where slopewash would be most effective.

The methodological level of the study is set at empirical generalisation and a combination of deductive and inductive reasoning is used. There are two reasons for this methodology. First, slopewash theory is inadequate to make accurate quantitative predictions of the efficacy of wash processes in different climatic zones. Second, comparisons with data currently available are limited by the paucity of measurements which have been made in the permafrost zone. As a result, conceptual hypotheses are both drawn from, and tested with, the field data.

CHAPTER 2

THE STUDY AREA

### Location

Fieldwork was undertaken in a tundra environment in north-central Banks Island (Figure 2) during the summers of 1977, 1978 and 1979. This island, with an area of approximately 70,000 km<sup>2</sup>, is the fourth largest in the Canadian Arctic Archipelago and is the most western of the major islands.

Sachs Harbour (population c.150), the only permanent settlement on Banks Island, is situated in the extreme south-west. Johnson Point, an oil company logistical base located on the east coast, was occupied between 1972 and 1976.

The selection of the field location was based on (1) the areal representativeness of a potential region, and (2) logistics. The Thomsen River study area is reasonably typical of the lowlands of central Banks Island (Figure 2) and comparable areas may exist elsewhere in the western Arctic, north of the tree-line. Results from this study may be applied to other locations if climate, terrain and vegetation are similar. Logistics for the study area were good in 1977, and research continued at the same location in 1978 and 1979 because the process studies required reasonably long-term measurements in a single locality.

The study area is located 230 km north-east of Sachs Harbour (base camp coordinates, latitude 73°14'N., longitude 119°32'W.). A Geological Survey of Canada (G.S.C.) tundra hut was moved to this location in 1975 and was used as a base for the duration of the investigations. Most of the research effort was concentrated at four runoff plots located 1.2 - 2.0 km

south of the base camp, but during the study, additional ground observations were made over an area of approximately 75 km<sup>2</sup>.

### Climate

The climate of Banks Island is polar continental, characterised by long cold winters, short cool summers and low but variable annual precipitation. Nine months of the year exhibit mean air temperatures below 0°C, and for nearly three months in winter and summer, continuous darkness and daylight respectively exist. As a consequence, the surface energy budget is dominated by radiative rather than turbulent terms (Vowinckel and Orvig, 1971) (Figure 3).

Climate data from Banks Island are limited both in duration and in the number of recording stations. Sachs Harbour possesses the sole lengthy record, dating from November 1955 (Table 1). Climatic records for four years (1972 - 1976) were collected at Johnson Point. G.S.C. parties recorded meteorological data from the interior of the island in the summers of 1974 and 1975, and the present study provides an additional three summers of information (1977 - 1979).

The summer climate of the Thomsen River study area is one of the major factors affecting sloopwash processes. The meteorological data collected during the present study, however, in themselves are too limited to be used in statements concerning long-term climatic conditions. Recourse must be made, therefore, to a comparison with the 25 years of records at Sachs Harbour. Two major differences exist between the two locations: (1) latitude, and (2) distance from the coast. Their effects must be

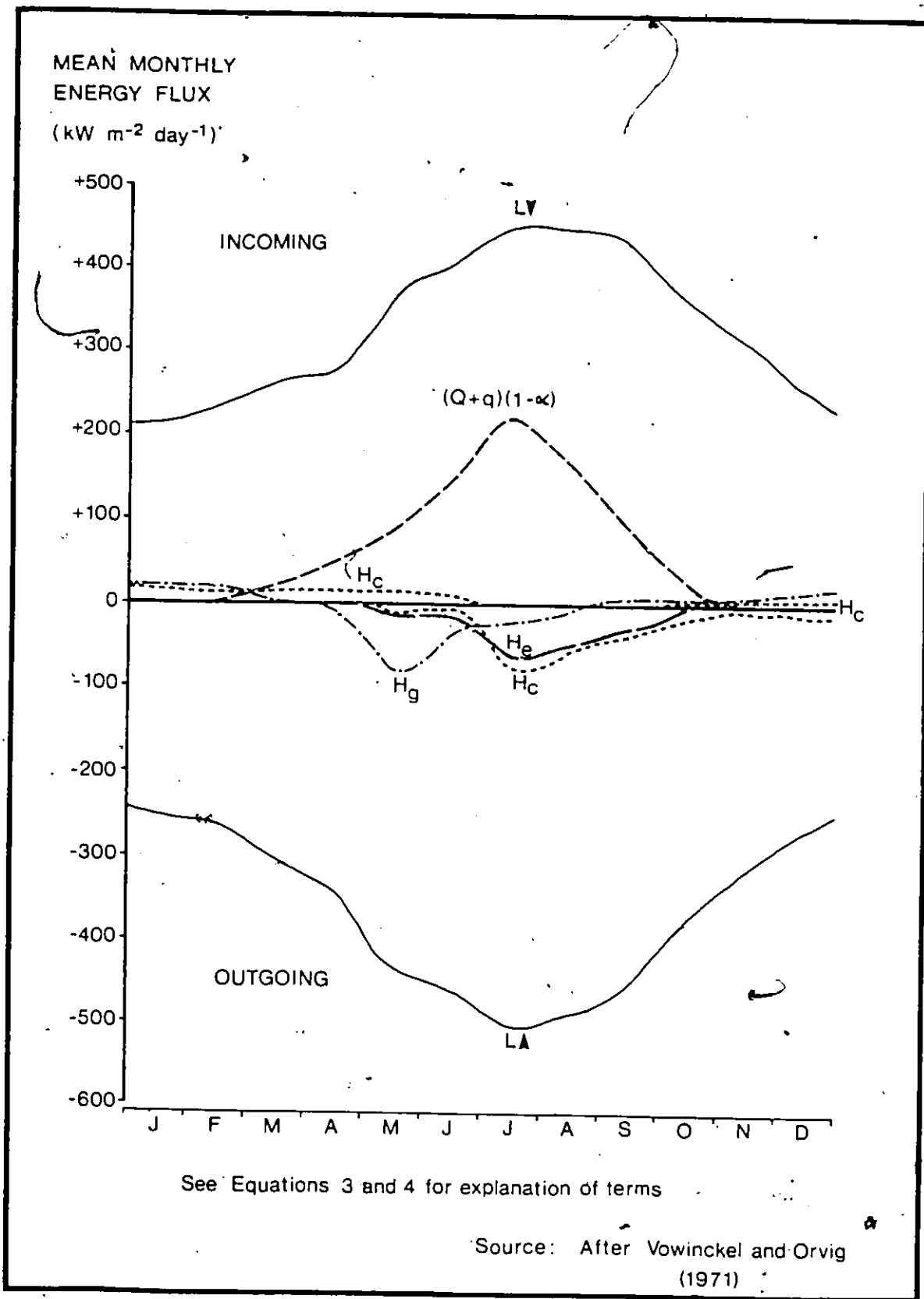


FIGURE 3 : SURFACE ENERGY BALANCE, SACHS HARBOUR, N.W.T., 1967

	J	F	M	A	M	J	J	A	S	O	N	D	Annual
<u>Air temperature (°C)</u>													
Mean (1955-1978) <sup>a</sup>	-30.7	31.0	-28.2	-19.8	-7.9	2.1	5.9	3.8	-1.8	-11.7	-22.4	-27.2	-14.1
Standard deviation <sup>b</sup>	2.9	3.2	2.9	3.0	1.9	2.4	1.8	1.9	2.0	3.0	2.3	3.4	
<u>Precipitation (mm)</u>													
Mean (1955-1978) <sup>a</sup>	3.0	4.1	3.3	4.3	8.1	7.6	15.7	22.4	16.3	17.0	6.9	4.3	113.5
Standard deviation <sup>a</sup>	2.8	4.3	3.0	4.6	8.1	6.6	10.9	12.4	9.9	11.9	3.8	3.0	34.0
<u>Date of thaw<sup>b</sup> (1956-1979)</u>													
Mean	June 9												
Standard Deviation	10.7 days												
Earliest	May 24, 1976												
Latest	June 29, 1979												
Probability Earlier	0.068												
Probability Later	0.023												
<u>Date of freezeback<sup>b</sup> (1956-1977)</u>													
Mean	September 7												
Standard Deviation	10.0 days												
Earliest	August 22, 1956												
Latest	September 27, 1976												
Probability Earlier	0.055												
Probability Later	0.230												
<u>Duration of thaw<sup>b</sup> (1956-1977)</u>													
Mean	91 days												
Standard Deviation	16.9 days												

<sup>a</sup> Canada Climate Centre personal communication

<sup>b</sup> Calculated from Monthly Record

TABLE 1: Selected climatological data, Sachs Harbour, N.W.T.

considered before the normality or otherwise of the study years can be assessed.

The effects of a change in latitude can be shown through comparison of data from Sachs Harbour and Johnson Point (Table 2). The annual temperature averages  $1.6^{\circ}\text{C}$  lower at the latter location, but differences are greatest in the months of March, April, May and June, retarding thaw dates by up to 10 days. Thaw dates are crucial at high latitudes because albedo changes are great during snowmelt, and up to 40% of annual incoming solar radiation is available during the months of June and July. If snowmelt is retarded, therefore, much of this potential energy is reflected from the snow surface (Vowinckel and Orvig, 1971). Dates of freezeback which influence the cessation of slopewash activity are likely to occur earlier at Johnson Point than at Sachs Harbour, resulting in a shorter thaw season (Table 2).

Both latitude and distance from the coast are increased in a comparison of the 11 summer months of field data and the Sachs Harbour records (Table 3). Interior locations average  $1.7^{\circ}\text{C}$  warmer than Sachs Harbour with extreme monthly differences as high as  $4.2^{\circ}\text{C}$ . During June and July, temperatures at Johnson Point average  $1.3^{\circ}\text{C}$  lower than those at Sachs Harbour suggesting that, with no large variation in relief, summer temperature differences between the northern interior of Banks Island and the coast are usually between  $2.5^{\circ}\text{C}$  and  $3.5^{\circ}\text{C}$ . The temperature differences between the northern interior and Sachs Harbour probably are  $1.0 - 2.0^{\circ}\text{C}$ , and dates of thaw may be earlier in the more northerly area. This information is important because it suggests, for example, that the temperatures recorded at Thomsen River in July 1977 (Table 3) were not as

	J	F	M	A	M	J	J	A	S	O	N	D	Annual
<u>Johnson Point air temperature(°C)</u>													
Mean (1972-1976) <sup>a</sup>	-33.5	-33.0	-32.2	-21.9	-9.5	0.9	5.0	2.4	-3.5	-14.3	-23.5	-29.0	-16.0.
<u>Air temperature difference(°C)</u>													
Johnson Point-Sachs Harbour <sup>a</sup>	-0.4	-1.7	-2.3	-3.1	-3.7	-2.0	-0.6	-1.3	-0.3	-1.7	-0.5	-1.3	-1.6

Year	Date of thaw		Date of freezeback		Duration of thaw (days)	
	Johnson Point	Sachs Harbour	Johnson Point	Sachs Harbour	Johnson Point	Sachs Harbour
1972	June 18	June 8	September 3	September 5	77	89
1973	May 27	May 26	September 9	September 23	105	120
1974	June 20	June 14	September 3	September 2	75	80
1975	May 31	May 31	Missing	August 30	Missing	91

<sup>a</sup> Data missing: January-March 1972; August 1975, May-December 1976  
 Source: Derived from Monthly Record

TABLE 2: Comparison of climatological data, Johnson Point and Sachs Harbour, N.W.T.

Air temperature (°C)

Year	Location	Dates of record	May	June	July	August
1974	Sachs Harbour <sup>a</sup>	1/5 - 31/8	-9.2	-0.8	5.8	2.4
	Johnson Point <sup>a</sup>	1/5 - 31/8	-9.9	-1.9	4.6	2.4
	G.S.C. Day <sup>b</sup>	24/6 - 17/8			8.5	
1975	Sachs Harbour <sup>a</sup>	1/5 - 31/8	-4.3	4.8	4.1	3.5
	Johnson Point <sup>a</sup>	1/5 - 27/7	-7.8	2.8	5.1	
	G.S.C. Day <sup>b</sup>	2/6 - 18/8		3.2	5.0	
	Thomsen River <sup>b</sup>	5/6 - 13/8		6.3	7.3	
1977	Sachs Harbour <sup>a</sup>	1/5 - 31/8	-5.7	5.3	8.1	6.0
	Thomsen River <sup>c</sup>	27/5 - 26/7		5.8	12.3	
1978	Sachs Harbour <sup>a</sup>	1/5 - 31/8	-11.9	-2.1	5.1	-0.2
	Thomsen River <sup>c</sup>	24/5 - 16/8		-0.9	9.1	
1979	Sachs Harbour <sup>a</sup>	1/5 - 31/8	-8.5	-1.8	7.0	4.1
	Thomsen River <sup>c</sup>	22/5 - 25/7		-2.2	7.4	
<u>Dates of thaw</u>		1977	1978	1979		
Sachs Harbour		May 30	June 25	June 29		
Thomsen River		May 30	June 21	June 26		
<u>Positions</u>						
Sachs Harbour		71°57'N., 124°44'W.				
Johnson Point		72°47'N., 118°30'W.				
G.S.C. Day		73°17'N., 120°40'W.				
Thomsen River		73°14'N., 119°32'W.				

N.B. Data not shown if more than 7 days missing from one month  
 Source: a Derived from Monthly Record b From Miller (1975) c Field data

TABLE 3: Summer climatological data, Banks Island, N.W.T.

extreme as they appear if the whole island is assumed to be represented by the Sachs Harbour mean July temperature.

Extreme events of temperature and precipitation may be significant to slopewash processes. The former may promote a particularly rapid snowmelt, while the latter (either heavy rainfall or a particularly large winter snow accumulation) may cause thresholds in surface wash production to be exceeded, potentially resulting in surface erosion. Variability is an important element in the temperature and precipitation regimes at Sachs Harbour (Table 1). In the 25 years of temperature records, differences from monthly means have exceeded two standard deviations on ten occasions. The maximum numerical monthly difference from the mean in winter was  $-8.6^{\circ}\text{C}$  (January 1975), while in the summer period, the maximum was  $+5.0^{\circ}\text{C}$  (August 1958). These figures imply that monthly temperatures differing from the mean by  $\pm 4 - 7^{\circ}\text{C}$  are not uncommon. Differences of this magnitude during the May-June and August-September periods account for the large standard deviations in dates of thaw and freezeback (Table 1). It is thought that variability of this order also occurs at Thomsen River, resulting in the large differences in monthly air temperatures recorded in the area between 1976 and 1979 (Table 3).

The probability of very heavy storms at Sachs Harbour appears to be less than at other locations in the Canadian Arctic Archipelago. At this station, the maximum recorded 24-hour storm was only 21.8 mm (24 years of records) and standard analysis indicates that on the basis of this admittedly short record, even the 100-year storm is only 25.8 mm (Figure 4). The value should be treated with caution as convective storms in the centre of the island may produce greater amounts of rainfall, but it remains far

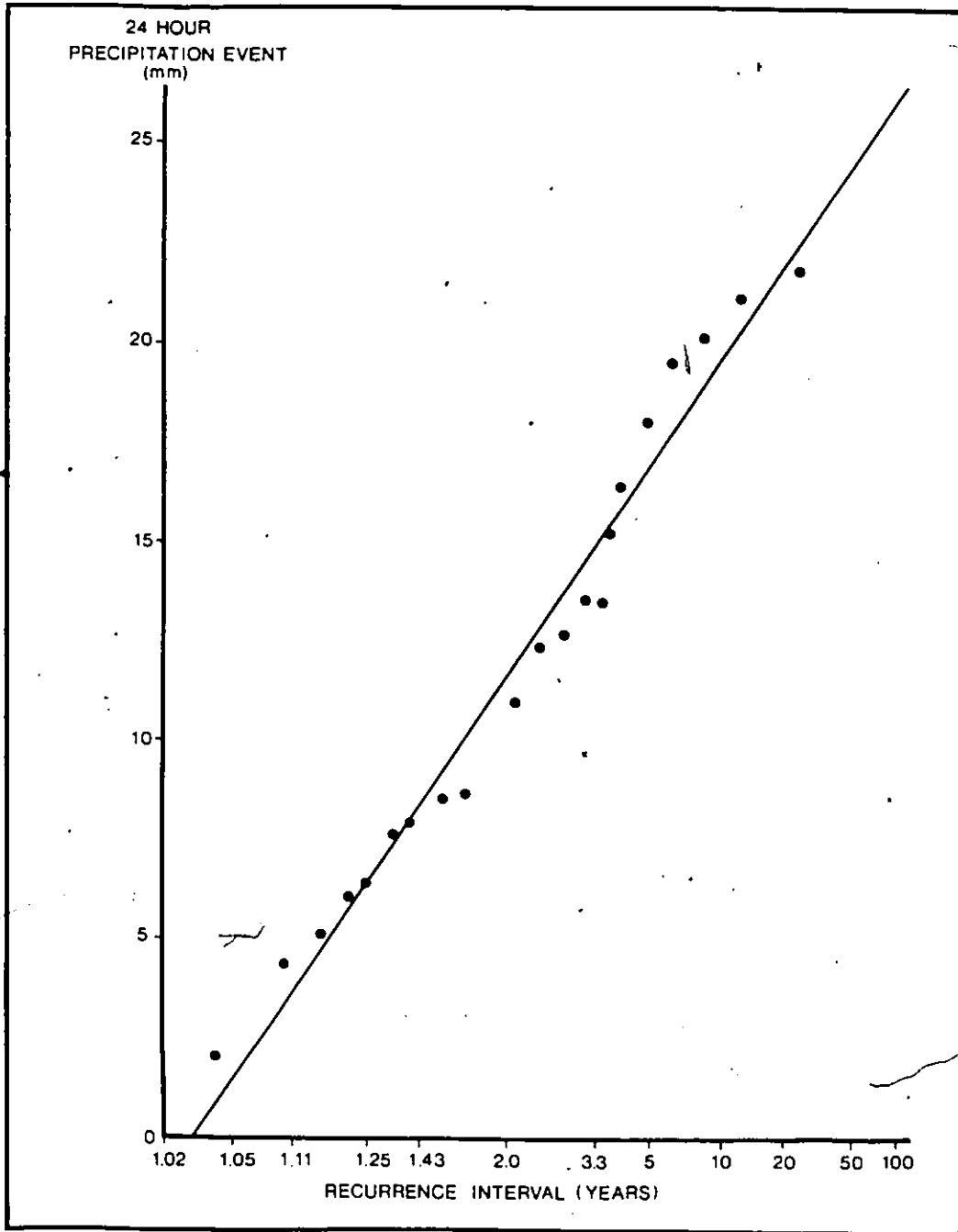


FIGURE 4 : RECURRENCE INTERVAL OF MAXIMUM ANNUAL 24-HOUR  
PRECIPITATION EVENT, SACHS HARBOUR, N.W.T.

below the 54.6 mm storm recorded on Ellesmere Island (Cogley and McCann, 1976) which probably has a recurrence interval of 25 - 50 years.

Winds are significant to slopewash processes in a high Arctic area in a number of ways. Their primary influence is shown in the redistribution of winter snowfall which results in a very thin covers on ridges and interfluves and deep accumulations on lee slopes. The prevailing winter wind direction at Sachs Harbour is south-east for all but the months of October and November when it is east (Monthly Record data, 1955 - 1970). It is possible, however, that a different direction dominates at Thomsen River. Field observations of snow accumulations suggest that the prevailing winter wind direction is north-west, causing the greatest snowbanks to develop on slopes facing south-east.

During the melt period, winds are significant in the energy balance, but little predictive capacity can be gained from examination of the Sachs Harbour data. A general comparison, using wind polygons for the summer months, reveals obvious differences (Figure 5). These may be the result of local conditions (e.g. the north-west - south-east trend of the Thomsen River data may be the result of funnelling by river valley) or may be caused by synoptic variation. It is interesting to note that there were considerable differences even between the two years at Thomsen River, with a more frequent north-east component in 1978.

Wind also plays an important role as the initiator of aeolian sediment movement. During the summer months of 1977, wind-borne sediment was observed on a number of occasions. Only local sediment movement occurred during lesser winds in 1978 and 1979. The strongest winds experienced (gusts up to  $18 \text{ m s}^{-1}$ ) occurred on May 24, 1979 when sediment source areas were still covered by snow, limiting aeolian sediment transportation.

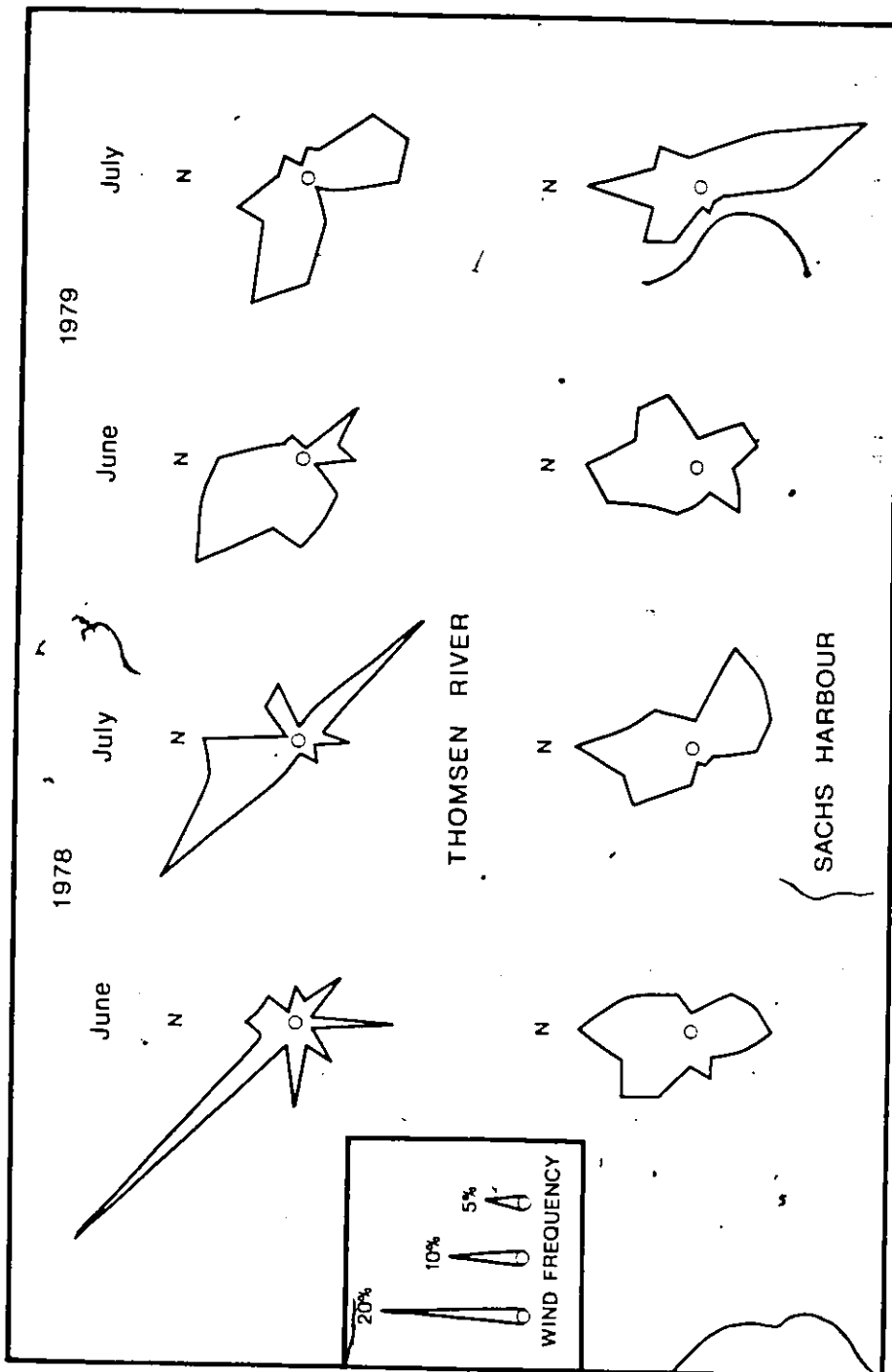


FIGURE 5 : WIND POLYGONS, THOMSEN RIVER AND SACHS HARBOUR, JUNE AND JULY 1978 AND 1979

Thomsen River data from 12 hour observations and 16 directions;

Sachs Harbour data from 6 hour observations, adapted from 36 directions.

### Permafrost

Banks Island lies within the continuous permafrost zone (Figure 1). In the study area, surface evidence of the existence of permafrost is provided by a number of distinctive ground ice forms. Beaded streams, low centred ice-wedge polygons, thaw lakes and thermokarst gullies have all been observed. Pingos and thermo-erosional thaw flow slides have been studied on Banks Island (e.g. French, 1974; 1975; 1976a; French and Egginton, 1973; Pissart and French, 1976; 1977) but apparently do not exist in the study area.

The temperature of permafrost at the level of zero annual amplitude can be estimated from the mean annual air temperature. The difference between the two is approximately  $3.3^{\circ}\text{C}$  in the Arctic (Brown, 1972), implying that the temperature of the permafrost at Sachs Harbour is  $-10.8^{\circ}\text{C}$  (mean annual air temperature is  $-14.1^{\circ}\text{C}$ ). Since mean annual air temperatures vary over the island, permafrost temperatures probably range between  $-10.0^{\circ}\text{C}$  and  $-12.0^{\circ}\text{C}$ . The only deep ground temperature data for Banks Island are from a borehole at Storkerson Bay, 150 km west of Thomsen River. The permafrost thickness there is estimated to be 500 m (Taylor and Judge, 1974), but it is possible that permafrost is somewhat thicker inland. If ground temperatures are between  $-10.0^{\circ}\text{C}$  and  $-12.0^{\circ}\text{C}$ , the geothermal gradients lie in the range of  $2.0 \times 10^{-2} - 2.4 \times 10^{-2} \text{ }^{\circ}\text{C m}^{-1}$ .

Above the permafrost, active layer thicknesses generally decrease with increasing latitude (Brown, 1972). Local factors, however, such as aspect, substrate thermal diffusivity, and vegetation cover are predominant, and variation within a single area can be considerable. In the study area,

for example, a thawed layer thickness in excess of 0.8 m was measured on a poorly vegetated sandy interfluvium in mid-July 1977, while the total active layer thickness of a vegetated silty-clay terrace in 1978 was only 0.5 m.

Permafrost is an important influence on slope hydrology in three ways. First, its presence usually prevents the deep percolation of water. This affects the flow routing and speed of response of a slope subjected to water inputs from snowmelt or precipitation. Second, the lowering of the frost-table that occurs during the summer months, results in a daily change in the hydrological parameters of a slope, as greater volumes of water are permitted to move as subsurface flow. Third, disturbance of the ground thermal regime may cause permafrost degradation and the melting of ground ice. If ground ice thaw is selective, this may lead to the channeling of surface wash and the initiation of rills and gullies.

### Physiography

#### (a) Terrain

In terms of its effects on slopewash processes, terrain is a significant factor at a number of scales. At the macro-scale, the general form of the landscape - highland, upland and lowland - and its dissection, affect relief differences and slope angles. At the meso-scale, slope profiles, and particularly breaks of slope, are dominant factors in the development of large snowbanks. At the micro-scale, the formation of hummocks or solifluction stripes can result in the concentration of surface flow.

The study area lies within the lowlands of central and western Banks Island (Fyles, 1962) (Figure 2) and probably is reasonably typical of this physiographic region. The maximum elevation is 150 m, compared with the maximum for the whole of Banks Island of 730 m. Typical height differences between ridge crests and adjoining valley floors are 20 - 50 m, and the lowest point in the study area is on the floor of the Thomsen River valley at an elevation of 30 m (Figure 6).

The study area (Figure 7) consists of an upland (Figures 8 and 9) dissected by a wide (2.0 - 2.5 km) valley that trends south-east to north-west. The Thomsen River flows north in the southern part of the area in a relatively narrow (0.8 km wide) valley with poorly developed alluvial terraces (Figure 10). It enters the main valley at its confluence with the "Sarfarsuk", 1.4 km south of the base camp. When joined by a second major tributary, Dissection Creek, 4.0 km downstream of the base camp, its flow direction is north-west.

The main valley (Figure 11) shows the development of two terraces above the present flood plain (Figure 12). On the main (second) terrace, which may date from Wisconsinan times, remnants of former islands are evident (Figures 7 and 12), as are patterned ground phenomena such as low centred thermal contraction crack polygons and thermokarst lakes (Figure 11). A number of gullies dissect the main river terrace on the west bank of the Thomsen River just south of the base camp, and their form suggests initiation along ice wedges. Ice wedge exposures have not been observed in the study area, but beaded streams and polygonal ground in the upland indicate their presence in colluvial materials (see Figure 8). Some of the thaw lakes on the second terrace show forms indicating expansion and amalgamation (Figures 7 and 11), but they are not of the oriented type.

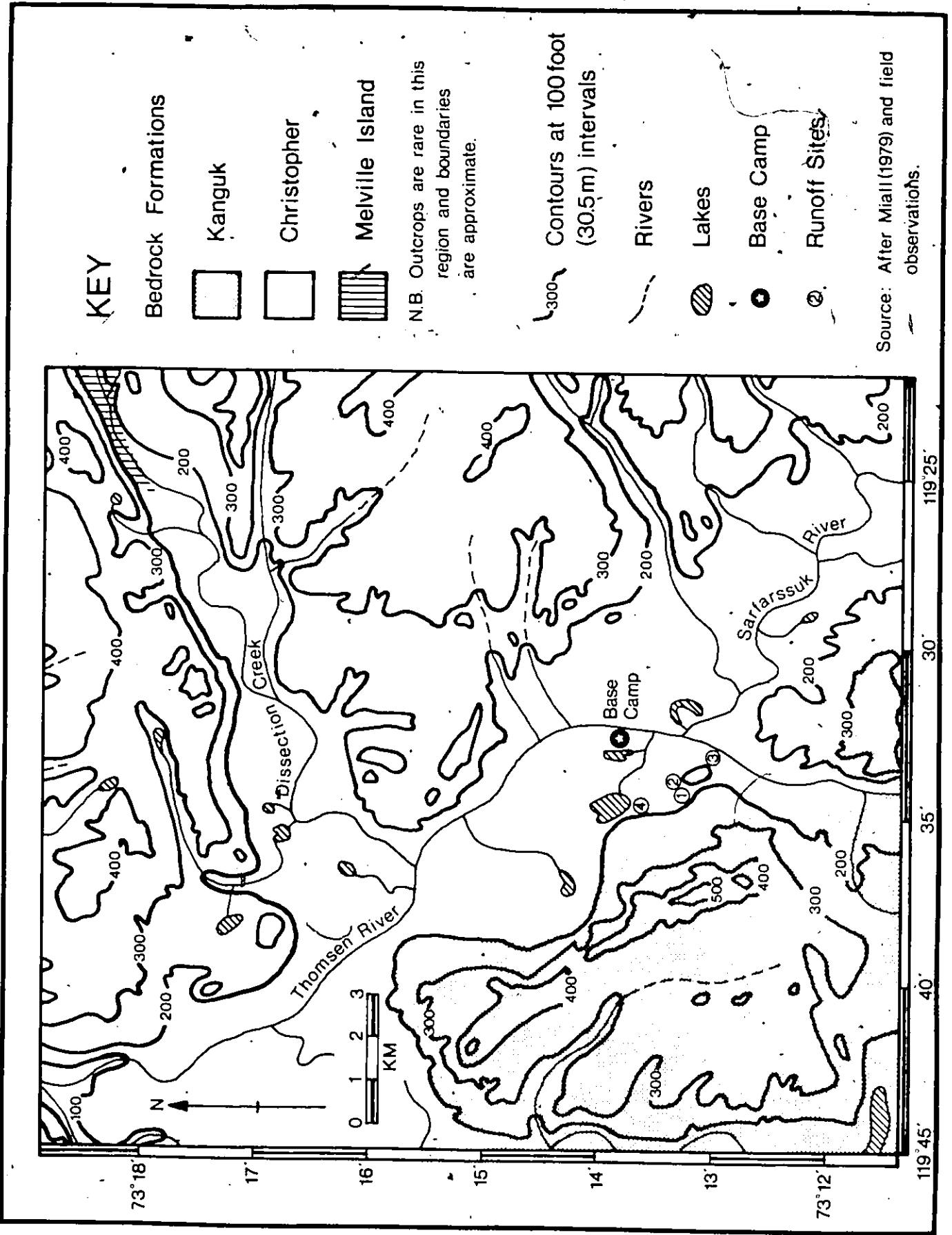


FIGURE 6 : RELIEF AND BEDROCK GEOLOGY OF THE STUDY AREA



FIGURE 7 : AIR PHOTOGRAPH OF THE STUDY AREA  
Enlargement of part of A - 17379 - 40.  
Position of base camp marked by circle.



**FIGURE 8 : EDGE OF THE UPLAND AREA, JULY 3, 1977**  
 Aerial view south-west with plot 2 centre left, marked by arrow. Note ice-contact deposits on slope crests, nonsorted stripes, ice-wedge polygons in colluvial materials, and beaded stream.



**FIGURE 9 : UPLAND AREA, JUNE 5, 1977**  
 View east towards base camp and Thomsen River valley. Note unvegetated ridges trending south-east to north-west and the presence of large snowbanks while interfluvies are snow-free.



FIGURE 10 : THOMSEN RIVER, SOUTH OF CONFLUENCE WITH THE  
'SARFARSSUK', 10 DAYS AFTER PEAK DISCHARGE,  
JULY 10, 1978

View from crest of site 3 looking south-east. Note extensive  
surface wash downslope of snowbank and aeolian deposition  
on snow surface.

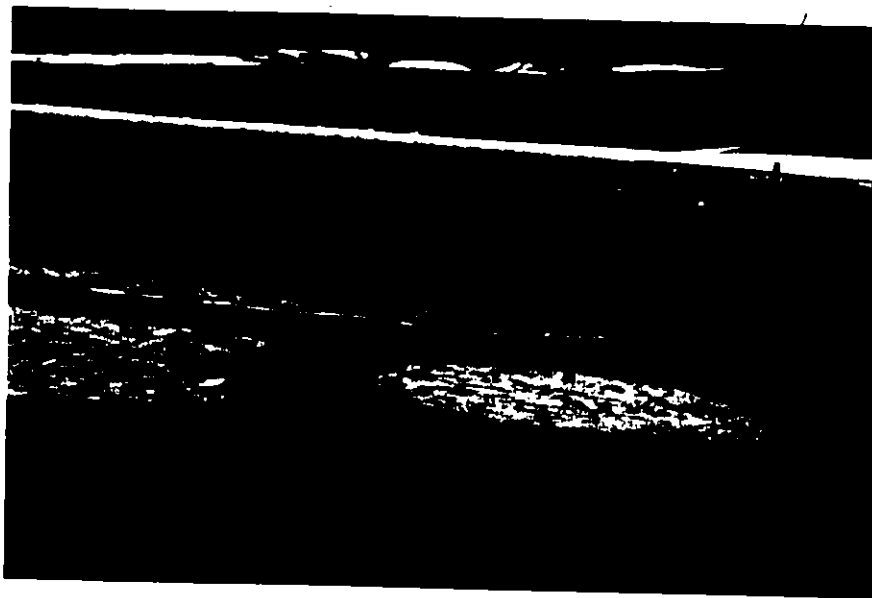


FIGURE 11 : THOMSEN RIVER VALLEY, MAIN TERRACE,  
JULY 10, 1979

Aerial-view east to base camp showing tundra ponds  
and lakes, and low centred polygons. River flow is  
from right to left (south to north).

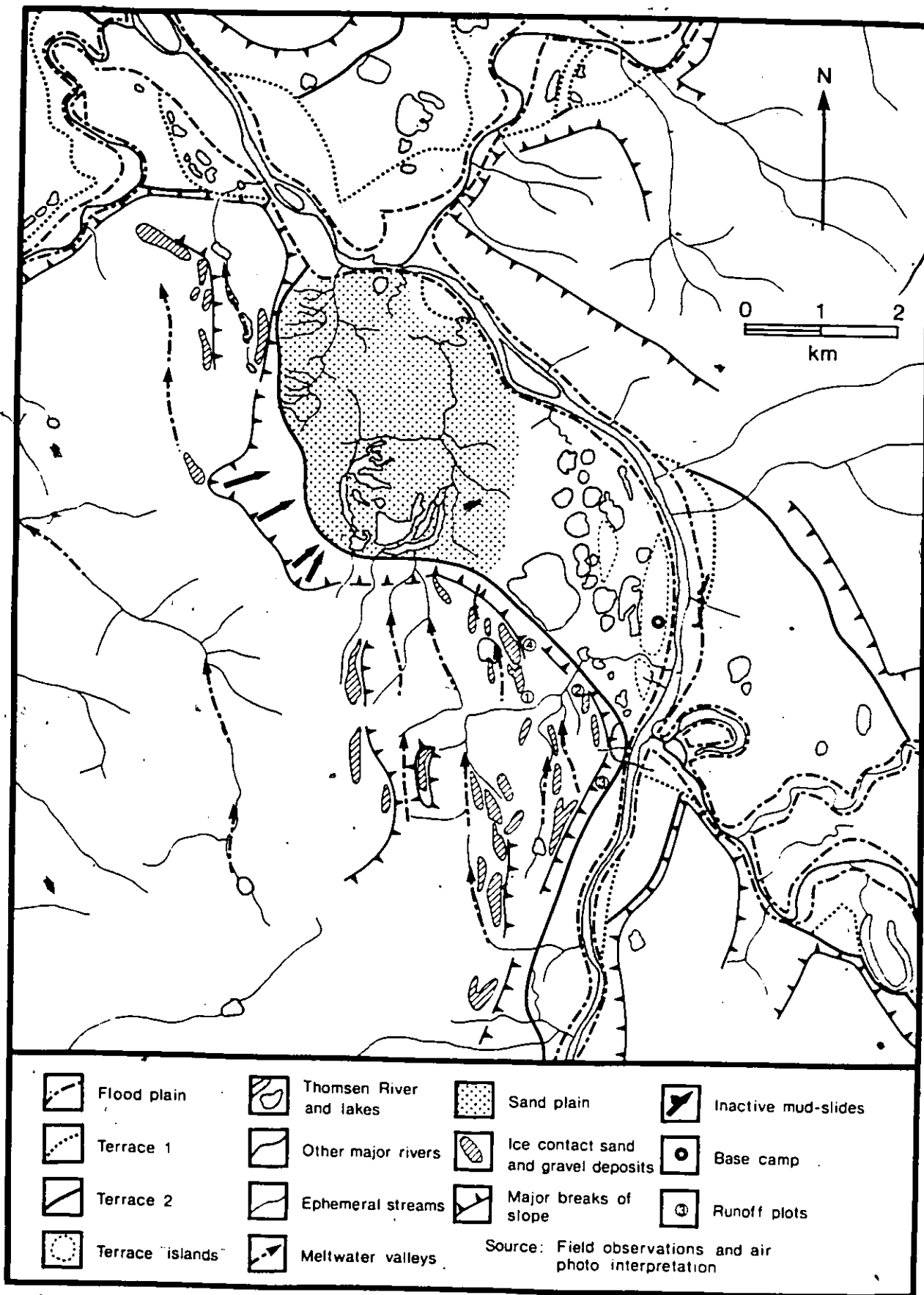
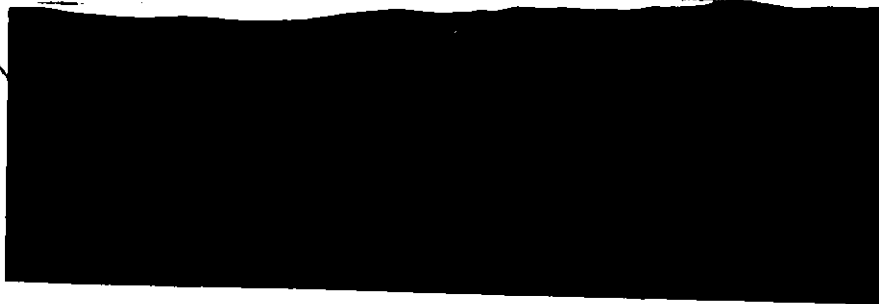


FIGURE 12 : MORPHOGENETIC MAP OF THE STUDY AREA

Slope profiles in the study area typically are convex at the crest, grading into low-angled rectilinear or concave segments. There are no free-face sections and slope angles seldom exceed 15 - 20°. Large breaks of slope are not uncommon, but the areas affected by long-lasting snowbanks constitute only a small percentage of the study region. Because slopes are relatively short, there are almost no slopes where one large snowbank is upslope of another.

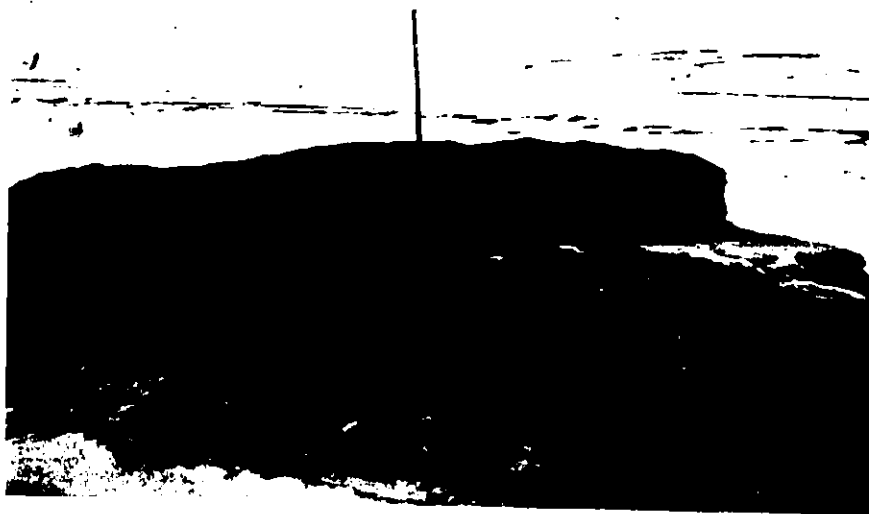
Other terrain elements that are significant to slopewash processes include unvegetated areas. These occur on interfluvies where moisture inputs are insufficient to support a closed vegetation cover (see Figure 9), and in the sand plain on the second terrace where active dunes and blowouts exist (Figures 13 and 14). Both of these sets of bare areas may contribute aeolian sediment to snowbanks, affecting both ablation rates and sediment outputs.

Micro-relief forms in the study area include earth hummocks and solifluction stripes. The stripes are nonsorted (see Figure 8): cross-sections show that if channeling of subsurface wash takes place along the unvegetated lines, it is insufficient to cause the removal of fines. This is in contrast to observations made on Cameron Island, N.W.T. where wash sorting is apparent (personal communication, H.M. French, 1977). Hummocks are consistent features of snowbank locations in the Thomsen River area. They range in size up to 0.7 m in diameter and to 0.3 m in height, and usually are linked to distinct vegetation assemblages dominated by Cassiope tetragona or Dryas integrifolia. The exact mechanism of hummock formation is not fully understood (for contrasting views see Mackay (1979) and Tarnocai and Zoltai (1978)), but hummocks affect surface wash by appearing through the snowpack first, thereby diverting and channeling subsequent meltwater flow.



**FIGURE 13 : ACTIVE SURFACE IN THE SAND PLAIN,  
JULY 13, 1978**

Note long 'tails' of sand developed in the lee of small stabilised hummocks, resulting from strong northerly winds on July 11, 1978.



**FIGURE 14 : BLOWOUT IN THE SAND PLAIN, 1.5 KM NORTH-WEST  
OF THE BASE CAMP, MAY 25, 1978**

Snow cover of surrounding area was close to 100%, but early exposure of sand allowed local aeolian erosion and transport to occur. Note candling of snow where deposition has taken place. Rod is 0.75 m long.

(b) Drainage.

There are obvious differences between the drainage patterns and regimes of the upland area and the valleys of the Thomsen River and its tributaries.

In the upland, streams are ephemeral and follow typical arctic nival regimes (Church, 1974). Many of them are underfit and flow in old glacial meltwater channels, resulting in a poorly integrated drainage system. Some of the streams are beaded and do not follow well-defined beds during the spring flood. Lakes are uncommon features in the upland valleys.

The Thomsen River drainage network includes three rivers that flow throughout the summer, and fall into the category of large river nival regimes (Church, 1974). Within the river valleys, many shallow lakes are present, although some of the smaller tundra ponds on the second terrace contain water only during the spring runoff season. The streams that dissect the terrace are well-defined gullies in their lower reaches and probably were initiated along ice wedges. In the sand plain, a large lake (possibly an ox-bow) once existed, but only remnants remain (Figure 7). A series of ox-bow lakes are present on the flood plain of the "Sarfarssuk" River (Figure 7).

Most surface wash in the study area reaches stream channels without becoming concentrated into rills or gullies, primarily because of the relatively gentle slopes and the role of moisture in promoting protective vegetative growth. Drainage densities (calculated from air photographs at a scale of 1:13,000) are low in the upland, between

2.5 - 3.0 km km<sup>-2</sup>, and the average slope length from crest to channel is 175 - 250 m. The average surface flow travel distance probably is somewhat less than this, since much of the runoff is produced by large snowbanks which usually do not develop at the watersheds.

The Thomsen River exhibits a low suspended load and through a large part of the study area flows over a bed armoured with cobbles and boulders. The "Sarfarsuk" River and Dissection Creek are actively eroding their channel banks and their sediment concentrations probably are higher (Anderson and Durrant, 1976). The small upland streams possess high turbidities during the melt season but much of their load appears to be vegetative debris.

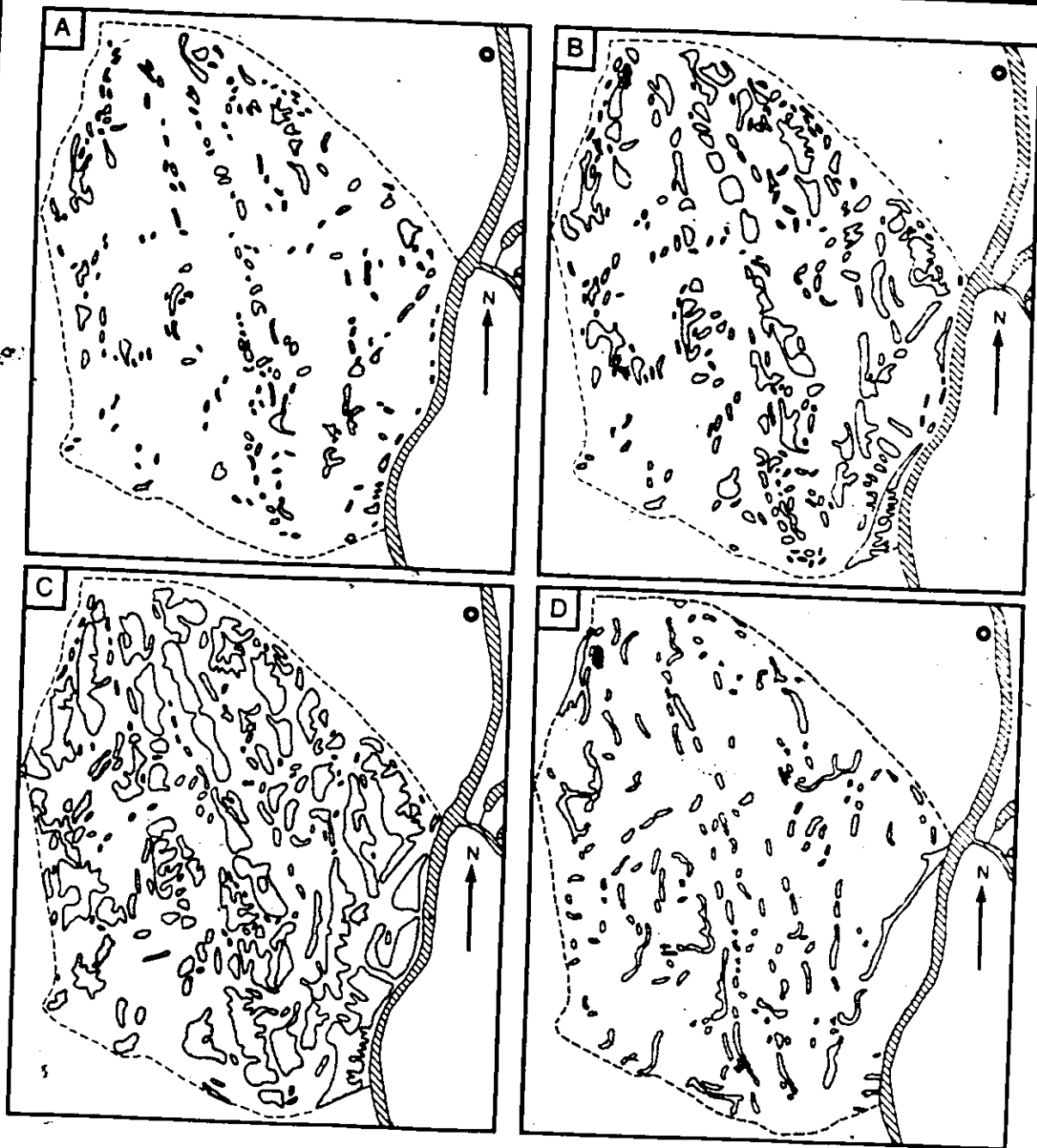
The patterns of snowmelt and streamflow in the upland area are closely linked. Snowcover is continuous at the end of winter, but evaporation while air temperatures remain below 0°C soon results in the development of bare areas on the ridges. As air temperatures rise to 0°C and above, more areas of the upland become snow-free, and water begins to saturate the snow in the valley bottoms. Within one or two days, the saturated patches link up and flow begins within the small streams. In 1979, for example, flow began in the headwaters of the valley between runoff plots 1 and 2 (Figure 7) during the evening of June 30. The stream reached the exit of the valley, flowed onto the terrace by mid-day on July 1, and into the Thomsen River via a deeply incised gully at 2100 h, July 1. At this time, the Thomsen River had been flowing over bedfast ice for two days. Estimated peak discharge of the small streams occurred 4 - 5 days after the initial flow (i.e. on July 3 - 4, 1979) and was followed by a relatively rapid decline.

Changes in stream discharge can be related to the pattern of snowmelt (Figure 15). The ridges became snow-free first, and by June 18, 1979, 15% of the ground within a 7 km<sup>2</sup> area of the upland was exposed (Figure 15A and B). Between June 18 and June 27 (Figure 15C), a further 17% of the area became clear of snow and water began to collect in the valley bottoms. The greatest change, however, took place during the period June 27 - July 6 when all the valley bottoms and most of the valley-side slopes lost their snow covers, leaving only major snowbanks (7% of the area) to supply streamflow (Figure 15D). This period of rapid melt corresponded to the period of maximum streamflow in the small streams, although the Thomsen River crested some days later.

#### Bedrock geology

The geologic succession of Banks Island includes rocks of many ages, but more than 80% of the land surface is underlain by Mesozoic and Tertiary strata (Figure 16) (Miall, 1979). With limited exceptions (such as the Devonian beds in the north-east and the Glenelg Formation in the south of the island), most of the bedrock is poorly lithified sands, clays and shales. In addition, much of Banks Island is covered by surficial deposits.

There are two main rock types in the Thomsen River study area and both are of Cretaceous age (Miall, 1979). The Lower Cretaceous Christopher Formation is a soft, dark grey shale. The Upper Cretaceous Kanguk Formation lies unconformably on the Christopher, and in the area consists of a lower bituminous member overlain by a silty shale member. The dip of the rocks is approximately west and at a low angle.



0 0.5 1.0 1.5 2.0  
KM



FIGURE 15 : THE PATTERN OF SNOW ABLATION IN THE THOMSEN RIVER UPLAND AREA, 1979

- A - 7% snow-free, June 12;  
 B - 15% snow-free, June 18;  
 C - 32% snow-free, June 27;  
 D - 93% snow-free, July 6.

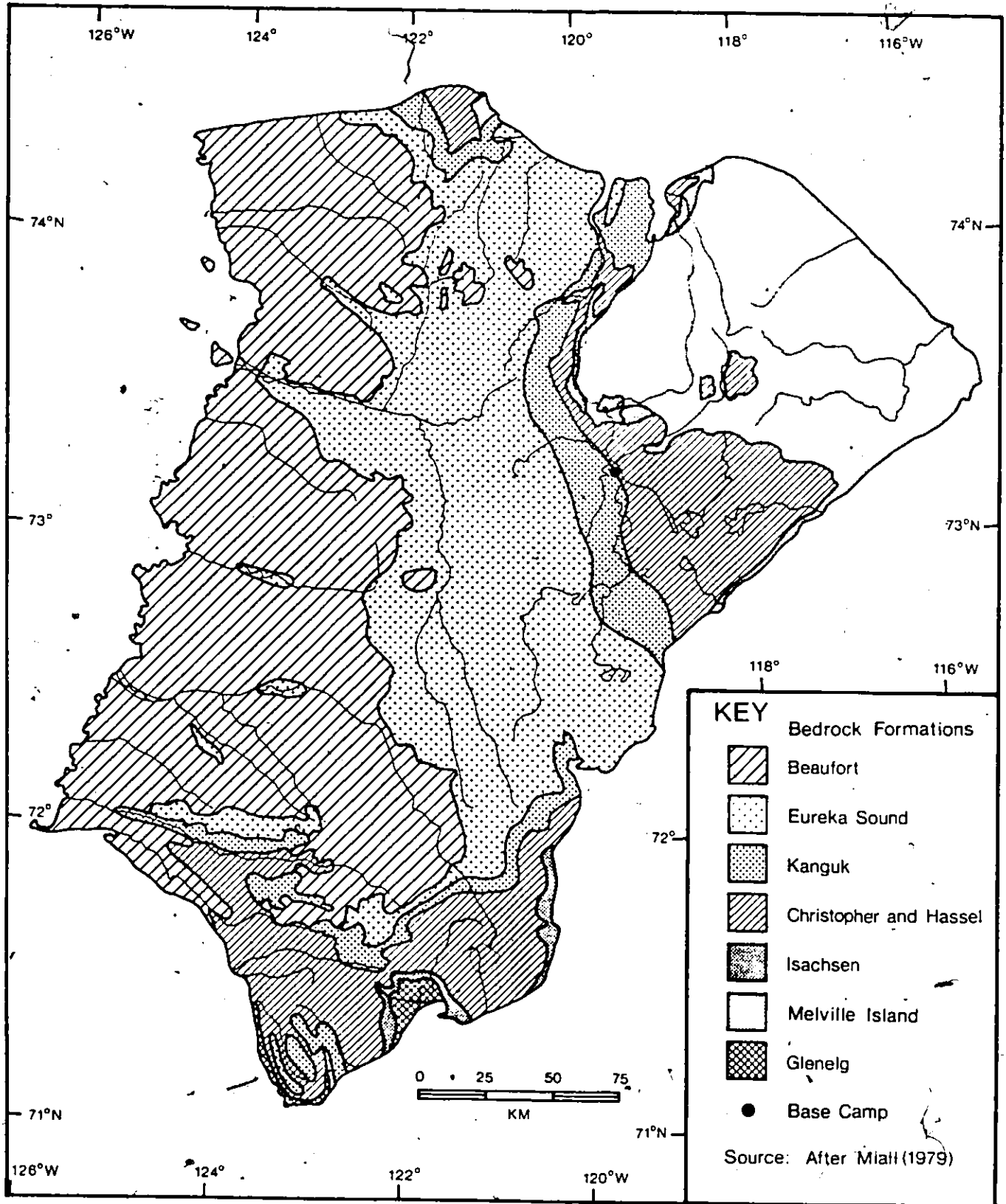


FIGURE 16 : BEDROCK GEOLOGY OF BANKS ISLAND

The position of the boundary between the Kanguk and Christopher Formations is a matter for dispute. Both the old standard work (Thorsteinsson and Tozer, 1962) and the new (Miall, 1979) place the boundary of the Christopher on the edge of the upland to the west of the camp. It is shown as an approximate edge, however, because the extensive cover of surficial materials has made accurate boundary positioning difficult without drilling. Field observations suggest that the contact is actually 0.5 - 1.0 km further west in the study area (Figure 6). Surface exposures of Christopher shale have been observed within the upland, and old mud slides on the western edge of the sand plain provide supplementary evidence (Figure 12). The only definite location of an outcrop of the Kanguk Formation is on the east bank of the Thomsen River 5 km south of the base camp, where the bituminous shale member has oxidised to its distinctive red colour.

The bedrock geology is relatively unimportant to slopewash processes in the study area because of the near-complete cover of surficial materials. Slopewash takes place on the ground surface and within the active layer, so that if surficial materials exceed one metre in thickness, the bedrock does not directly affect the processes. Bedrock is more important when considering the relative significance of slopewash and other processes. Christopher shale, for example, is susceptible to rapid mass movements so that the relative importance of slopewash in an area underlain by this rock may be less than in one of more resistant deposits.

#### Quaternary geology

The Thomsen River region is shown as an area glaciated during the Wisconsin period on the first Glacial Map of Canada (Geological

Association of Canada, 1958). Prest et al. (1968) and Vincent (1978b), on the other hand, show the furthest advance of the Wisconsin ice sheet ending east of the Thomsen River. All agree, however, that the study area was glaciated during pre-Wisconsinan advances (Figure 17).

There is little evidence of the earliest recognized glaciation (the Banks Glaciation (Vincent, 1978a)) in the study area, but the Thomsen Glaciation (which pre-dates the Wisconsin) included a lobe that flowed north down the Thomsen valley. It deposited the ice-contact sandy gravels which are to be found on the ridges in the upland area south-west of the base camp (Figures 8, 12 and 18). During this period, the numerous small meltwater channels (Figure 9) that trend south to north in this locality, probably were eroded, although most of the drainage must have taken place in the large meltwater valley further to the west (Figure 18). The ice-contact deposits contain high percentages of gravel-sized particles which are not easily removed by surface wash and form a lag deposit. On ridges where the gravel cover is incomplete, surface erosion is much more apparent.

The study area may have been submerged during a marine transgression (the Big Sea (Vincent, 1978a)) following the Thomsen Glaciation, but no evidence exists for this event within the study area itself. The furthest advance of the Viscount Melville Glaciation (=Amundsen Glaciation ?) is marked by morainal deposits characterised by high-centred polygons 15 - 20 km east of the base camp. Meltwater from this glacial episode probably increased the width of the "Sarfarssuk" - Thomsen River valley to its current extent and infilled it with fluvio-glacial materials. The sand plain on the second terrace may have been laid down as an outwash delta into Lake Ivitaruk. This lake resulted from the blocking of northward

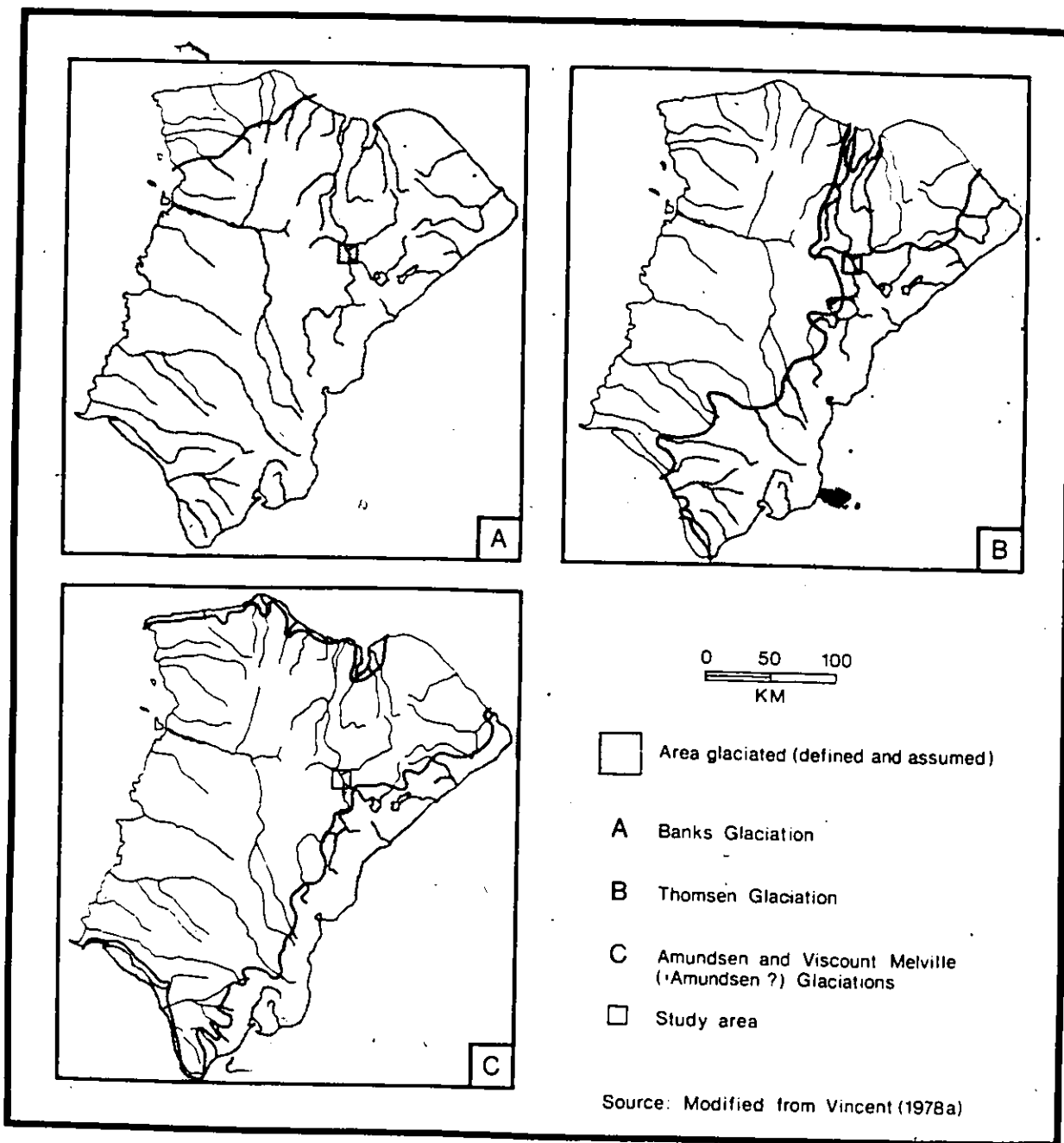


FIGURE 17 : BANKS ISLAND: GLACIAL LIMITS

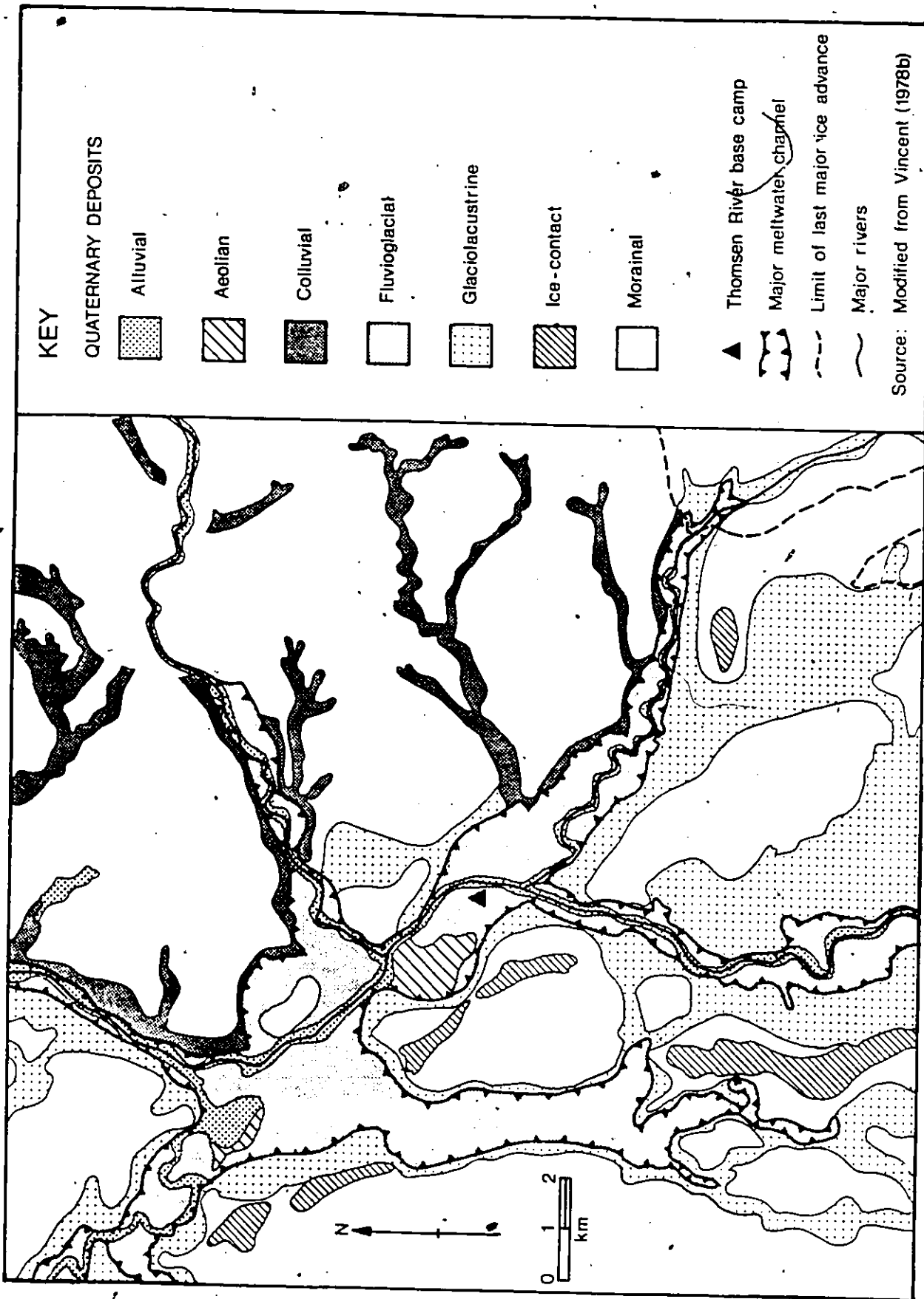


FIGURE 18 : SURFICIAL GEOLOGY OF THE THOMSEN RIVER VALLEY AND SURROUNDING AREA

drainage by the McClure lobe during the Wisconsin advance, and the Thomsen River valley was flooded to an elevation of 90 m (Vincent, 1980, p. 90). The sedimentary evidence to support the hypothesis of a delta is limited because of reworking of the sand by fluvial and aeolian processes. The sand plain at present acts as a source of aeolian sediment during high winds in summer.

### The runoff sites

The locations of the intensive study sites in the Thomsen River area are shown in Figure 12. The four sites, each with a surface runoff plot are illustrated in Figures 19 - 26. The criteria for the choice of the sites are outlined above (pp. 9-10). On each slope the plot was defined by a combination of surface flow observations, topographic surveying, and in some cases by use of plastic edging to divert flow into the collector. The positions of the plots on the slopes remained the same throughout the study at sites 2 and 4, but varied between years at sites 1 and 3.

At site 1, the plot was set up in three different positions across the slope during the study years (Figure 27). The edges of the plot in 1977 were defined by edging, but observations during the melt season suggested that the boundaries were not parallel to the flow paths. This problem was confirmed by surveying at the end of the field season which showed that the edges did not cross the contours orthogonally, and as a result, the position was deemed unsatisfactory for a second year of measurement. In 1978, the surface flow collector was set up at a point which became snow-free early in the spring, but this proved to be a location with a triangular shaped contributory area, implying divergent surface flow in the area upslope. This did not invalidate the results obtained, but the collector was moved



FIGURE 19 : PLOT 1, JUNE 2, 1977

Note break-up of the thin snowpack into discrete pieces and use of plastic edging to divert surface flow to collector. Thaw was particularly early this year (compare snow cover with Figure 20).

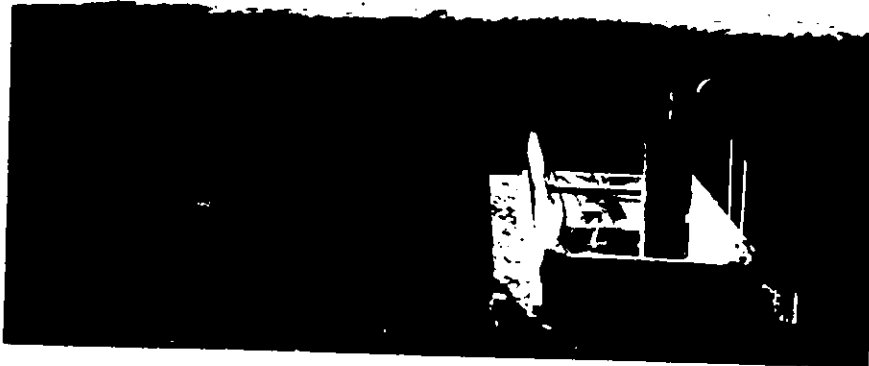


FIGURE 20 : PLOT 1, JUNE 29, 1979

Location of 1977 plot is 15 m west. Note use of V-slot weir and Leupold-Stevens water-level recorder to obtain a continuous record of surface flow.

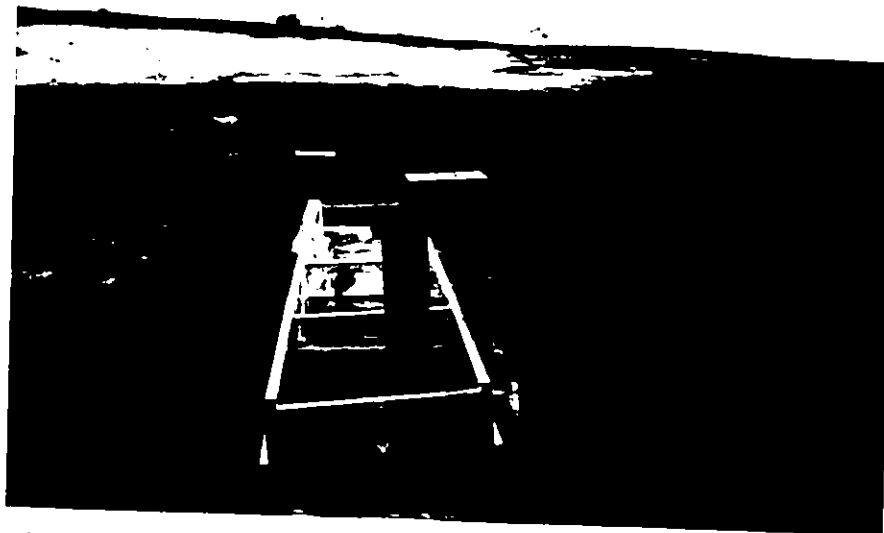


FIGURE 21 : PLOT 2, JUNE 30, 1978

Note Honda tricycle at crest of slope for scale. Photo taken at time of maximum annual water release from the snow on the plot, resulting in extensive surface flow.

(A more general view of site 2 is shown in Figure 8).



FIGURE 22 : SITE 3, JULY 10, 1979

Aerial view south-west with Thomsen River at high stage in the foreground. Note absence of snow cover over most of the upland area while a large snowbank remains at site 3.



FIGURE 23 : SITE 3, MAY 30, 1978

Note thin snow cover on terrace downslope of large snowbank. The terrace snow was lost by evaporation before air temperatures rose above 0°C.



FIGURE 24 : PLOT 3, JULY 20, 1979

Note depletion of the snowbank in comparison with Figures 22 and 23, and the thick basal ice layer. The continuous cover of tundra vegetation, developed in response to long-lasting moisture supply, is clearly shown.



FIGURE 25 : SITE 4, JULY 10, 1979

Aerial view south-west showing the large snowbank that develops in winter below a pre-Wisconsin ice-contact deposit. On this date, only 6% of the upland area remained snow-covered.



FIGURE 26 : PLOT 4, JULY 22, 1979

Note continuous meadow tundra vegetation cover and basal ice layer. Ablation stakes are visible on left of the photograph and surface flow weir on the right.

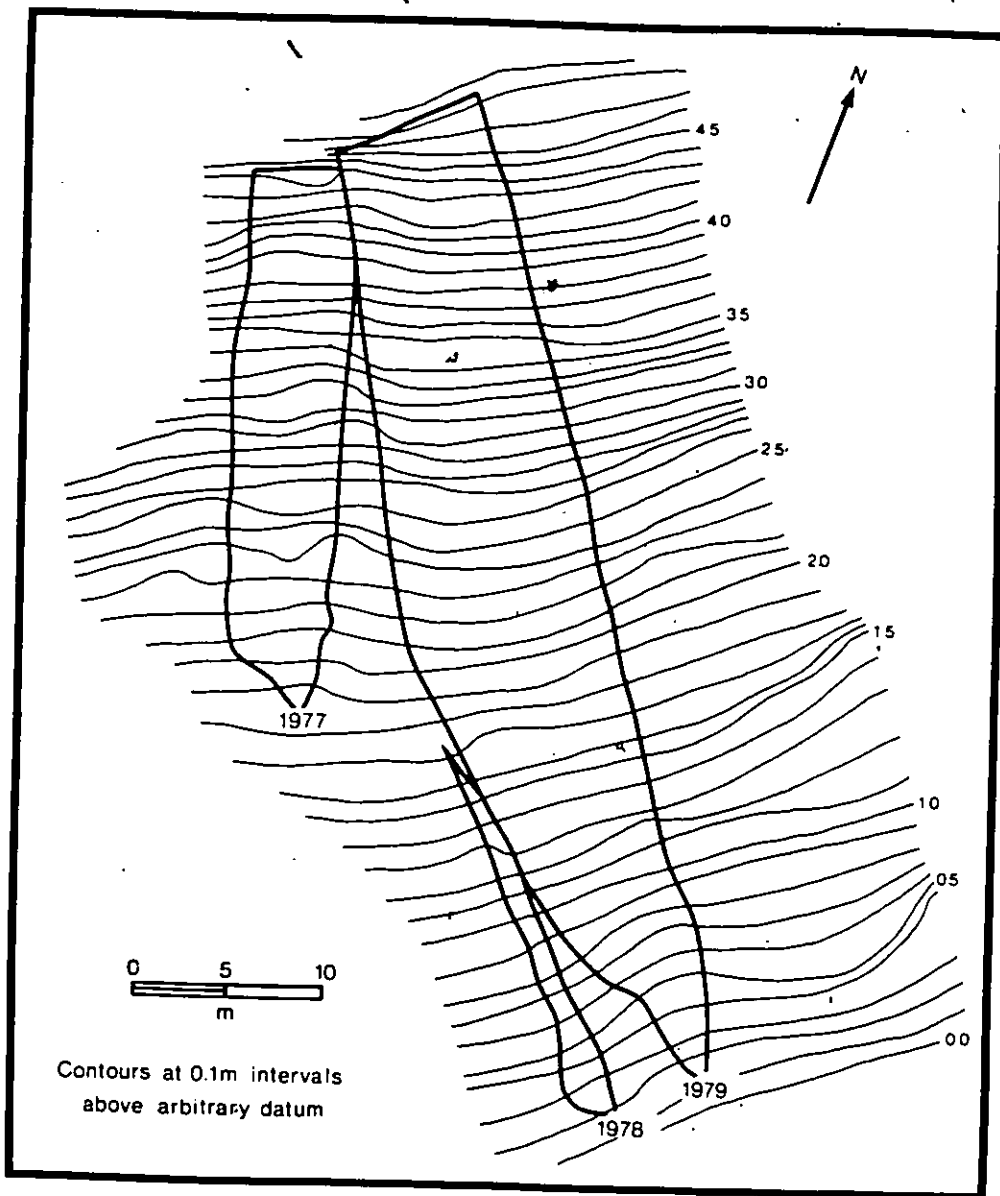


FIGURE 27 : TOPOGRAPHIC MAP OF SITE 1 : POSITIONS OF 1977, 1978 AND 1979 RUNOFF PLOTS.

again for the 1979 season to a small depression to obtain a greater contributory area and higher discharges.

At site 3, the location of the surface flow collector was changed between the 1977 and 1978 field seasons (Figure 28). This move was necessary because the 1977 position had been chosen during the summer after much of the snowbank had ablated. It was too far upslope for monitoring snowmelt during the early part of the ablation season.

The positioning of the plot at site 2 in 1977 was satisfactory and no changes were made in subsequent years (Figure 29). The location of the plot at site 4 was not undertaken until the end of the 1977 season so measurements are available only for 1978 and 1979. The position of this plot remained unchanged for these two years (Figure 30).

The hydrological inputs and outputs of each of the runoff plots are influenced by the plot dimensions, their slope angles, and their aspects (Table 4). Other characteristics, such as snow accumulation, vegetation cover and substrate composition, and particularly any downslope variation of these factors, are relevant to both the hydrologic and geomorphic elements of slopewash processes (see below, Chapter 3). These site-specific characteristics are summarized in Figures 31-34.

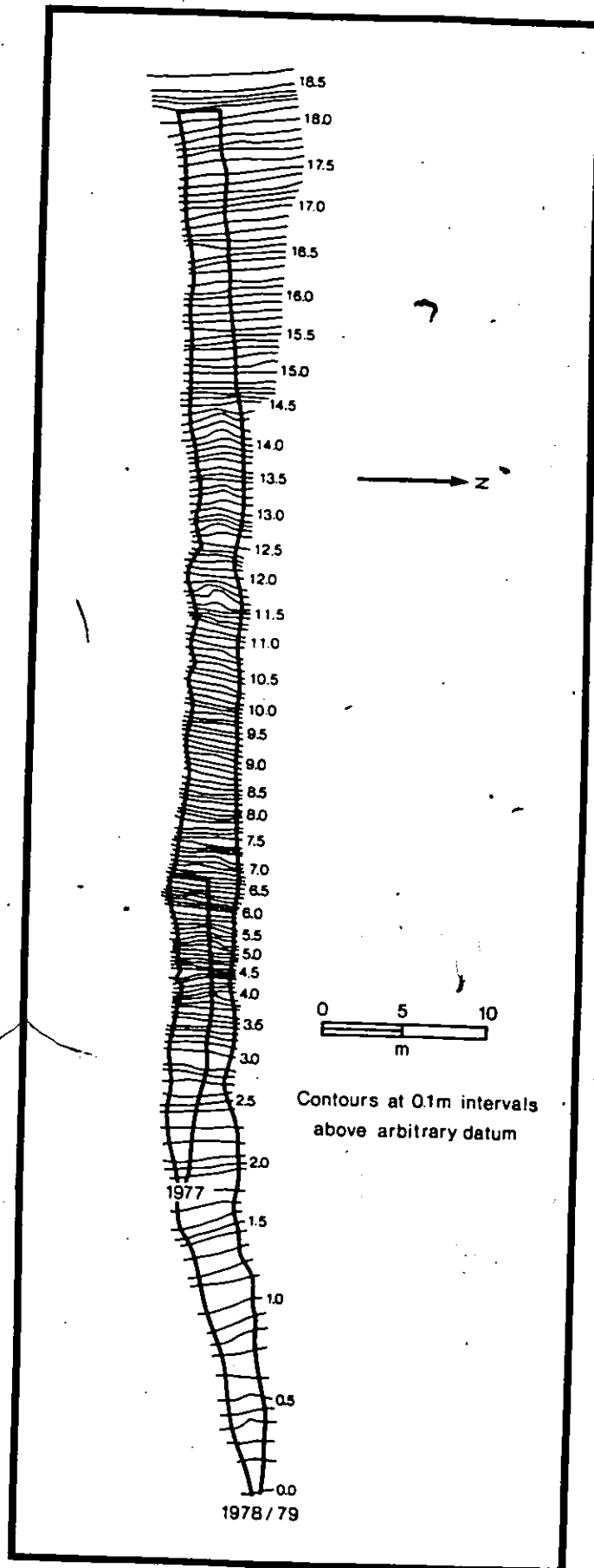


FIGURE 28 : TOPOGRAPHIC MAP OF SITE 3 : POSITIONS OF 1977 AND 1978-1979 RUNOFF PLOTS

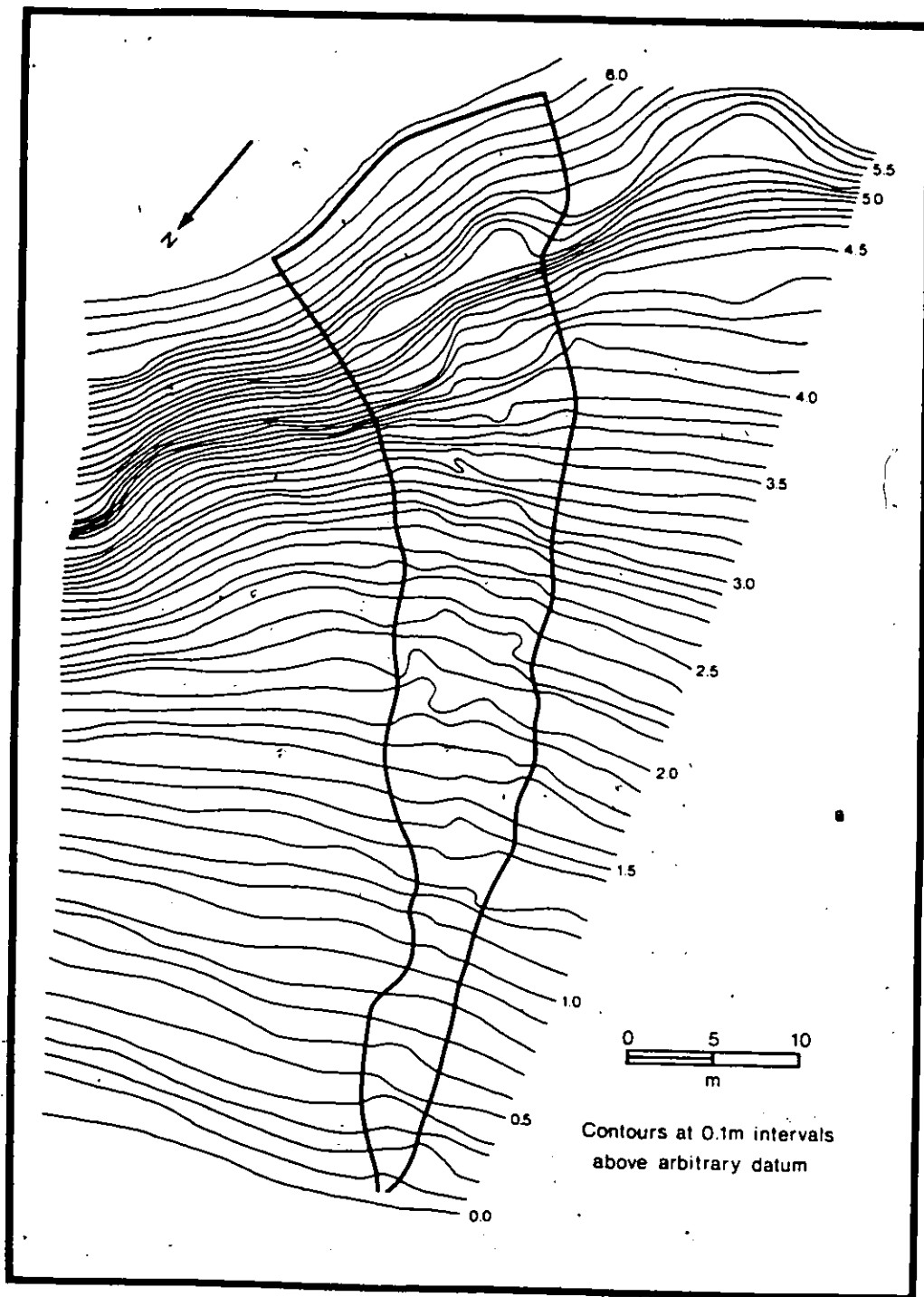


FIGURE 29 : TOPOGRAPHIC MAP OF SITE 2 : POSITION OF RUNOFF PLOT FOR 1977-1979

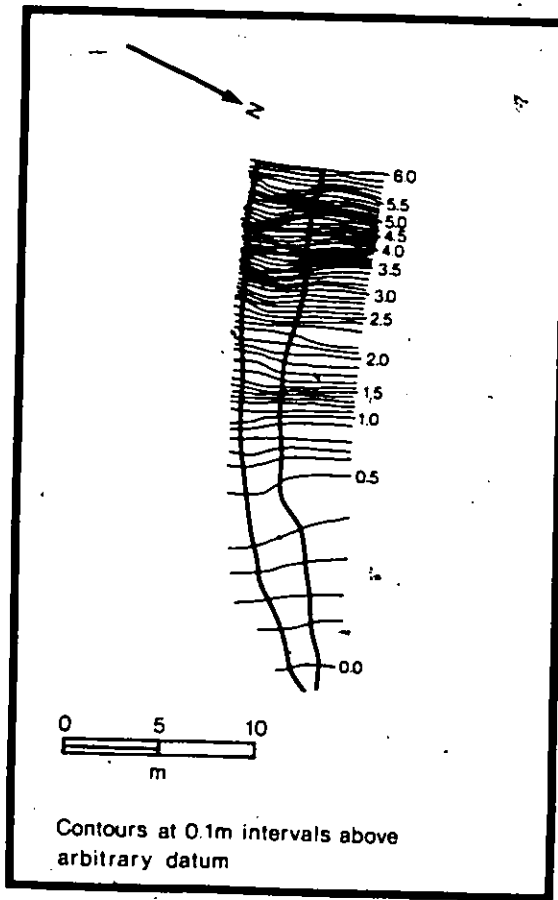


FIGURE 30 : TOPOGRAPHIC MAP OF SITE 4 : POSITION OF RUNOFF PLOT FOR 1978-1979

	Year	Plot 1	Plot 2	Plot 3	Plot 4
Area (m <sup>2</sup> )	1977	*149.6	524.7	26.6	-
	1978	25.0	524.7	262.3	67.0
	1979	437.7	524.7	262.3	67.0
Length (m)	1977	28.7	62.2	18.8	-
	1978	21.0	62.2	92.3	28.1
	1979	52.9	62.2	92.3	28.1
Average Width (m)	1977	5.2	8.4	1.4	-
	1978	1.2	8.4	2.8	2.4
	1979	8.3	8.4	2.8	2.4
Average slope angle	1977 to 1979	5°	5°	12°	13°
Aspect	1977 to 1979	125°	315°	125°	70°

TABLE 4: Dimensions of the runoff plots

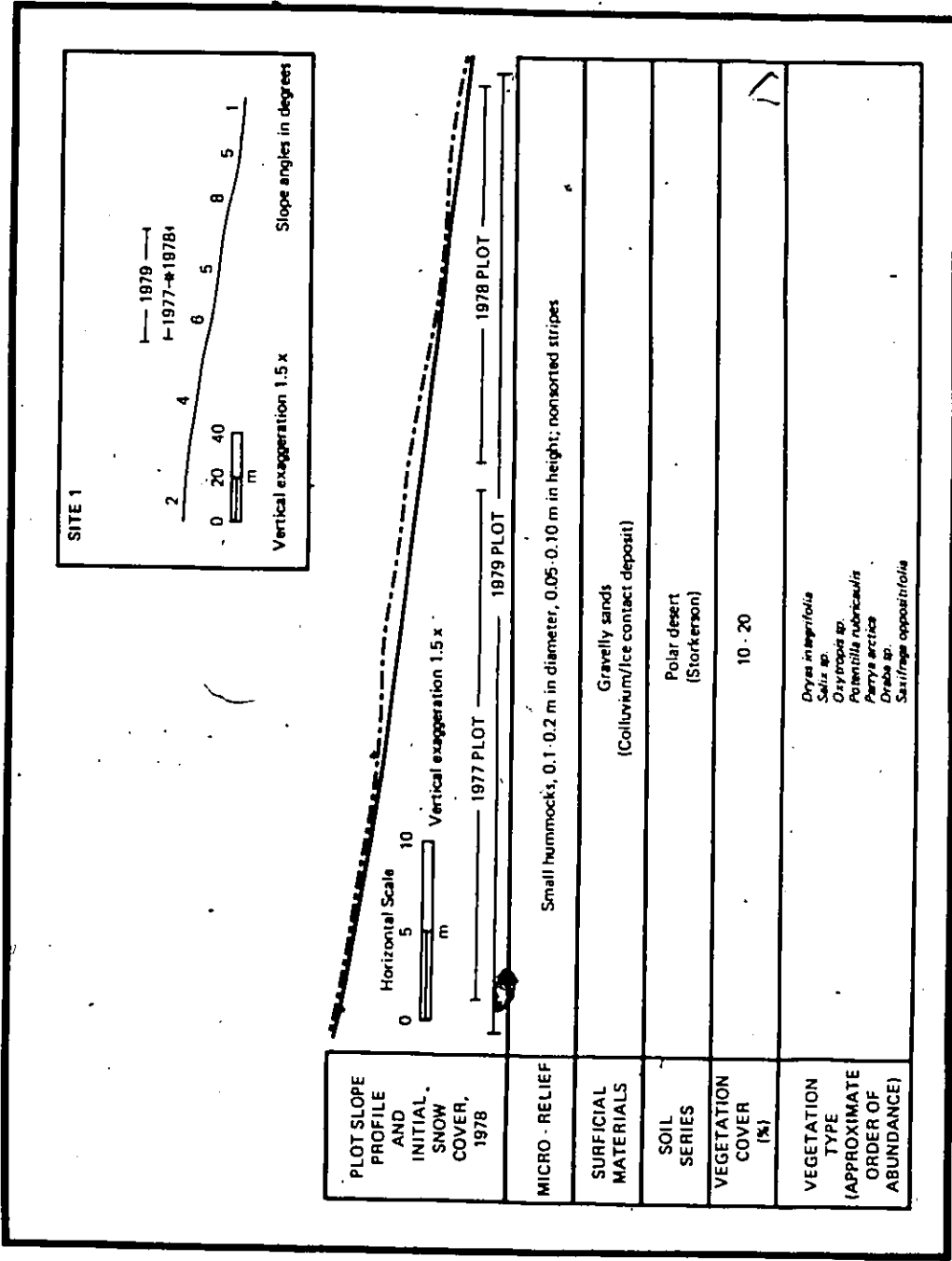
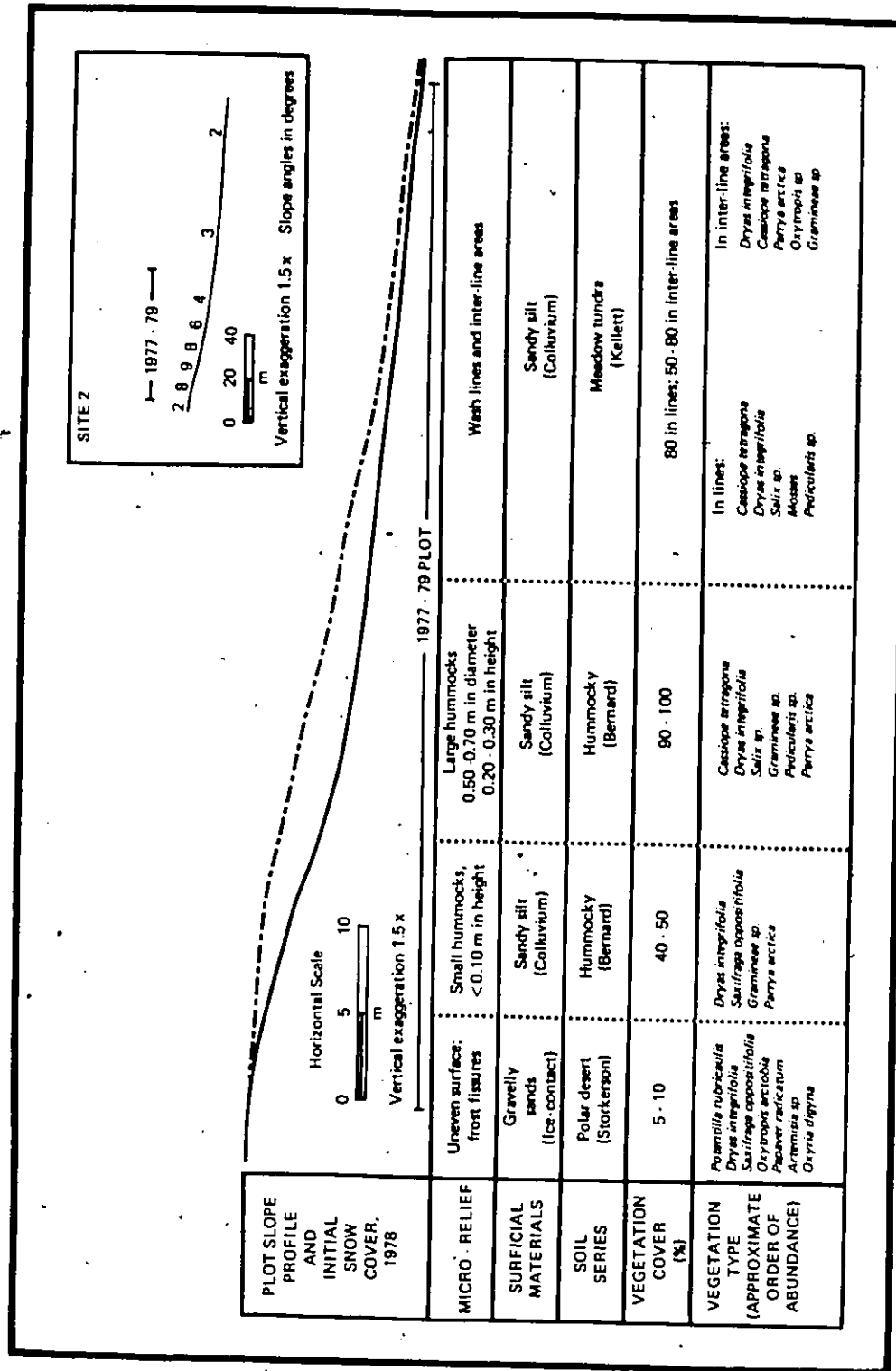


FIGURE 31 : SITE 1 : SLOPE PROFILE AND DOWNSLOPE VARIATION OF PLOT SURFACE CHARACTERISTICS  
Soil series after French (1971; 1976b, p. 200) and Tedrow (1974).



**FIGURE 32 : SITE 2 : SLOPE PROFILE AND DOWNSLOPE VARIATION OF PLOT SURFACE CHARACTERISTICS**  
 Soil series after French (1971; 1976b, p. 200) and Tedrow (1974).

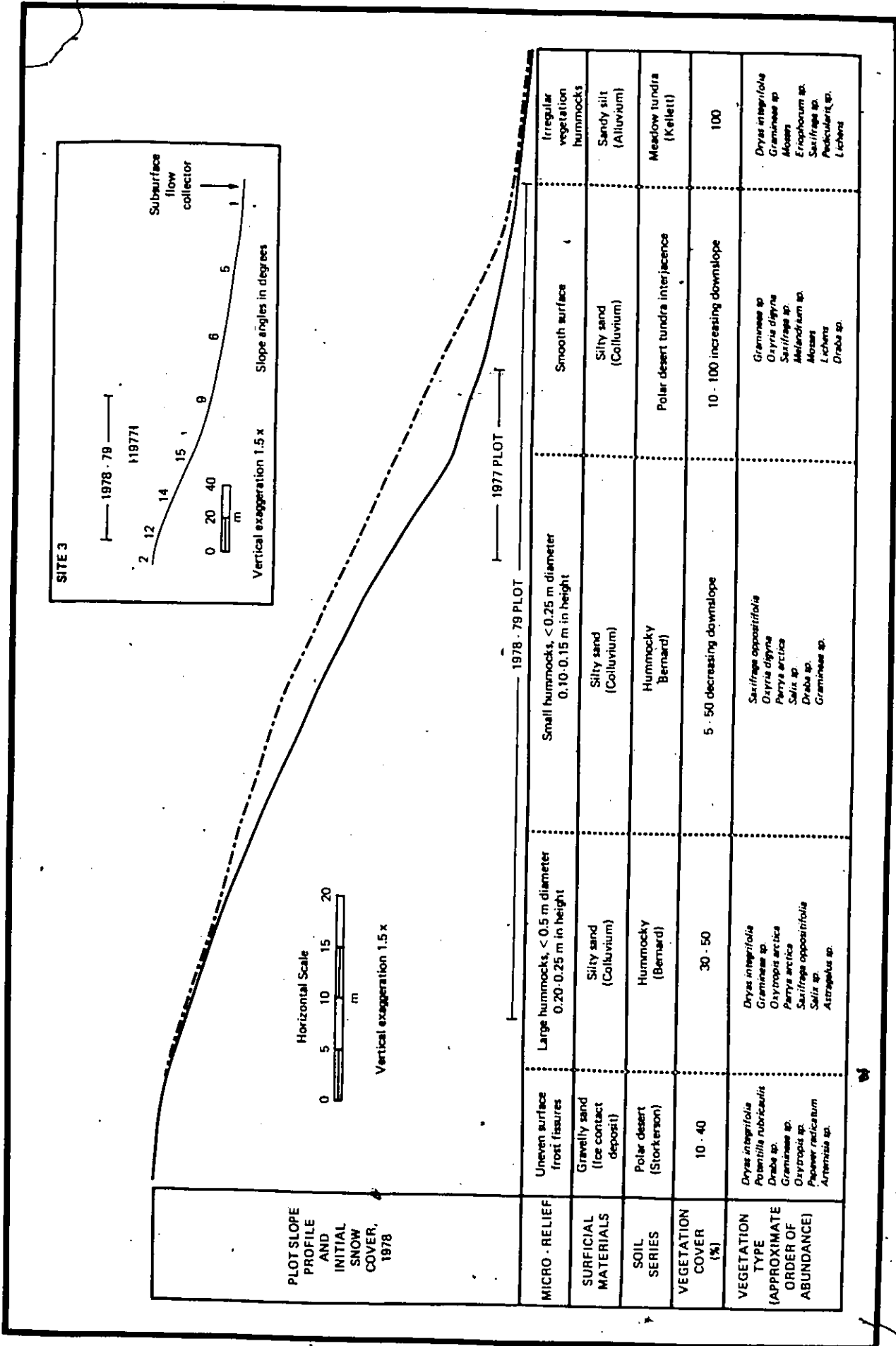
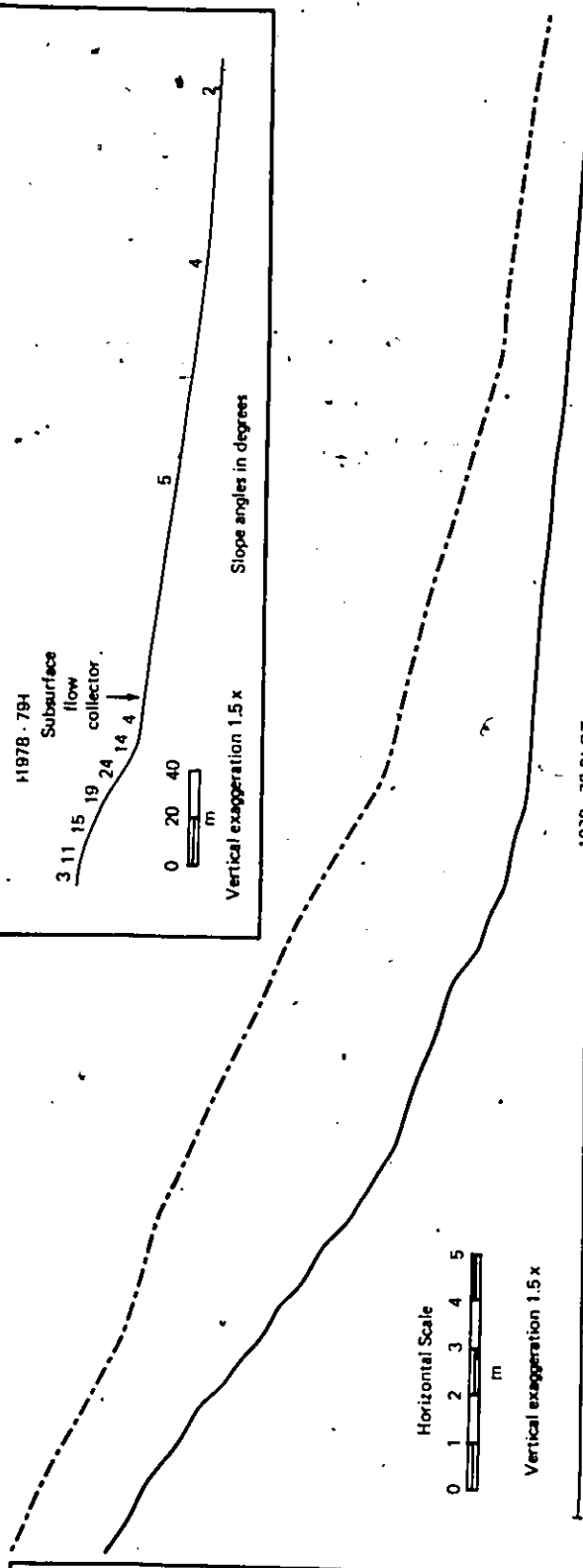
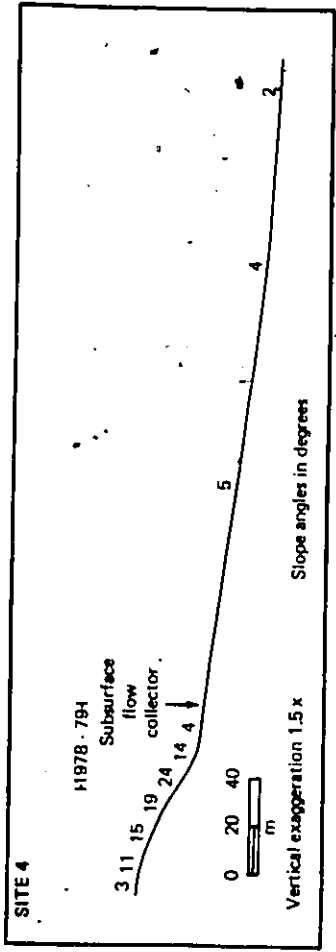


FIGURE 33 : SITE 3 : SLOPE PROFILE AND DOWNSLOPE VARIATION OF PLOT SURFACE CHARACTERISTICS

Soil series after French (1971; 1976b, p. 200) and Tedrow (1974).



<b>MICRO - RELIEF</b>	Large hummocks	Small hummocks	Smooth surface	Irregular vegetation hummocks
<b>SURFICIAL MATERIALS</b>	Silty sand (Colluvium)	Silty sand (Colluvium)	Silty sand (Colluvium)	Silty sand (Colluvium)
<b>SOIL -SERIES</b>	Hummocky (Bernard)	Hummocky (Bernard)	Polar desert tundra interjaccence	Meadow tundra (Kellert)
<b>VEGETATION COVER (%)</b>	20 - 50	10 - 40 decreasing downslope	10 - 100 increasing downslope	100
<b>VEGETATION TYPE (APPROXIMATE ORDER OF ABUNDANCE)</b>	<i>Dryas integrifolia</i> <i>Gramineae</i> sp. <i>Oxytropis</i> sp. <i>Saxifraga oppositifolia</i> <i>Silix</i> sp. <i>Papaver radiculatum</i> <i>Draba</i> sp.	<i>Oxynia digyna</i> <i>Silix</i> sp. <i>Draba</i> sp. <i>Gramineae</i> sp. <i>Parnassia</i> sp. <i>Papaver radiculatum</i> <i>Saxifraga</i> sp.	<i>Gramineae</i> sp. <i>Oxynia digyna</i> <i>Saxifraga</i> sp. <i>Parnassia</i> sp. <i>Pedicularis</i> sp.	<i>Dryas integrifolia</i> <i>Gramineae</i> sp. <i>Mosses</i> <i>Saxifraga</i> sp. <i>Pedicularis</i> sp. <i>Lichens</i>

**FIGURE 34 : SITE 4 : SLOPE PROFILE AND DOWNSLOPE VARIATION OF PLOT SURFACE CHARACTERISTICS**  
 Soil series after French (1971; 1976b, p. 200) and Tedrow (1974).

CHAPTER 3

SLOPEWASH THEORY

### The slopewash system

Three systems diagrams can be used to illustrate the processes and factors controlling the slopewash system in a continuous permafrost area (Figures 35 - 37).

Figure 35 shows the slope hydrological system and is divisible into five component subsystems. Subsystems A, B and C represent the inputs of energy and mass, while subsystem D can be either an input or output depending on meteorological conditions (hence its title "interchange"). Subsystem E represents all the other variables involved in the outputs of the slope hydrological system and includes surface and subsurface flow.

Figures 36 and 37 identify the factors affecting suspended sediment concentrations in surface wash, and dissolved solids in both surface and subsurface wash. These systems are not as readily divisible into subsystems since while the principal variables can be easily distinguished (for example, in Figure 36, these are (1) erodibility of the surface, (2) erosivity of the surface flow, (3) erosivity of the rainfall, and (4) direct aeolian sediment inputs) they are influenced by numerous other factors, some of which affect two or more of the major variables (e.g. slope angle).

Suspended sediment within subsurface wash is not considered in Figures 36 or 37. This process may be significant over long periods (e.g. in forming a soil catena), but has not been shown to be effective in the short-term (Carson and Kirkby, 1972, pp. 232 - 233; Young, 1972, p. 71).

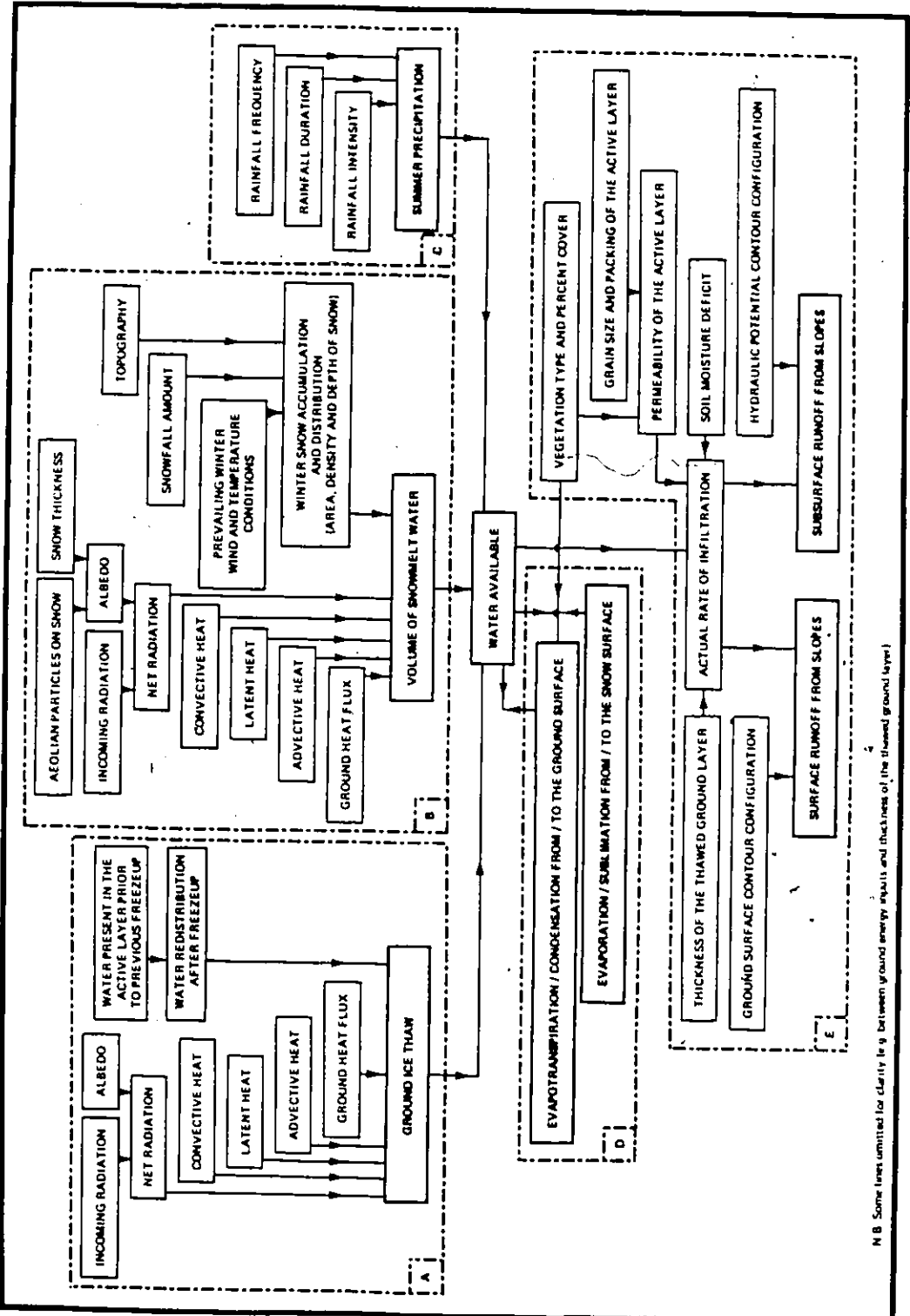


FIGURE 35 : MAJOR FACTORS AFFECTING THE SLOPE HYDROLOGICAL CYCLE IN AN AREA OF CONTINUOUS PERMAFROST

- A - Ground ice supply and energy factors;
- B - Snow supply, energy inputs and snowmelt;
- C - Summer precipitation;
- D - Subaerial interchange;
- E - Surface and subsurface water balance factors.

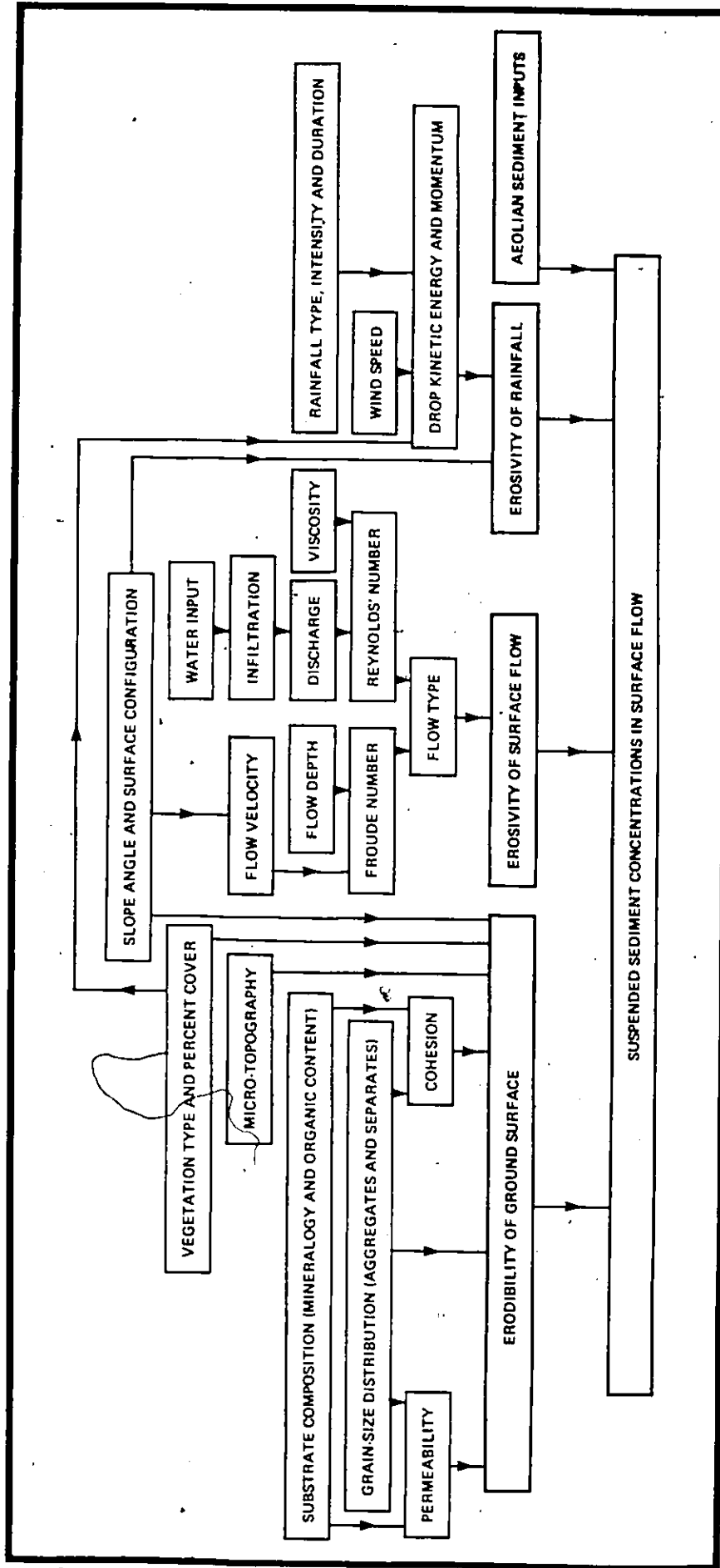


FIGURE 36 : MAJOR FACTORS AFFECTING THE CONCENTRATION OF SUSPENDED SEDIMENT IN SURFACE WASH IN AN AREA OF CONTINUOUS PERMAFROST

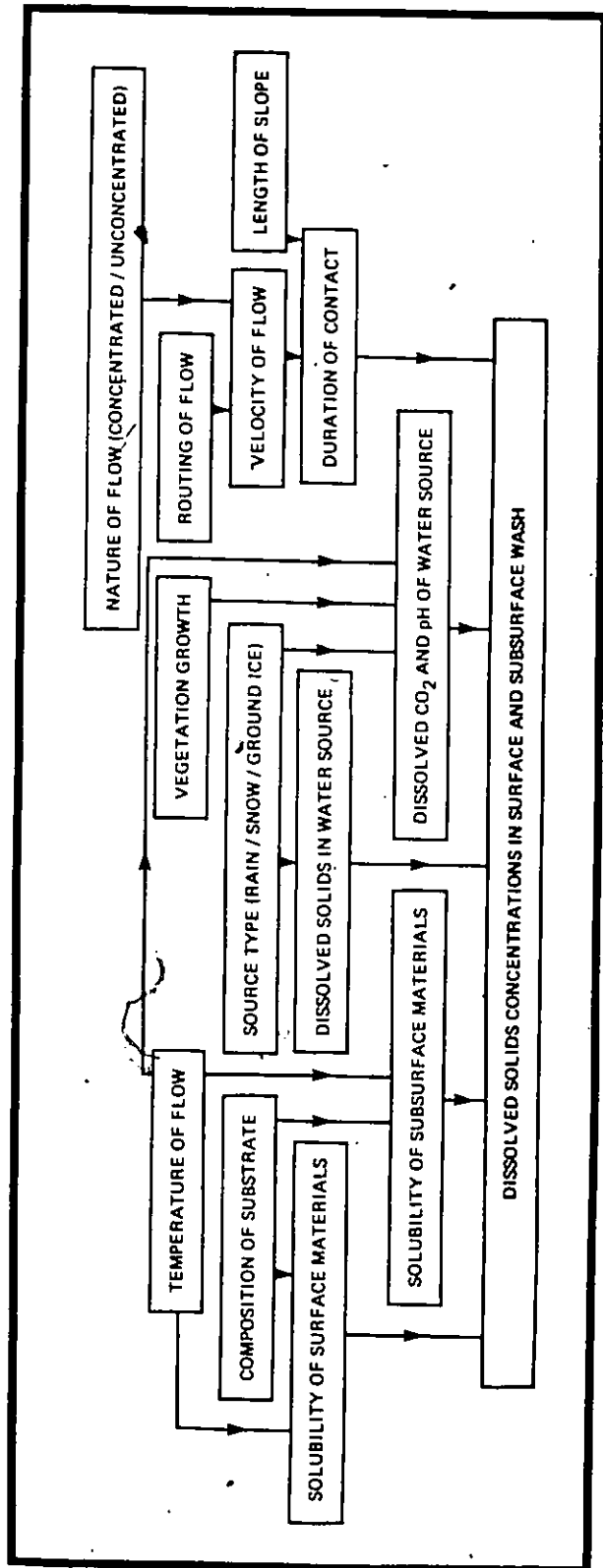


FIGURE 37 : MAJOR FACTORS AFFECTING THE CONCENTRATION OF DISSOLVED SOLIDS IN SLOPEWASH IN AN AREA OF CONTINUOUS PERMAFROST

The related processes of "thermo-erosional wash" and "thermo-planation" (Dylik, 1972; Kachurin, 1962) await quantitative assessment.

### Hydrological theory

With the exceptions of (1) the presence of permafrost itself, and (2) ground ice thaw, the factors affecting slopewash hydrology in a permafrost area (Figure 35) are the same as those acting in a nonpermafrost location. Within the permafrost milieu, however, each azonal factor acts with a different frequency and at a different magnitude relative to its nonpermafrost counterpart.

The slope hydrological factors are discussed below, grouped into the subsystems shown in Figure 35, following a description of the hydrological effects of permafrost.

#### (a) Hydrological effects of permafrost

The presence or absence of permafrost is crucial to slope hydrology, but is little mentioned in standard texts (e.g. Kirkby, 1978a). Most hydrological studies assume that permafrost acts as an impermeable barrier, so that water balance and streamflow calculations can ignore deep groundwater inputs and losses (e.g. Anderson and MacKay, 1974; Church, 1972; Landals and Gill, 1973; Marsh and Woo, 1977; McCann et al., 1972).

On slopes, lateral flow generally is assumed to be confined to the surface and to within the active layer. In addition, however, flow may take place within the permafrost. Some water in the ground remains unfrozen at

temperatures below  $0^{\circ}\text{C}$  because of the presence of dissolved salts and/or pore pressures greater than one atmosphere (e.g. Anderson and Morgenstern, 1973; Koopmans and Miller, 1966; Nerseova and Tsytovich, 1966; Tice et al., 1978; Williams, 1963; 1964). Moreover, frozen soils possess low but distinct hydraulic conductivities which are temperature-dependent within a few degrees of  $0^{\circ}\text{C}$  (Burt and Williams, 1976; Harlan, 1974). If the active layer is underlain by "dry" permafrost in which pore ice is absent, movement of water may take place through a thin upper layer whose temperature is close to  $0^{\circ}\text{C}$  (Dingman, 1975). In this case, there is no distinct hydrological boundary between areas of water movement and stasis, but rather a zone in which hydraulic conductivity decreases from some finite value at the thawing plane down to zero at depth. If on the other hand, the upper part of the permafrost is saturated with pore ice or contains excess ice, no infiltration can take place because the void spaces in the frozen sediments are already full (Williams and van Everdingen, 1973). In this, the more common case, infiltration is restricted to the seasonally thawed zone, with four important consequences for slopewash processes.

First, water movement is confined to a relatively thin layer of thawed ground, which if composed of fine sediments, may possess low hydraulic conductivity. Surface flow can be produced under conditions of water inputs (snowmelt or rainfall) which would be insufficient to generate it in temperate catchments where totally impermeable subsurface strata are rare. Since routing of water to the surface allows it to pick up and transport sediment, depending on the other factors in the system (Figure 36), this may have significant erosional effects.

Second, hydrologic conditions are not constants in a permafrost area. As the thawed layer becomes thicker during the summer period, it allows more water to move as subsurface flow, because the rate of subsurface discharge is proportional to the cross-sectional area available (if conductivity and hydraulic head are constant, Darcy's Law).

Third, in most continuous permafrost areas, despite the insulating effects of the snow cover, the active layer freezes completely in winter and frozen ground exists immediately beneath a snowbank. Any upslope derived subsurface flow is forced to the surface by this barrier of frozen ground (Ballantyne, 1978; Dutkiewicz, 1967) with consequent effects on sediment transport.

Fourth, saturation of the substrate by summer rainfall (assuming rainfall intensity does not exceed infiltration capacity) and resultant surface flow generation, can be achieved after relatively low precipitation totals in a permafrost environment since only the active layer void spaces require filling. Summer precipitation in the continuous permafrost zone, however, is often extremely light. Despite the influence of permafrost, it may be rare for surface flow to be generated by this process (Lewkowicz et al., 1978).

(b) Hydrological effects of ground ice thaw  
(Subsystem A, Figure 35)

At an annual time-scale, the effects of ground ice thaw on the water balance of a catchment can be deemed insignificant (e.g. Marsh and Woo, 1979; Woo, 1976b). This is permissible because if active layer

thicknesses and soil moisture conditions are similar from one freezeup season to the next, storage changes are small. The same principle applies to slope water balance studies, and on a yearly basis ground ice thaw can be ignored.

Over time periods of less than a year, the thawing of ground ice may have some effect on the slope water balance. The total volume of water present as ice in the active layer prior to spring thaw is equivalent to, or less than that present during freezeback the previous autumn. Considerable spatial variation in ground ice distribution is likely, therefore, with larger amounts on slopes which have been supplied with water for most of the summer by long-lasting snowbanks, and much smaller quantities on interfluves where soil moisture deficits exist prior to freezeup.

In addition to variation in plan, ice is unevenly distributed vertically within the active layer, as a result of water migration towards the freezing planes during the freezeup period. Ice lenses grow preferentially at the top and bottom of the active layer (e.g. Gell, 1974; Lewellen, 1972; Vtyurina, 1973) with a desiccated area left in between (Mackay and MacKay, 1976). This redistribution is particularly noticeable in fine-grained mineral soils and may have geomorphological effects (e.g., the formation of mud hummocks (Mackay, 1979; Mackay and MacKay, 1976)). The process is less important in coarse-grained mineral or organic soils which possess low unsaturated hydraulic conductivities (Slaughter and Kane, 1976). A further influence on the vertical distribution of ice in the active layer is the potential diffusion of water from the top 10 - 20 mm of the ground into the overlying snowpack, during the winter period (Santeford, 1978).

The information concerning the distribution of ice within the active layer can be combined with knowledge of the near-asymptotic nature of thaw layer development (Terzaghi, 1952) to produce an hypothesis concerning the variation of ground ice hydrological inputs to a slope during the summer. It is thought that the thawing of ground ice may be important during the first one or two days of active layer development (Journaux, 1976) when the thawed layer achieves typical thicknesses of 0.10 - 0.15 m and the ice lenses close to the ground surface melt. For much of the remainder of the summer, the desiccated zone within the active layer thaws and ground ice contributions probably are insignificant. Near the end of the summer, the second, lesser, concentration of excess ice is encountered by the thawing plane, but thaw rates are low and it is likely that only small quantities of water are released daily.

No field evidence is available to support the hypothesis of ground ice hydrological inputs at the beginning of active layer thaw, probably because the amounts of water released are small in comparison to those produced by contemporaneous snowmelt. The small contributions to streamflow produced by the melting of segregated ice at the end of the summer, have been noted in one catchment study (Wedel et al., 1978), but not in others (e.g. Brown et al., 1968; Dingman, 1971).

Estimates of the daily inputs of ground ice thaw to a slope can be made using predictions of active layer thickness and ground ice distribution. The rate at which the active layer thickens is predictable using a computer program based on an equilibrium surface temperature solution (e.g. Outcalt, 1972; Smith, 1975; 1977). Most programs are

point-predictive, however, and until recently little work has been undertaken on two-dimensional modelling of subsurface heat flow (exceptions are French and Smith (1980); Hwang (1976); Smith and Hwang (1973)). No model is known that can simulate active layer thaw, and thereby ground ice thaw, along a slope transect possessing varying surface properties and micro-climatic inputs. A simple equation, however, can be derived from Terzaghi's (1952) formula, and is suitable for prediction on slopes morphologically similar to sites 2, 3 and 4 at Thomsen River (Figure 38):

$$V_n = P \cdot S \cdot B [ ((n)^{0.5} - (n-1)^{0.5}) r_1 + ((n-1)^{0.5} - (n-2)^{0.5}) r_2 + \\ ((n-2)^{0.5} - (n-3)^{0.5}) r_3 + \dots + ((2)^{0.5} - (1)^{0.5}) r_{n-1} + r_n ] \quad \dots (1)$$

where:  $V_n$  is the volume of ice in  $m^3$  thawed on day  $n$ , per  $m$  of slope width;  
 $P$  is the average porosity of the active layer (dimensionless);  
 $S$  is the coefficient of saturation (dimensionless);  
 $B$  is the empirical coefficient from Terzaghi's (1952) equation, which varies from 0.07 to 0.15 depending on the type of soil and its moisture content;  
 $r_n$  is the upslope distance retreated by the downslope edge of the snow cover on day  $n$  (m).

Inputs from areas upslope of the snowbank are not considered in equation (1) and it is assumed that snow ablation takes place in discrete steps from the base of the slope, with no active layer development before an area is clear of snow. A uniform distribution of ice through the active

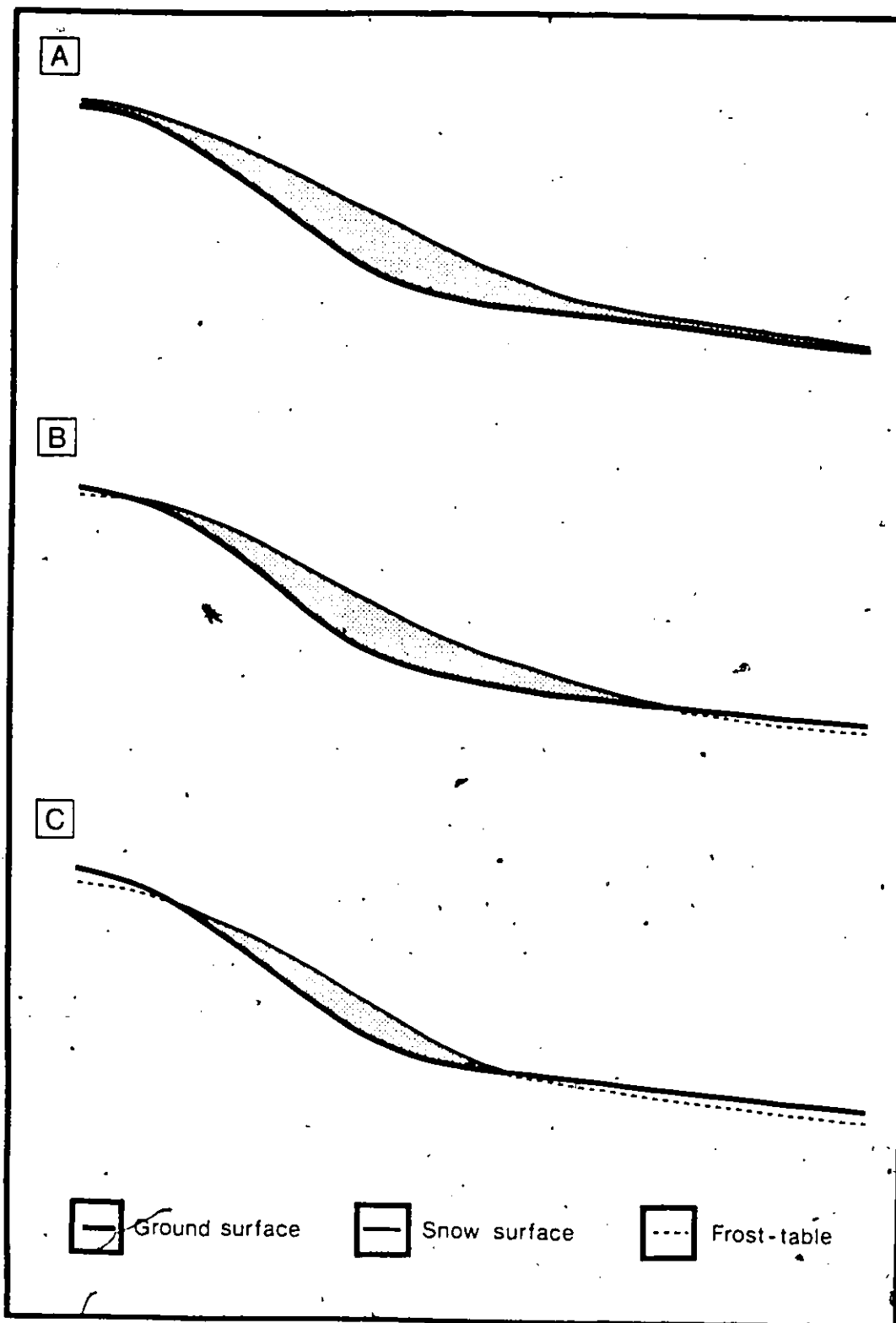


FIGURE 38 : SCHEMATIC DIAGRAM SHOWING TYPICAL SNOWBANK AND GROUND THAW PROGRESSION, THOMSEN RIVER AREA

layer is assumed, although the equation could be adapted to incorporate field data on vertical ice distributions.

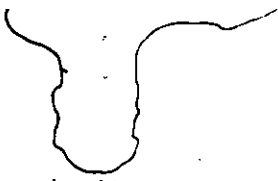
The discussion above concerns average conditions for ground ice thaw on a slope. It should be noted, however, that catastrophic thermokarst may also contribute to a slope's water balance. The top of the permafrost is often ice-rich due to the presence of ice wedges and ice veins (e.g. Babb, 1977; Mackay, 1971; 1972; 1974; Price et al., 1974; Tarnocai and Zoltai, 1978; Williams, 1968) and is particularly vulnerable to thermokarst following changes in the ground thermal regime. Inputs to the hydrologic system, however, depend on site-specific conditions such as the forms of segregated ice masses and the energy available for melt, so that generalisations are not possible. Since thermokarst was not observed operating on the slopes in the Thomsen River area during the three field seasons, the extensive literature on the process is not discussed further.

(c) Snow supply, energy inputs and snowmelt

(Subsystem B, Figure 35)

In most non-glacierized basins of the continuous permafrost zone, snowmelt produces the highest stream discharge of the year. Snowmelt also is responsible for most of the total annual stream runoff and the resultant typical nival flow regime (e.g. Anderson, 1974; Church, 1972; 1974; Cook, 1967; Marsh and Woo, 1979; McCann and Cogley, 1973; Robitaille, 1960; Ryden, 1977; St-Onge, 1965).


The significance of snowmelt to slopewash depends on (1) the amount and distribution of snow available for melt; (2) the energy supplied



to the snow; and (3) the timing of water movement through the pack and the rate of outflow from the snow.

North of the tree-line, the development of the winter snowpack is determined by interactions between topography and wind. Snow is redistributed by wind, and drifts build up in depressions and behind flow obstructions. Variations in winter conditions cause only minor changes in the exact position of snowbank faces and in snow depths. Typical locations for snow build-up are gullies, valley bottoms and concave segments of valley-side slopes. Contrarily, accumulation on ridges and hill-tops depends upon the roughness of micro-topography and vegetation, and in polar desert areas, interfluves may be swept almost bare in the winter period.

The total snow accumulation within a catchment affects slopewash activity. In years of average or below average winter snowfall, surface wash may be confined to areas downslope of major snowbanks. In other years, surface wash may be more extensive. Evaluation of the water equivalent of the winter snowpack, however, is a problem. Standard measurements made at meteorological stations are unreliable, ranging from underestimates of 25 - 30% in the subarctic and Alaska (Hare and Hay, 1971; Lewellen, 1972), to 65% in the high Arctic (Marsh and Woo, 1979; Woo and Marsh, 1977; 1978). One solution is the subdivision of catchments or larger areas into landscape units (such as hilltops, low flats, long slopes, gullies and valleys) each characterised by distinctive average depths and densities (Gray et al., 1979; Woo and Marsh, 1978). Other viable methods of snow accumulation assessment involve measurement of ablation rates and mapping areas as they become free of snow (see Chapter 5, pp. 143-144).



Winter snow density depends on the initial density of the particles, the packing force (i.e. wind strength), and the amount of snow metamorphosis that takes place. The abrasion of snow particles during transport and their small size, results in high densities for the snowpack north of the tree-line. An empirical equation developed by Billelo (1969) relates average snow density to air temperature and wind speed:

$$\rho_s = 0.152 - 0.0031 T + 0.019W \quad \dots(2)$$

where:  $\rho_s$  is the snow density ( $\text{Mg m}^{-3}$ );  
 $T$  is the average seasonal air temperature ( $^{\circ}\text{C}$ );  
 $W$  is the average seasonal wind speed ( $\text{m s}^{-1}$ )

The normal range of winter snow densities in tundra areas is  $0.35 - 0.40 \text{ Mg m}^{-3}$ . (Slaughter and Crook, 1974) with higher values for snowbanks.

Substituting meteorological data from Sachs Harbour (Table 1) into equation (2) gives an average snow density of  $0.323 \text{ Mg m}^{-3}$ . In reality, a single value for snow density is an oversimplification since it may vary spatially. Moreover, densities tend to increase during the winter period (e.g. from  $0.320 \text{ Mg m}^{-3}$  to  $0.380 \text{ Mg m}^{-3}$  at Resolute Bay, Cornwallis Island, N.W.T. (Williams, 1957)).

The energy supply to the snowpack can be assessed by measuring or estimating the sum of all energy inputs and losses. The pack reacts in two ways to the energy inputs: first, by warming, and second, by melting.

Before melt, the energy absorbed by the pack is used to warm the snow surface. Heat is transmitted to the underlying pack by simple conduction, vapour diffusion, and by meltwater produced at the surface

refreezing at depth and releasing latent heat. The water equivalent of the snow remains constant except for evaporation losses, or gains from sublimation.

After the pack becomes isothermal at  $0^{\circ}\text{C}$ , meltwater produced at the surface is able to leave without refreezing. The heat balance of the snow is given by:

$$H_m = H_r + H_c + H_e + H_p + H_g \quad \dots(3)$$

where:  $H_m$  is the heat available for melting snow;  
 $H_r$  is the net radiative heat flux;  
 $H_c$  is the sensible heat flux;  
 $H_e$  is the latent heat flux;  
 $H_p$  is the heat gained from precipitation;  
 $H_g$  is the ground heat flux.

All terms are in  $\text{W m}^{-2}$ .

In equation (3),  $H_g$  may be assumed to be zero (Heron and Woo, 1978; Male and Gray, 1975; Price and Dunne, 1976; Weller and Holmgren, 1974) and if rain does not fall during the melt period,  $H_p$  also is zero. Thus, the energy balance simplifies to three terms -  $H_r$ ,  $H_c$  and  $H_e$ .

The net radiative heat flux to the snow surface,  $H_r$ , is the balance of incoming and outgoing radiation:

$$H_r = (Q + q) (1 - \alpha) + L_n \quad \dots(4)$$

where:  $Q$  is the direct beam solar radiation ( $\text{W m}^{-2}$ );  
 $q$  is the diffuse solar radiation ( $\text{W m}^{-2}$ );  
 $\alpha$  is the snow albedo (dimensionless);  
 $L_n$  is the long-wave radiation balance ( $\text{W m}^{-2}$ ).

The relative contributions of  $Q$  and  $q$  at any one site and time, depend upon meteorological conditions and are independent of the snow surface. In comparisons between sites, however, aspect, slope angle, latitude and solar declination-all influence the amount of energy received per unit area and the timing of its variation. Moreover, ablation of snowbanks may cause changes both in their angles and azimuths relative to the incoming radiation (Peterson, 1974). These changes affect the direct beam component and are important in the subarctic where differences in aspect can alter daily totals of solar radiation by 13% (Price and Dunne, 1976). They are, however, less significant in the high Arctic (e.g. Heron and Woo, 1978).

The albedo,  $\alpha$ , of fresh snow, usually ranges between 0.80 and 0.85 (Geiger, 1959, p. 129), but as the snow undergoes metamorphosis the albedo decreases to a typical range of 0.42 - 0.70. Relationships exist between  $\alpha$  and snow density and grain-size (i.e. snow aging), cloudiness and sun angle (e.g. Bergen, 1975; Petzold, 1977). At low sun angles on days with high  $Q$  values, for example, the snow albedo may be increased by specular reflection, which may account for maximum  $\alpha$  values of 1.0 measured at Truelove Lowland, Devon Island, N.W.T. (Courtin and Labine, 1977, p. 88).

As snow ablates, any aeolian sediment within the pack appears at the surface and together with any summer sediment input, causes a substantial reduction in surface albedo. The reduction is site-specific and dependent on the sediment concentrations, but by the end of the ablation season the snow surface may appear quite dark, with an  $\alpha$  value as low as 0.24 (Peterson, 1974).

A further seasonal change in albedo occurs as the snowpack thins. Studies in Alaska and elsewhere during the melt season, indicate that approximately 20% of the energy transmitted from the snow surface reaches a depth of 0.15 m (e.g. Gerdel, 1948; Weller and Holmgren, 1974). Furthermore, the albedo of the underlying terrain has little effect on the energy absorbed (O'Neill and Galbraith, 1979). In the high Arctic, therefore, radiation penetration is important in the thin snowpacks of ridges and hilltops, and at the edges of snowbanks where the snow is less than 0.15 - 0.20 m thick.

The long-wave radiation balance,  $L_n$ , is dependent on emission from the atmosphere and return radiation from the snow. In tundra areas,  $L_n$  is important because of the magnitude of the terms involved in equation (4) (Price, 1977, p. 16). If absorbed short-wave radiation amounts are low, long-wave terms become relatively large and small percent differences in  $L_n$  translate into large over- and under-estimates of  $H_r$ .

Downward clear-sky long-wave radiation can be estimated using either empirical (e.g. Brunt, 1952; Brutsaert, 1975; Swinbank, 1963) or physically-based equations (e.g. Idso and Jackson, 1969). These can be modified for cloudy skies to give the long-wave balance (e.g. Smith, 1975). An alternative method is simply to measure the all-wave balance with a net radiometer linked to a recording device. Unfortunately, these instruments are particularly difficult to maintain in the field.

The sensible and latent heat fluxes over snow,  $H_c$  and  $H_e$ , depend on the gradients of air temperature, vapour pressure and

wind speed above the snow surface, and on variations in the surface roughness. These gradients differ from those normally found over ground surfaces because conditions over snow usually are stable, rather than neutral or unstable (Heron and Woo, 1978; Price and Dunne, 1976). Changes in the atmospheric gradients occur on both daily and seasonal bases, and affect snowmelt and slopewash processes at the same time-scales.

Diurnal cycles of air temperature are "driven" by radiation cycles with a lag of 0.5 - 1.5 hours. If the snow is isothermal, the magnitude of  $H_C$  is directly related to changes in air temperature. Since air temperatures tend to increase during the melt season, ablation also tends to become more rapid.

Wind speeds above the snow surface, which influence the sensible and latent heat transfer coefficients, are related to overall meteorological conditions, wind direction, and to changes in the snow surface. On a lee slope, for example, wind speeds above the snow tend to decrease as the snow ablates, reducing energy inputs from sensible heat, but also reducing latent heat losses.

The roughness of the snow surface increases during the melt season as the crystal structure changes. Particularly rapid changes occur when vegetation begins to emerge through the snow (Price, 1977, pp. 77 - 79; Weller et al., 1972) and these cause increases in turbulent transfer.

The complexity of the energy balance of snow is revealed in measurements of  $H_r$ ,  $H_c$  and  $H_e$ . At Barrow, Alaska, radiation proved to be the most important energy source for snowmelt, providing 90-100% of the energy inputs (Weller et al., 1972; Weller and Holmgren, 1974).

Sensible heat transfer was far less important and contributed 0% during the ripening period and 10% of the inputs during melt. Evaporation used up only 1-13% of the available energy and the remainder was used to melt snow (Weller and Holmgren, 1974). Figures similar to these were obtained on Axel Heiberg Island, N.W.T. (Ohmura and Muller, 1976) where  $H_r$  contributed 100% of the energy during melt. Other data from northern areas support these findings, e.g.  $H_r$  contributed 90% of the energy for melt in an alpine catchment in northern British Columbia (Jordan, 1979), and 98% in a forested area in southern Alaska (Storr, 1978).

In one study in a permafrost area, however, the proportions of the energy balance components recorded are at variance to those outlined above. At a catchment on Cornwallis island, N.W.T.,  $H_r$  contributed only 58% of the energy to the snowpack and  $H_c$  supplied 42% (Heron and Woo, 1978).  $H_e$  was an important energy loss prior to ripening of the pack, but became progressively less significant as melt continued. Overall, 80% of the energy to the snow was used for melt, and 20% for evaporation. These data are quite similar to others obtained from locations in the Canadian Prairies where  $H_r$  is dominant and  $H_c$  unimportant when the snow cover is complete (93% and 7% respectively) and  $H_r$  is of lesser significance when the cover becomes patchy ( $H_r$ : 56%,  $H_c$ : 44%) (Gray and O'Neill, 1974).

It is not clear which of the two proportions of energy balance components is the more common in a permafrost area. Two generalisations, at least, apply. First, in all cases, radiation is the most important energy input to the snowpack. Second, a change in the energy balance occurs

between periods when a complete snow cover exists and when it becomes patchy. This latter is problematic if a predictive model based on the energy balance is desired, because of the breakdown in boundary conditions and the importance of local advective heat inputs (e.g. Heron and Woo, 1978).

The interaction of the snow present on a slope and the energy supplied to it, is revealed in the movement of water within the snow and downslope from it. In areas of continuous permafrost, infiltration beneath the snowpack (see above, pp. 62-64) may be eliminated by the presence of frozen ground. In this case, the movement of water within the snow can be divided into two phases: (1) vertical percolation from the snow surface, and (2) downslope movement at the base of the snowpack.

Measurement of vertical percolation has been attempted using snow lysimeters (e.g. Anderson, 1968; Fohn, 1973; Jordan, 1979; Price, 1977; Woo and Slaymaker, 1975), but downslope flows have proved more difficult to measure without destroying the pack. An alternative approach, involving physically-based predictive modelling has been developed by Colbeck (1971; 1972; 1973; 1974a; 1974b; 1975; 1977). His equations have been partially tested in an alpine area (Jordan, 1979) and more fully in the subarctic (Dunne et al., 1976; Price, 1977).

In essence, Colbeck's model treats the vertical movement of water as unsaturated flow through a porous medium, and horizontal water movement through the snow as saturated flow. The assumptions of snow homogeneity within the saturated and unsaturated zones are simplifications because ice layers may be significant obstacles to flow (Colbeck, 1973; Langham, 1974),

but they are reasonably correct for permafrost areas where mid-winter thaws are uncommon. The speed of flow within the saturated zone is generally two orders of magnitude greater than that within the unsaturated zone, so that with the exception of a very thin pack, the lag between surface melt and outflow is chiefly determined by the time taken for vertical movement (Colbeck, 1974a).

In the unsaturated zone within the snow, the rate of vertical movement of the meltwater wave depends not only on the initial characteristics of the pack, but also on the melt flux. This makes prediction complicated since early morning flux rates are usually low, and therefore are slow-moving. They can be overtaken by afternoon fluxes which move faster, producing a "shock front" (Colbeck, 1971) which overrides the slower-moving fluxes and is in turn overridden by larger, faster fluxes.

The input of water to the saturated layer at the base of the pack can be assessed by use of the continuity equation (Colbeck, 1974b). The thickness of the saturated layer varies both with distance downslope, and time, in response to additions of water from the unsaturated zone. Inputs arrive in waves, simultaneously if the surface melt rate and the thickness of the pack are constants for the whole slope. A wave is then generated within the saturated zone and transmitted downslope. The discharge at the base of the hillslope can be derived by integrating the inputs from the unsaturated layer to the saturated zone, over a period equivalent to the time taken for the wave to travel through the saturated layer. The flux rate and the snowmelt hydrograph can be predicted by dividing the discharge by the slope area.

One of the few studies to examine the snowmelt inputs to the slope hydrologic system using Colbeck's equations, was undertaken at Schefferville, Quebec (Dunne et al., 1976; Price, 1977). Predicted and measured hydrographs showed reasonable correspondence, indicating that if the assumptions of the model are met (particularly snow homogeneity, constant snow thickness and slope angle), quite accurate predictions can be made.

(d) Summer precipitation

(Subsystem C, Figure 35)

Rainfall occurring during the summer period in the Canadian Arctic Archipelago, has been thought to constitute as much as 50% of the total annual precipitation (Thompson, 1967). This figure may be an overestimate, however, because of the under-measurement of snow accumulation (see above, p. 70). For example, at Truelove Lowland, Devon Island, rainfall formed only 26% of the total precipitation over a three year period (Ryden, 1977), and at Resolute Bay, Cornwallis Island, it constituted 15 - 20% of the total during two years of study (Marsh and Woo, 1979).

In addition to its relatively low overall hydrological significance, the importance of rainfall to slopewash is further reduced as a result of its distribution throughout the summer. The rapidity of snowmelt in the continuous permafrost zone means that approximately 70-80% of the total annual precipitation is released in 7-10 days, while the remainder falls over a number of months. Moreover, summer rainfall may only just balance evapotranspiration losses (e.g. Marsh and Woo, 1977), or may be greatly exceeded by them (e.g. Woo, 1976a) further reducing its significance to slopewash.

Most rain falls as light showers in the high Arctic, with typical intensities of less than  $0.5 \text{ mm h}^{-1}$ . Usually these rates are too low to produce surface flow, notwithstanding the effect of permafrost which may permit the generation of surface flow at rainfall intensities lower than those necessary in nonpermafrost regions (Newbury, 1974). The perennial seeps found in valley bottoms in temperate catchments and in the discontinuous zone (Dingman, 1971) which permit saturated overland flow to develop after long periods of low intensity precipitation, are rare in the high Arctic because the active layer generally drains rapidly immediately following the cessation of snowmelt. As a result of these factors, it is likely that surface flow is generated by rainfall only if it occurs during or immediately following the ablation period. Under these circumstances, locations downslope from long-lasting snowbanks are the probable source areas for runoff. Dry areas possessing thick active layers (e.g. sandy interfluves) are unlikely to generate flow.

The preceding assessment of the importance of summer rainfall to slopewash applies to the majority of years, but not to those containing rare, extreme events. The variability of Arctic precipitation has been widely noted (e.g. Church, 1972; Pissart, 1967; Rudberg, 1963) and although meteorological records are of relatively short duration, storms greater than 50 mm in 24 hours have been recorded at Pond Inlet, Baffin Island (58.4 mm (Atmospheric Environment Service, 1971)) and at Vendom Fiord, Ellesmere Island (54.6 mm (Cogley and McCann, 1976)). During the latter storm, the maximum one-hour precipitation was 6.5 mm. An intensity of this magnitude is sufficient to exceed the infiltration capacities of some fine-grained materials (typically  $2.5 - 5.0 \text{ mm h}^{-1}$  (Gregory and Walling, 1973, p. 284)) and could potentially cause widespread surface flow.

The magnitude of high recurrence interval storms at Sachs Harbour appears to be less than in other areas of the Arctic Archipelago (see Figure 4), so that it is not certain that even a rare event would have significant geomorphic and hydrologic effects in the study area at Thomsen River. It should also be noted that while a 40 mm storm at Longyear dalen, Spitsbergen in 1972, produced widespread wash which was concentrated in gullies, its geomorphic effects only represented modification of existing forms produced by annual snowmelt (Jahn, 1975; 1976).

(e) Subaerial interchange

(Subsystem D, Figure 35)

Depending upon vapour pressure gradients, net moisture movement between ground and snow surfaces and the atmosphere can be in either direction. For the most part, however, water is lost by evaporation from the snow and by evapotranspiration from the ground. The hydrological effect of evaporative loss from the snow is to reduce the amount of water available for the recharge of the active layer and for runoff processes.

Evapotranspiration takes place throughout the summer months, diminishing the importance of the summer rainfall and creating soil moisture deficits.

Evaporation from snow is relatively difficult to measure accurately and usually it has been calculated with an energy balance approach. This method is valid prior to melt, but it may underestimate values when the snow cover is discontinuous. The proportion of the snowpack which is lost by evaporation is estimated to be 2% for Barrow, Alaska (Weller and Holmgren, 1974), with pre-melt rates of  $0.2 \text{ mm d}^{-1}$ . Similar calculations for Truelove Lowland, Devon Island, give lower rates of  $0.04 \text{ mm d}^{-1}$  for March and  $0.16 \text{ mm d}^{-1}$  for April. However, a larger value of  $0.35 \text{ mm d}^{-1}$  in May (Ryden, 1977) results in an average

total evaporation of 20 mm prior to melt, representing 15% of the snowpack water equivalent. On Cornwallis Island, Heron and Woo (1978) estimate that 20% of the energy available for melt is used for evaporation from snow, which may equal only 3% of the water equivalent of a snowbank.

Evaporation from snow usually is not an important part of the water losses during the melt period and has been ignored in some studies (e.g. Marsh and Woo, 1979). Spatially, however, losses may be significant in areas where the winter snow accumulation is only a few millimetres. Evaporation from these localities, caused particularly by radiation penetration of the snow and warming from below, may be sufficient to reduce to zero the amount of water available for infiltration or runoff.

The importance of evapotranspiration to permafrost slope hydrology lies in its magnitude relative to the total water balance. Evapotranspiration rates are surprisingly high in Arctic locations north of the tree-line (Table 5). In some cases, total evapotranspiration approximately equals summer precipitation (e.g. Marsh and Woo, 1977), and in others, greatly exceeds it (150 mm compared with 13 mm (Woo, 1976a)). In this latter case, evapotranspiration causes drying of the active layer after the end of snowmelt and a rapid cessation of streamflow. In terms of the total water balance, estimates of evapotranspiration losses range from 13 - 23% (Marsh and Woo, 1979) to 44% (Ryden, 1977).

Evapotranspiration can be determined with lysimeters and other water balance approaches, by energy budget and aerodynamic approaches, and by combination formulae. For example, lysimeters were used by Addison (1977) to separate the influences of transpiration and evaporation at micro-sites in the Truelove Lowland, Devon Island. The energy budget approach has been used in Alaska (Weller and Holmgren, 1974) in conjunction with evaporation pan measurements. In this method, evapotranspiration rates are

Author	Location	Rate of Evapotranspiration	Total
Gray et al. (1974)	Tuktoyaktuk, N.W.T. (Evaporation pan on open tundra)	June: 1.8-7.5 mm d <sup>-1</sup> average: 4.5 mm d <sup>-1</sup> July: 0.9-8.7 mm d <sup>-1</sup> average: 4.1 mm d <sup>-1</sup>	Not given
Marsh and Woo (1977)	Small internal basin, Vendom Fiord, Ellesmere Island, N.W.T.	0.45-1.00 mm d <sup>-1</sup>	27.2 mm (6 weeks)
Marsh and Woo (1979)	4 small basins, Resolute Bay, Cornwallis Island, N.W.T.	Not given	Average of 40 mm for 2 years
Ryden (1977) using figures from Addison (1977)	Truelove Lowland, Devon Island, N.W.T.	Raised beaches: 0.63-1.24 mm d <sup>-1</sup> Sedge meadows: 3.20-3.39 mm d <sup>-1</sup>	75 mm 185 mm Average of 90 mm for 2 years
Weller and Holmgren (1974)	Wet tundra, Barrow, Alaska	Post melt } maximum: 6.0 mm d <sup>-1</sup> period. } average: 4.5 mm d <sup>-1</sup> Mid-summer average: 2.6 mm d <sup>-1</sup> Freeze-up average: 1.1 mm d <sup>-1</sup>	Not given
Woo (1976a)	Alluvial terraces, Vendom Fiord, Ellesmere Island, N.W.T.	1.42 - 4.45 mm d <sup>-1</sup>	150 mm (8 weeks)

TABLE 5: Some evapotranspiration rates measured in permafrost areas

calculated from the residual in the energy balance ( $H_e$ ) divided by the latent heat of vaporization of water. The two measurement techniques corresponded satisfactorily. An attempt to use the aerodynamic approach near Tuktoyaktuk, N.W.T., however, (Gray et al., 1974) was unsuccessful because well-developed temperature and humidity profiles did not exist. The absence of profiles was attributed to undulating topography; local water bodies, and the existence of a layer of stagnant air equal to the height of the vegetation, which reduced vapour transfer. Since this combination of terrain characteristics is common in tundra areas, it is likely that the same difficulties would be experienced in other localities.

One of the more successful methods of determining actual evapotranspiration rates in high latitudes is the combination model, which incorporates both the aerodynamic and energy balance approaches (Wilson, 1971). In this model, the evapotranspiration rate is a function of the available radiant energy, the ground heat flux, and the air temperature. It may be used when the surface and overlying air adjust to one another and produce conditions known as "equilibrium" evapotranspiration (Slatyer and McIlroy, 1961). In order to extend the concept of equilibrium evapotranspiration beyond middle values of moisture, however, a coefficient must be introduced. Its value is 1.26 for saturated surfaces (Davies and Allen, 1973; Priestley and Taylor, 1972) including wet tundra meadows (Stewart and Rouse, 1976a; 1976b), and decreases as a function of soil moisture content to an average of 0.95 for non-transpiring upland tundra surfaces (Rouse and Stewart, 1972). For polar deserts and lithosols, it may be as low as 0.10 during dry periods (Marsh and Woo, 1979). The combination model has proved particularly useful in high latitudes because of its robustness (Rouse et al., 1977), but with the exception of the aerodynamic approach, all others outlined above appear suitable for use in permafrost areas.

(f) Surface and subsurface water movement(Subsystem E, Figure 35)

In contrast to the other hydrologic subsystems, little attention has been paid to the movement of surface and subsurface water on slopes in permafrost areas. Theory is poorly developed, partly because ideas concerning stream runoff production in nonpermafrost areas have not been rigorously applied to catchments in the continuous permafrost zone. Further, until recently, relatively few quantitative measurements existed, either of slope or stream discharge. Church (1974), for example, provides discharge data for less than 10 rivers in the continuous zone.

It is convenient to describe water movement on slopes in two ways: (1) in terms of total slope discharge, and (2) as a system divisible by infiltration, which determines the proportions of flow that take place on and beneath the ground surface.

In most areas of continuous permafrost, deep groundwater inputs can be assumed to be zero. As a result, streams are fed only by direct channel precipitation and by surface and subsurface flow from slopes. Streamflow in high Arctic areas usually is the major hydrologic output, so surface and subsurface flow volumes can be inferred to be very significant.

Their magnitude can be assessed by:

$$q_s = \frac{Q}{2L} \quad \dots(5)$$

where:  $q_s$  is the annual average unit slope discharge (surface and subsurface flow) at the base of a slope ( $m^3 m^{-1} a^{-1}$ );  
 $Q$  is the annual stream discharge ( $m^3 a^{-1}$ );  
 $2L$  is the average channel length (m).

Use of this equation is possible if annual stream discharge and average drainage density are known for a drainage basin. It will not be completely accurate, however, since it omits consideration of direct channel precipitation (either as snow accumulation or rainfall), supra-nival flow on slopes, and subsurface flow in the thawed layer beneath the stream bed (e.g. Church, 1972). At least two published studies of streamflow in the continuous permafrost zone supply sufficient information to calculate  $q_s$ . At d'Iberville Fiord, Ellesmere Island,  $q_s$  in 1973 was  $63.1 \text{ m}^3 \text{ m}^{-1}$  (see Ambler (1974) and Church (1974)), while on Bathurst Island in 1976 it varied between  $23.9$  and  $33.6 \text{ m}^3 \text{ m}^{-1}$  for a sub-basin and a main basin, respectively (see Wedel et al. (1978)). All three figures imply that only small sediment concentrations are necessary in slopewash to remove relatively large amounts from slopes.

An alternative method of calculating  $q_s$  for a particular slope is to evaluate the upslope water balance. If the overall moisture content of the active layer is assumed to be the same at thaw and freezeup:

$$q_s = M + R - (E + E_+) \quad \dots(6)$$

where:  $q_s$  is the annual unit slope discharge;  
 $M$  is the upslope snow volume (water equivalent);  
 $R$  is the upslope summer precipitation;  
 $E$  is the upslope snow evaporation;  
 $E_+$  is the upslope evapotranspiration.

All terms are in  $\text{m}^3 \text{ m}^{-1} \text{ a}^{-1}$ .

Equation (6) is valid for any point on a slope if upslope inputs and outputs are calculated, but it does not describe the spatial variation from slope base to ridge crest. A conceptual model, however, can be generated to consider this variation (Figure 39).

In the simplified case of a catchment underlain by permafrost receiving only spatially uniform rainfall inputs,  $q_s$  increases from zero at the watershed edge, to a maximum at the stream bank (Figure 39A). If infiltration capacities are exceeded by rainfall intensities, all of this increase will be absorbed by surface flow (as in Horton's model (1945)). The linear augmentation of  $q_s$  occurs because each point on the slope receives not only its own rainfall, but the sum of all upslope inputs.

The conceptual model behaves differently for a more realistic, irregular, slope where snow is the principal precipitation input. Snow is distributed unevenly on the slope, generally with a thin cover close to the crest and deep snowbanks on the valley floors and/or valley-side segments (Figure 39B). In this case, overall increases in  $q_s$  between crest and streambank do occur, but they are not related linearly to distance from slope crest. Snowbanks reduce  $q_s$  values by delaying the penetration of seasonal thaw, and thus diverting any flow from upslope areas into the snow (Ballantyne, 1978; Wilkinson and Bunting, 1975). Immediately downslope of snowbanks, there is a dramatic increase in  $q_s$  as these areas receive both upslope contributions and water from melting snow. Further downslope,  $q_s$  varies in relation to the thickness of the snow cover present at the beginning of the season. If slopes are short, it is probable that  $q_s$  increases to a maximum value at the stream bank.

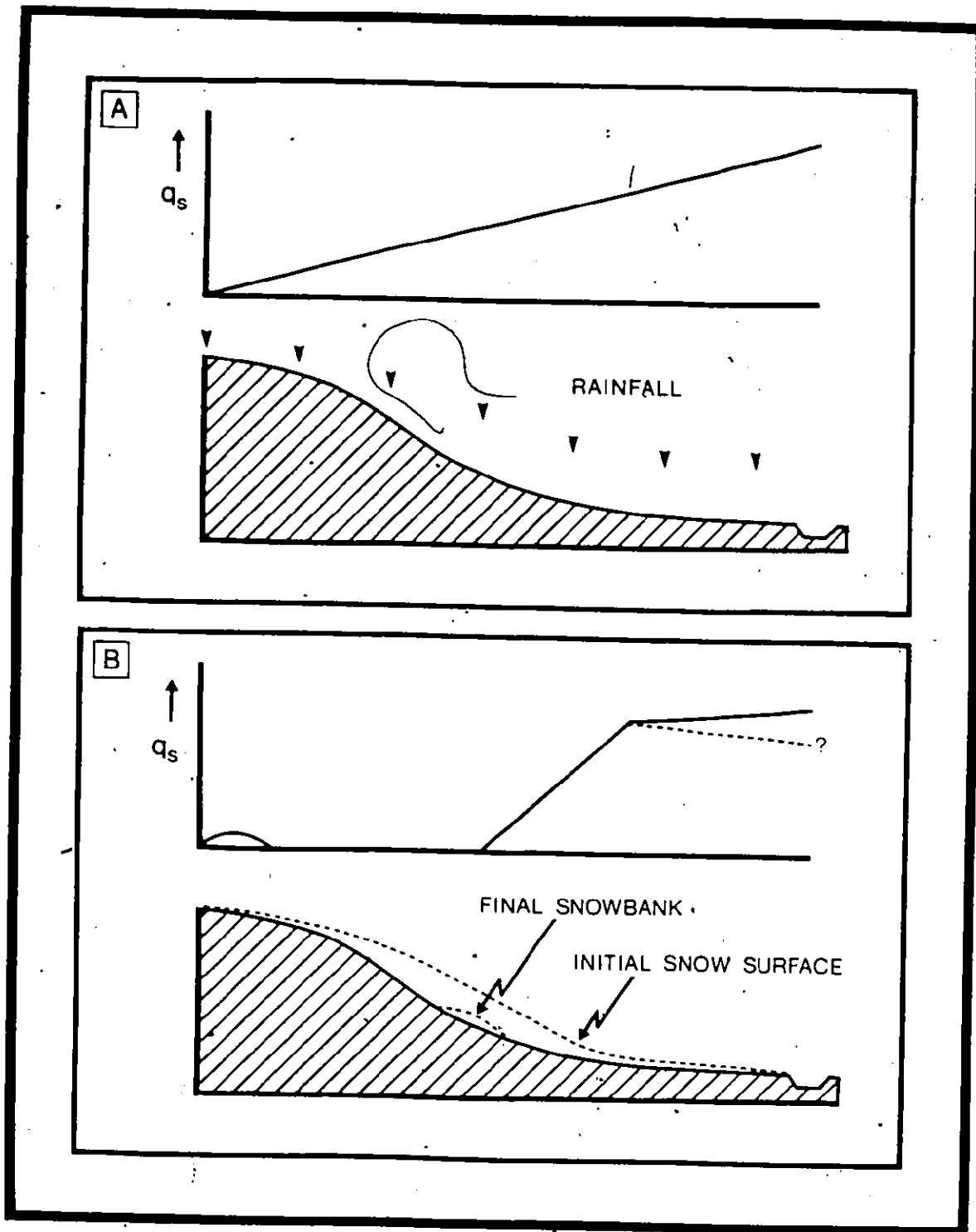


FIGURE 39 : CONCEPTUAL MODELS OF THE DOWNSLOPE VARIATION OF UNIT SLOPE DISCHARGE

A - Permafrost area,  $q_s$  produced by rainfall;

B - Permafrost area,  $q_s$  produced by snowmelt.

It is relatively easy to conceptualise the spatial distribution of combined surface and subsurface discharge. If each is treated as a separate entity, however, and water supplies are constant, an increase in one involves a decrease in the other and vice versa. The division of flow is dependent on the interrelationships of infiltration rate, antecedent moisture conditions, water supply and thawed layer thickness. For example, in one study in the discontinuous permafrost zone, near Yellowknife, N.W.T., eight runoff plots were monitored during the snowmelt period (Landals and Gill, 1973). Surface runoff coefficients of the plots, whose substrates included bedrock, vegetated unconsolidated deposits, and muskeg, ranged from 1.0% to 81.7%. The two major variables controlling the surface runoff coefficients were identified as the aspect of the plots, in relation to snow distribution and energy inputs, and the substrate material. Included in the latter was the effect of the degree of soil saturation prior to the previous year's freezeup.

In nonpermafrost areas, the theory of partial area contribution to stream runoff describes the spatial variation of surface and subsurface slope discharge. Storm runoff in streams is ascribed to direct channel precipitation, to rain falling on small, perennially saturated areas of the catchment generally contiguous with the channel itself, and in some cases to rapid subsurface flow. The runoff source areas are kept at or near saturation by a high water-table. Rain, while not able to exceed infiltration capacities in the strict Hortonian sense, cannot infiltrate because the ground is already saturated. The same mechanism has been

hypothesised for a subarctic catchment in the discontinuous permafrost zone (Dingman, 1971). The possibility exists, however, that runoff generation does not operate in the same way in the continuous permafrost zone. Because of the rapid depletion of active layer moisture following the cessation of snowmelt, the conditions necessary for the formation of perennial seeps may be rare. At present, evidence to confirm or deny this hypothesis is not extant (Dingman, 1975). Even if partial area contribution does operate in a catchment underlain by permafrost, it is likely that the source areas vary from the period of initial snowmelt through to freezeup.

At the start of snowmelt, infiltration is prevented by the presence of frozen ground. The process by which frozen ground is thawed by meltwater has been investigated in nonpermafrost areas. In these environments, infiltration can be described as four successive stages (Alexeev et al., 1972): (1) infiltration rates decrease as percolating water freezes in the pore spaces; (2) infiltration is zero as the soil becomes saturated above the blocking layer; (3) infiltration increases as the blocking layer is gradually destroyed; (4) steady infiltration occurs with no obstacles to flow. Clearly, stage (4) does not exist in the permafrost zone, but it can be hypothesised that stages (1) - (3) recur cyclically as the ground thaws by refreezing of meltwater at depth, and through the absorption of radiative and sensible heat fluxes.

In general terms, surface flow is promoted in areas which have just become snow-free since the thawed layer is thin or nonexistent and the

underlying ground is impermeable following stage (1). In areas possessing relatively thick active layers, surface flow is reduced. Therefore, two distinct slope hydrologic regimes can be recognised (Woo, 1976b). First, on slopes with large snowbanks, the water-table in the active layer remains high, and surface detention and flow are promoted in the area downslope from the snow. Second, on slopes without a snowbank or on segments upslope of one, the water-table within the active layer declines in response to evapotranspiration and subsurface flow. Clearly, if rain occurs, it is more likely to generate surface flow in those areas at or near saturation, i.e. those with high water-tables. The partial or variable source area concept, therefore, appears to be viable for permafrost areas. However, it is not related to stream-bank seeps, but rather to the topographic control of snowbank distribution. Only if a rainstorm occurs when large areas of the catchment are frozen (before widespread snowmelt) or in extreme storm events, is genuine Hortonian runoff generation likely to occur.

The distinction between surface and subsurface flow is significant in terms of slope hydrology because of the differences in controlling variables and flow velocities. These factors also are significant geomorphologically because of their effects on the erosivity of flow. In temperate areas, surface flow velocities commonly exceed subsurface flow speeds by three orders of magnitude (Kirkby, 1969). The only reported velocities of both flow types in a permafrost area are for Devon island where interflow velocities were measured as  $0.006 - 0.055 \text{ m s}^{-1}$  and surface flow in rills as  $0.40 - 0.60 \text{ m s}^{-1}$  (Wilkinson and Bunting,

1975). The differences in this case are one to two orders of magnitude, probably because the interflow was through an extremely permeable matrix (a talus slope). In areas where a thick vegetation mat exists, even the high permeability of sphagnum moss retards flow considerably and results in delayed responses in streamflow and long recessions (Dingman, 1971).

Surface flow can be laminar, transitional or turbulent, and tranquil or rapid. The flow type that prevails depends on the relative magnitudes of viscous, inertial and gravity forces, and can be described quantitatively by Reynolds' and Froude numbers. Despite a number of experiments (e.g. Emmett, 1970; Horton et al., 1934), the critical values for the Reynolds' number have not been determined accurately for thin flows, and information concerning their values in permafrost areas is not available.

One study of surface flow in a nonpermafrost area suggests that rapid flow does not occur on natural slopes since the maximum Froude number recorded was 0.2 (Emmett, 1970). Carson and Kirkby (1972, p. 38), however, indicate that laminar rapid flow is quite common on smooth, steep slopes. Schumm (1956) describes a type of pulsating flow due to the formation and break-up of micro-dams, and this may account for Emmett's low Froude numbers, since average velocities would remain low while rapid flow still occurred.

Subsurface flow is a response to a gradient of hydraulic potential. Moisture within the soil matrix primarily possesses potential energy, made up of gravitational potential and pressure potential. Rate of

flow in a porous medium depends on the gradient of hydraulic potential and on the permeability of the medium (Darcy's Law). Permeability varies from zero when a soil is dry to a maximum at saturation. The relationship is nonlinear because it is affected not only by the proportion of pore spaces filled with water, but also by the sizes of the water-filled pores. Furthermore, the relationship between moisture content and permeability is hysteretic and depends partly on the previous history of wetting and drying (e.g. Atkinson, 1978).

Subsurface flow in the soil matrix is assumed to be laminar because the permeability of soil materials is low (e.g. for silt, between  $10^{-7}$  and  $10^{-4}$   $\text{m s}^{-1}$  (Gregory and Walling, 1973, p. 62)). Even in the case of sphagnum moss whose permeability is much greater and within the range of gravels ( $>10^{-3}$   $\text{m s}^{-1}$  (Dingman, 1971)), flow is laminar. Flow within frozen soil materials depends on the thickness of the unfrozen water films, and is several orders of magnitude slower (e.g.  $10^{-10}$  -  $10^{-9}$   $\text{m s}^{-1}$  (Hoekstra, 1966)).

### Sediment movement

#### (a) Suspended sediment in surface wash

##### (Figure 36)

Sediment concentrations in surface wash are determined by the forces acting on the ground surface and the resistance of the surface to those forces. Permafrost has no direct influence over sediment detachment and transportation in surface wash, but it does influence the frequency and efficacy of the processes through its effects on the slope hydrological system.

Sediment can be entrained in surface wash produced by (1) snowmelt, (2) rainfall, or (3) a combination of the two. In the high Arctic, snowmelt is the source of the majority of surface runoff (see above, pp. 34-35). In temperate and tropical zones, raindrop impact is regarded as the major agent of sediment detachment (Ekern, 1950; Ellison, 1945; Feodoroff, 1965; Gerlach, 1976; Hudson, 1971, Mutchler and Young, 1975) and its absence during most of the surface wash production in tundra and polar desert areas reduces the potential erosivity of the flow. When rainfall does occur, as Jenness (1952, p. 239) notes, its intensity usually is too low to increase sediment concentrations through rainsplash processes. Geomorphologically, the action of an extreme summer rainfall event (e.g. Jahn, 1976; Rüdberg, 1963) involving splash as well as wash, can be viewed as a modifier of pre-existing wash features produced by more frequent meltwater erosion. However, in view of the low frequency of such an event and its absence during the three years of the present study, the extensive literature on rainsplash is not discussed further.

In general, there is a paucity of work treating the concepts of erodibility and erosivity in response to surface flow without rainsplash. Moreover, simple empirically-derived equations relating overall sediment discharge to overland flow (e.g. Kirkby, 1978b, p. 338) cannot be utilised for meltwater flow because they include the erosive effects of rainsplash. Attempts have been made to separate the effects of wash and splash in some laboratory experiments (e.g. Bryan, 1969; 1974a; 1974b; 1976; 1979; Meyer and Monke, 1965; Moeyersons and De Ploey, 1976; Poesen and Savat, 1978; Savat, 1975; Savat and Poesen, 1977; Young and Wiersma, 1973), while in

others, wash alone has been examined (e.g. Bubenzer et al., 1966; Lyle and Smerdon, 1965; Savat, 1977). Certain generalisations concerning erodibility and erosivity in permafrost areas may be derived from these papers.

The concentration of sediment in snowmelt-derived surface wash (Figure 36) is a function of: (1) the erodibility of the surface in relation to wash (sensu stricto, the inherent susceptibility of the soil to erosion as influenced by its physical properties (Hudson, 1976)); (2) the resistance of the vegetation cover; (3) the erosivity of the surface flow; and (4) the occurrence, if any, of direct aeolian sediment inputs. These factors are considered below.

The erodibility of soil is not necessarily equal with respect to both splash and wash (Ellison, 1947; 1948). Erodiability is affected by the soil properties which influence infiltration rates and therefore surface flow discharges, and by those properties which affect rates of dispersion, splashing, abrasion and transportation in rainfall and runoff (Wischmeier and Smith, 1965). In relation to wash alone, only the resistance to dispersion and the shearing force of surface flow need be considered.

The establishment of an erodibility index in terms of a soil's physical characteristics has been the aim of many laboratory tests. Bryan (1968) has reviewed more than 30 indices, most of which use textural or chemical attributes, with either one or several parameters. In terms of surface wash, the most erodible grain size is said to be silt (0.002 - 0.05 mm) (Ellison, 1948), but Bryan (1974b) argues that this is an oversimplification because it places too much emphasis on textural separates (as do many indices (e.g. Wischmeier et al., 1971)) and not enough on

aggregates. Further, it has been shown that in some cases, particles larger than silt may be the most erodible (e.g. Poesen and Savat, 1978; Savat and Poesen, 1977).

Bryan's (1974a; 1974b; 1976) experiments indicate that the water-stability of aggregates is an important, but not straightforward, influence on erodibility. Aggregates affect infiltration rates if they disperse and fill pores, reducing infiltration and increasing depth of surface flow and shearing force. Further, aggregates are larger than separates of which they are composed, and in most flow types, require greater velocities to move them. Thus water-stability is negatively correlated with wash erosion. Aggregation is promoted by organic matter (Robinson and Page, 1950), fine grain sizes, and by resistance to slaking (Yoder, 1936) and frost action (Bryan, 1971b). Aggregate water-stability is not a constant, and the erodibility of soil may vary seasonally, particularly in response to frost action. This dynamic view of erodibility makes its effects particularly difficult to isolate in the field.

The additional resistance to surface wash erosion provided by vegetation is relatively obvious. Root mats tend to bind the soil together so that a greater shear stress is necessary to induce particle movement (e.g. Waldron, 1977; Weaver, 1937). Roots which die and decay promote infiltration, and transpiration by plants also diminishes surface flow volumes. Plant shoots tend to reduce flow velocity and consequently its erosivity. As a result, and also because a vegetation cover is effective in breaking the impact of raindrops, it is a crucial factor in Wischmeier and Smith's (1965) Universal Soil Loss Equation.

World vegetation communities can be ranked according to the usual resistance they offer to surface wash and rainsplash. According to Young (1972, p. 225), polar desert and tundra communities are respectively, the second and third least protective. Despite this proximity in the list, the difference in the augmentation of resistance to erosion achieved by these vegetation types is important. For example, in the Thomsen River area, there is an obvious difference in the erodibility (sensu largo) of interfluves, possessing low percentage covers of polar desert species, and tundra meadows, completely covered by mosses, lichens and vascular plants. This variation correlates with moisture supply (i.e. snow distribution) so that the surfaces subjected to the greatest unit discharges are also the best protected.

The erosivity of surface flow depends on forces acting on particles at its base or protruding into the flow. These forces are: (1) dynamic pressure, exerted by the speed of the current, (2) Bernoulli hydrodynamic lift, resulting from the difference in mean downstream velocity between the top and bottom of the particle, and (3) in turbulent flow, randomly directed forces provided by eddies which lift some particles into the flow (Carson and Kirkby, 1972, p. 38). The parameters determining the magnitude of these forces, and consequently the flow erosivity, are average flow velocity and velocity gradient, flow depth and slope, discharge and friction. These are not completely independent variables and their interaction prohibits easy determination of their quantitative influence on flow erosivity.

In the laboratory, it is possible to describe surface flow numerically in terms of Reynolds' and Froude numbers. This is a more difficult task in the field because surface flow is often unsteady and spatially concentrated ("subdivided" flow (Horton, 1945)), in response to particles which exceed its depth (Emmett, 1978) and topographic irregularities which need be no greater than 1 mm in size (Bryan, 1979). Flow may also be affected by surface tension forces (Young and Mutchler, 1969). Thin surface wash is far more complicated than channel flow for which most discharge equations (e.g. Manning, Chezy, and Darcy-Weisbach) and sediment transport equations (e.g. Bagnold, 1956; Meyer-Peter and Muller, 1948; Yalin, 1963) have been developed. Modifications of the flow formulae have been undertaken (e.g. Emmett, 1970; 1978; Savat, 1975; 1977) and the methods involved can be employed in the field to a certain extent. However, the accurate determination of particular parameters (e.g. flow depth) is difficult, and their application in vegetated areas questionable (Dunne, 1978).

Sediment transport equations are still harder to adapt to thin flow problems because the fundamental processes are inadequately understood. Slope angle can be used as an example of this problem. Intuitively, it seems likely that the erosivity of surface flow increases with slope angle, but the form of the relationship is not obvious. Most researchers have found that the relationship between soil loss (in this case, an analogue of flow erosivity) and slope angle is a power equation:

$$E = c.S^m$$

... (7)

where: E is the rate of soil loss;  
S is the slope (percent);  
c, m are constants.

Kirkby (1969), Musgrave (1947) and Zingg (1940) found the exponent m in equation (7) to be 1.4, but others suggest a value of 2.0 (e.g. Hudson and Jackson, 1959). Still other studies show no definite relationship between slope angle and soil loss (e.g. Lal, 1976; Leopold et al., 1966). Wischmeier and Smith (1965) suggest a quadratic form, and Bryan (1979) found that a polynomial equation best predicted the rates of soil loss.

It is clear that the establishment of a unifying and generally accepted surface wash equation is unlikely. It is rendered still more so by problems associated with flows that do not transport at capacity. The Yalin (1963) equation, for example, is a relatively simple model that can predict sediment transport within a factor of two in the field (Foster and Meyer, 1972), but only when wash is transporting at its capacity. In the case of snowmelt on slopes, the flow transport capacity is likely to exceed the amount of sediment in suspension because the absence of raindrop impacts reduces the probability of detachment. The problems associated with evaluating the erosivity of thin surface flows, together with those concerning soil erodibility, justify a field approach to the study of surface wash erosion, but render generalisations from specific measurements more difficult.

Direct aeolian input to suspended sediment in surface wash is dependent on site-specific conditions and there is little theory to discuss.

Input can occur during the ablation period directly into the surface flow, or can result from the melting of snow containing sediment deposited during the winter. In the latter case, some particles subsequently may be deposited onto the ground (Warren Wilson, 1958), so that a particular sediment concentration in the snow does not imply the same amount in derived meltwater wash. Within the permafrost zone, measured concentrations of sediment in snow range from 274 - 607 mg L<sup>-1</sup> on Bathurst Island (Wedel et al., 1978) to 2000 mg L<sup>-1</sup> on Spitsbergen (Czeppe, 1965). In the basal layers of snowpatches at Radstock Bay, Devon Island, sediment concentrations varied between 2000 mg L<sup>-1</sup> and 27000 mg L<sup>-1</sup> (Wilkinson and Bunting, 1975), while up to 23 g m<sup>-2</sup> of material (largely organic) was collected from the snow surface at Truelove Lowland, Devon Island (Teeri and Barrett, 1975).

Despite the gaps inherent in slopewash theory, only two studies attempt to provide quantitative data on the efficacy and magnitude of surface wash erosion in permafrost areas.

A. Jahn measured surface wash transport during two summers on seven unvegetated slopes at Hornsund, southern Spitsbergen (Jahn, 1960; 1961). Sediment traps were used so sediment concentrations in the surface flow are not known. Two main conclusions were reached. First, although sediment removal was low on short slopes (average of 1 g m<sup>-2</sup> a<sup>-1</sup>), wash was capable of removing at least 18 g m<sup>-2</sup> a<sup>-1</sup> at a snowbank location. This figure corresponds to a surface lowering of approximately  $7 \times 10^{-3}$  mm a<sup>-1</sup>, which is lower than many others obtained for nonpermafrost areas (Young, 1974, p. 69). Second, most sediment was transported by meltwater, not rainfall. These data can be

contrasted with snowmelt-induced wash erosion measured near Sudbury, Ontario, which produced two orders of magnitude more sediment. However, this constituted only 17 % of the annual erosion (Pearce, 1976a; 1976b), indicating the importance of rainfall in temperate areas in promoting erosion.

T. Wilkinson measured the movement of sediment for two years in a series of rills on a low-angled slope at Radstock Bay, south-west Devon Island (Wilkinson and Bunting, 1975). As in Jahn's study, sediment was trapped, but direct comparisons between the two are not possible because erosion rates were not calculated in the Devon Island study. Sediment concentrations of up to  $1700 \text{ mg} \cdot \text{L}^{-1}$  were measured, but both these and flow velocities declined downslope. Some rills carried very low concentrations of sediment and appeared clear, while others were turbid. Average transport was  $65 \text{ g m}^{-1}$  slope width (30 kg across a slope 450 m wide).

The work of Jahn (1960; 1961) and Wilkinson and Bunting (1975) can be regarded as the first steps towards assessing the quantitative importance of surface wash as an erosional and transportational agent in a permafrost region.

(b) Dissolved solids in surface and subsurface wash

(Figure 37)

Dissolved solids are discussed in reference to wash as a whole rather than to its constituents, because the processes of chemical solution are the same for both types of flow.

Solution operates when water comes in contact with the products of weathering, either at the surface or within the regolith. A number of key variables govern the rates and type of chemical reactions that occur (see Figure 37): (1) the composition of the soil and the solubility of its constituents; (2) the concentrations of solutes within the water; (3) the temperature of the solvent; (4) the duration of the contact. These variables in turn are influenced by factors such as vegetation growth, water source, and flow routing.

The solution process is azonal and variations between permafrost and nonpermafrost areas are a matter of degree, rather than the result of the introduction of new factors. The existence of permafrost, for example, usually prevents deep percolation and long residence times, and therefore can be expected to reduce the probability of very high dissolved solids concentrations in subsurface water. Further, most rates of chemical reactions are reduced at low temperatures, so that if all other factors are equal, to achieve the same solute concentrations, residence times must be greater in permafrost regions than in lower latitudes. In areas where groundwater springs occur, however, concentrations of salts can be high (e.g. Anisimova, 1978; Pisarskiy, et al., 1978; van Everdingen, 1974).

Substantial differences exist in the chemical composition of high Arctic soil types (Tedrow, 1966; 1974) but their effects on solution have not been investigated. Polar desert soils commonly are alkaline, hummocky soils near-neutral, and tundra soils acidic, but there are exceptions depending on the composition of the underlying unweathered material. In some areas on the margins of the true high Arctic, such as central Banks

Island, a quasi-catena of soil types develops. Polar desert soils exist on interfluves, hummocky soils on slopes, leading through transitional soils of the polar desert-tundra interjacency to tundra soils in the valleys (French, 1971; Tedrow, 1974). These changes are linked to moisture supply and vegetation distribution and are likely to be significant to solution processes.

The distribution of wash on any slope is linked to the pattern of snow accumulation (see Figure 39B). On the upper convexity, unit discharges are low and solution transport processes correspondingly insignificant. Materials leached down the soil profile by snowmelt and rainfall, are precipitated as efflorescences during evaporation. The snowbank location itself, also is subjected to little solution activity. First, the concentration of dissolved solids in snow is low (e.g. total hardness was measured as 12 - 13 mg L<sup>-1</sup> as CaCO<sub>3</sub> by McCann and Cogley (1971) and less than 5 mg L<sup>-1</sup> by Thorn (1974, p. 159)). These data indicate that snow does not react strongly either with the substrate or with sediment trapped within it. Second, unit discharges are low along the snowbed, because snowmelt and upslope subsurface flow contributions travel within the snow itself and not along the ground surface (see above, p. 64).

The downslope margin of the snowbank is affected by high unit discharges (see Figure 39B) and may be the area of maximum solution activity because of the presence of dissolved carbon dioxide (CO<sub>2</sub>) in the meltwater. Concentrations of CO<sub>2</sub> within a snowbank and in meltwater, however, are matters for discussion. Williams (1949) measured percentages in a snowbank which were greater than in the atmosphere, and

this increase was attributed to melting and refreezing of the snow. Further, the  $\text{CO}_2$  could be absorbed by the meltwater because the saturation concentration of  $\text{CO}_2$  is highest at  $0^\circ\text{C}$ . The hypothesis that meltwater is a very aggressive solutional agent was supported by Corbel (1959a; 1959b). Subsequent research, however, indicates that this conclusion may have been premature. First, measurement of the pH of snowbanks shows that it increases as snow ages, and meltwater consequently becomes less aggressive towards carbonates (Cogley, 1972; Thorn, 1974, p. 159). Second, the temperature of the reactions and the lack of biogenic  $\text{CO}_2$  production in many polar areas, appear to be just as significant as the effects of an increase in  $\text{CO}_2$  within the snow and/or higher saturation values (Smith, 1972; Woo and Marsh, 1977).

At the margin of a snowbank, when vegetation and soil are present, water takes material into solution either at the surface or from within the downslope-thickening active layer. As the flow stays in contact with the acidic tundra soil, the former becomes more acidic and more aggressive towards carbonates. Thus there is a rapid increase in solute concentrations away from the snowbank edge. For example, at one location, Cogley (1971) measured an increase in total hardness of  $60 \text{ mg L}^{-1}$  in a distance of 25 m.

The residence time and routing of flow ultimately determine whether solute saturations are reached before the slopewash arrives at the stream channel. Experiments indicate that in some cases equilibrium with soil is reached in a matter of hours (Bricker et al., 1968). The

predominance of rapid runoff during the melt period is indicated by the low solute concentrations in streamflow at this time (Woo and Marsh, 1977).

Following melt, slope unit discharges progressively decrease except after summer rainfall events, more water is routed as subsurface flow, and solute concentrations have a greater chance to reach saturation values. The routing factor is exemplified in the diurnal pattern of calcium and magnesium concentrations observed in rills on Devon Island (Wilkinson and Bunting, 1975). Concentrations were lower during the "daytime" hours when most flow was being produced by snowmelt, and higher at "night" when much of the flow was derived from subsurface flow through a talus slope.

The effects of summer rainfall on solutional activity are not definitely known, but the processes outlined above indicate that high concentrations of solutes could be expected in any rain-induced slopewash. This is particularly likely since rainfall often merely displaces water that has been in contact with the soil for some time and has reached equilibrium (Carson and Kirkby, 1972, p. 256). The relative importance of snowmelt and rainfall in solution transport is not known, but in view of the small volumes of rainfall-derived wash in normal years, it is likely that melt is the more important.

Few studies examine solution from the standpoint of it being a geomorphic process, or consider its effect on slope form (Carson and Kirkby, 1972, p. 238). Still fewer have been undertaken in the permafrost zone. The near-comprehensive, long-term process study of A. Rapp (1960) in the Karkevagge, northern Lappand, an area of discontinuous permafrost, assessed solution as the most active erosional process operating in the valley. The

calculated total of sediment removed, however, was based on relatively few concentration data and thus the absolute accuracy of the figure is open to question. Moreover, the value cannot be related to slopewash activity because it includes the effects of fluvial solution. Other studies of the solute load in Arctic rivers have shown that its importance varies during the summer, but is generally less than that of the suspended load (Chyurlia, 1978; Guilbault and Chacko, 1978; Wedel et al., 1978). Since these studies were short-term, however, more data are needed before these findings or those of Rapp can be accepted as a general rule for polar areas.

Available solute concentration data in slopewash in the continuous permafrost zone illustrate some of the points outlined above (Table 6). Particularly notable are downslope increases and the effects of biogenic activity in increasing concentrations in vegetated areas. A need remains, however, for information concerning the relative proportions of suspended and dissolved material in surface wash in a permafrost area, and more generally, for a comparison of the efficacy of surface and subsurface wash as erosional and transportational agents.

D

Author	Location	Material	Flow Type	Total Hardness (as mg L <sup>-1</sup> calcium carbonate)
McCann and Cogley (1971)	Devon Island, N.W.T.	Limestone (mainly unvegetated)	Rillflow  Talus interflow  Pool overlying vegetation	Increases downslope from snowbank edge: In 15m, from 26 to 64 In 25m, from 13 to 82  69 - 116  126 - 144
Smith (1972)	Somerset Island, N.W.T.	Limestone (mainly unvegetated)	Rill and sheet flow  Pool over- lying vegetation	Range: 18-95 Average: 60  Maximum: >130 Average: 106
Wilkinson and Bunting (1975)	Devon Island, N.W.T.	25-35% Calcareous rocks (unvegetated)	Rillflow  Talus interflow	Diurnal variation: 30-70  Range: 72-97
Woo and Marsh (1977)	Ellesmere Island, N.W.T.	Limestone and Outwash (vegetated)	Seepage  Soil water (in pit)	Average: 180 Maximum: 235  Average: 227 Maximum: 312

TABLE 6: Some total hardness values measured in surface and subsurface wash in the continuous permafrost zone

CHAPTER 4

FIELD TECHNIQUES

## Errors

In addition to the potential methodological problems of a process approach, there are errors of the precision and accuracy types (Schenck, 1961) inherent in all of the measurement techniques used in this study. Assessment of these inaccuracies is important because much of the analysis is comparative and errors may accumulate. Some check can be made of errors in water balance studies, but only if all factors are quantified. The risk of drawing incorrect conclusions is increased where residual estimates are employed.

The original aim was to keep accuracy within  $\pm 10\%$  for all measurements. In practice, a more realistic figure for certain data collected is  $\pm 20\%$  and in some cases errors may exceed even this estimate. Inaccuracies are attributable to a number of problems: (1) observer error (more probable in rigorous weather conditions); (2) equipment design (e.g. leaks beneath the surface flow collectors); (3) disturbance of the natural system (e.g. during snow ablation measurements); (4) limited time (e.g. between samples of continuously varying factors, such as evapotranspiration); (5) inadequate theory (e.g. the assumption that the snowpack contributory area is the same as that of the underlying terrain); (6) transformation of data (e.g. as in residual analysis). These difficulties are examined in more detail below, for each of the measurement techniques employed.

## Energy measurement techniques

During the three study seasons, the following factors relating to the energy balance were recorded, either at the runoff sites or at the base camp:

- (1) radiation;
- (2) ~~air~~ temperature;
- (3) wind speed and direction;
- (4) cloud cover, level and type;
- (5) barometric pressure.

Net all-wave radiation ( $H_r$  in equation (3)) was measured at the 1 m level during 1977 with a single Swissteco type S-1 net radiometer mounted on a tripod and linked to a Rustrack millivolt recorder. The radiometer was moved between the runoff sites during the ablation period in order to obtain comparative measurements relating to the influence of slope and aspect. During the 1978 field season, similar procedures were followed but two radiometers were used, the second measuring net short-wave radiation. In 1979, the radiometers were set up over the base camp terrace and remained over horizontal ground throughout the season.

Radiation proved to be one of the most difficult factors to measure accurately. Although changes in radiation balances were recorded on the Rustrack, the timing of the recorder had to be checked continually, and the values recorded were less than the true figures. Comparisons were made between the Rustrack and an accurate millivoltmeter and the results corrected. However, these problems meant that considerable transformation of the data was necessary before they could be used for analysis, and results are considered accurate to  $\pm 20\%$ .

Air temperature was monitored with a Weathermeasure hygrothermograph inside an instrument screen at a height of 1.75 m. During 1977, this apparatus was located at site 1, while for the two other years it was situated on the base camp terrace. The hygrothermograph was calibrated weekly against an accurate ( $\pm 0.05^\circ\text{C}$ ) thermometer, and the chart for

the week was corrected. Tests comparing chart-recorded maxima and minima and those obtained from accurate thermometers indicate that the corrected thermograph trace is accurate to  $\pm 0.7^{\circ}\text{C}$ .

Wind speeds were measured in 1978 and 1979 with a Cassela cup-type totalising anemometer. The instrument was mounted on a pole at the 1.75 m level, 75 m from all obstructions, and measured wind-run over 12-hour periods. In addition, spot readings (accurate to  $\pm 5\%$ ) of wind speeds were taken at the same intervals, and wind directions assessed using a flag and marked ground.

Cloud observations were recorded bi-daily during 1978 and 1979. Significant changes in conditions during the day were also noted. The accuracy of these observations depend on observer experience and error is difficult to estimate. Cloud types probably were identified correctly, cloud cover data likely are accurate to  $\pm 5\%$ , and cloud levels to  $\pm 10\%$  for low clouds and  $\pm 20\%$  for middle and high clouds.

Barometric pressure was measured at 12 hour intervals in 1979 only, with a Fuess aneroid barometer located at the base camp. Accuracy of this instrument, which was transported in an airtight container following calibration at Resolute Bay, was  $\pm 0.01$  kPa.

#### Water measurement techniques

The following factors relating to the slope water balance were measured during the study:

- (1) snow density, depth, ablation, areal change and temperature;
- (2) summer precipitation;
- (3) evaporation from snow;
- (4) evapotranspiration from the ground;

- (5) surface flow;
- (6) subsurface flow.

Snow measurements were made in a number of different ways. The data required concerned the ripening process and subsequent changes in the water equivalents of the snow at the runoff sites. In addition in 1979, successive surveys of snow-free areas over 7 km<sup>2</sup> of the study area were undertaken to examine the pattern of snow ablation (see Figure 15).

Snow density was measured during 1977 and 1978 with U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) type Q.5 L snow tubes, in conjunction with an Ohaus portable triple-beam balance accurate to  $\pm 0.1$  g. Densities were measured in snow-pits dug in the snow beside the runoff plots (Figure 40). The number of pits used varied with changing snow conditions and from one plot to another, but usually there were 4 or 5 at each site, located 10 - 30 m apart. Measurements within the pits were made wherever distinct changes in the snow crystallography were observed, or at depths of 0.05 m, 0.10 m, 0.20 m, and then at 0.2 m intervals down to the snowpack base. As a result of the time-consuming nature of this type of density measurement, particularly when the triple beam balance was affected by wind, densities generally were measured at intervals of 3 - 7 days throughout the ablation period.

During the 1979 field season, snow densities were measured with a 2 m Mount Rose snow sampler, in conjunction with a spring balance (tested accuracy of  $\pm 2\%$ ) and weighing cradle. This method allowed integration of snow densities over the complete snowpack thickness at sites 1 and 2, but only the top portions of the deep packs at sites 3 and 4 could be sampled. The Mount Rose sampler was easier to operate than the CRREL tubes



**FIGURE 40 : CRREL SNOW MEASUREMENT EQUIPMENT**  
Snow density tube and snow thermometer are visible in a shallow snow pit, site 2, June 5, 1977. Note low albedo of snow surface in contrast to clean snow at depth.

except during the ripening period, and measurements were made more frequently than with the snow-pit method.

Both the Mount Rose and the CRREL tubes were unable to sample the ice layer which developed at the base of all snowpacks. As a result, neither method provided completely accurate results. A greater problem, however, was the representativeness of the sampled snow since the CRREL tube cores sometimes contained air spaces. A comparison of the two methods (Table 7) indicates that: (1) neither method is completely consistent, but that the Mount Rose sampler gives results with a low degree of variability in deep packs; and (2) since the larger samples obtained with the Mount Rose kit are likely to be more accurate, the small tubes probably underestimate density by 10 - 20%. In view of this finding, density figures obtained using the CRREL tubes in 1977 and 1978 were increased by 15%. The overall accuracy of the corrected snow density data is thought to be  $\pm 10\%$ .

Snow ablation was measured, usually daily, with a calibrated metal probe in 1977 and 1978, and downwards from a fixed rope in 1979. A number of problems were experienced with both methods.

Use of a metal rod to measure changes in snow depths at fixed points on a snow course inevitably results in disturbance of the snow surface, with probable concomitant changes in snow albedo. Further, the task is fairly arduous, particularly when ice layers are encountered by the probe. Inaccuracies were introduced by the basal ice layer in the snowbank, which prevented measurement down to the ground surface. Normal errors were of the precision type (equal to  $\pm 20$  mm), and while these were relatively unimportant (<10%), on days of rapid ablation, they could constitute significant percentages on days of low melt.

Location	MEASURED SNOW DENSITY ( $\text{Mg m}^{-3}$ )		Difference between methods
	CRREL tube	Mount Rose sampler	
Site 3	Surface: 0.252;0.306 0.2 m: 0.358;0.342 0.4 m: 0.396;0.376 0.6 m: 0.366;0.386 0.8 m: 0.391;0.364 1.0 m: 0.366;0.388 1.2 m: 0.308;0.288	(integrated over 1.2m) 0.426 0.427 0.420	
	Average: 0.349	Average: 0.424	18%
	Site 1	Surface: 0.396;0.382 0.2 m: 0.322;0.352 0.4 m: 0.302;0.308	(integrated over 0.5m) 0.400 0.385 0.369
	Average: 0.343	Average: 0.385	11%

TABLE 7: Comparison of snow density measurements made with a Mount Rose snow sampling kit and a CRREL snow tube in conjunction with a triple-beam balance, June 8, 1979.

As a result of the problems experienced with the probing method, an alternative was employed in 1979. Poles, 2 m in length, were inserted down to the ground surface, and held upright by ropes tied to pegs embedded in the snow or the ground beneath (Figures 41 and 42). A third pole, secured in between the other two, acted as a benchmark. The rope slung between the two outer poles was tightened to the level of the benchmark for the purpose of ablation readings, and the distances between fixed points on the rope and the snow surface measured. The overall accuracy of this method was checked by comparing total measured ablation and initial snow depths (Table 8). On average, rope measurements underestimated true ablation by less than 10%, with absolute differences after 29 measurements averaging only 0.07 m.

The areal change of the snowbank is an important factor in water equivalent change, so the surface flow collector (see p. 125) was positioned so that it was possible to determine the topographic area of the slope which it drained. The edges of the snowbank within this area were marked, usually daily, with colour-coded dowels. These were surveyed at the end of the field season with a Wild NK2 level and stadia rod, to a spatial accuracy of  $\pm 0.10$  m and an elevation accuracy of  $\pm 1$  mm. Maps of the plots (e.g. Figure 43) were constructed from these surveys, and areal changes subsequently were measured with a polar planimeter.

The major error in assessing changes in the snowpack water equivalent is the assumption that the snowmelt contributory area is the same as the underlying slope drainage area. This error depends on the form of ice layers in the snow, on the shape of the basal ice layer, and on the thickness of the saturated zone within the snow, so that it need not be constant, even during a single day. It is hypothesised that in some cases

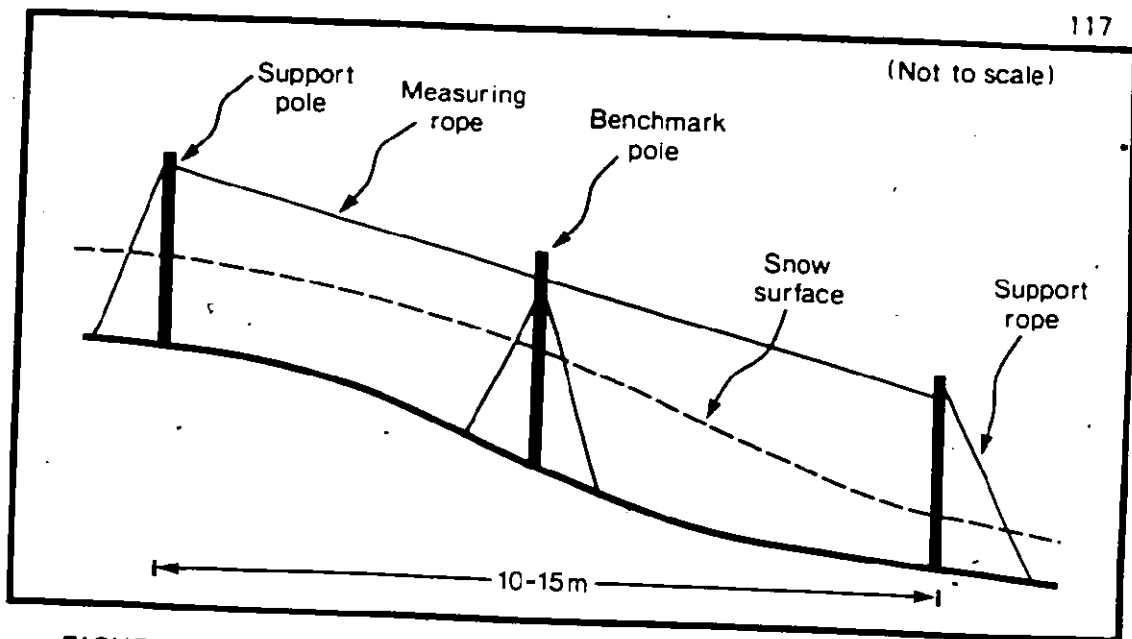


FIGURE 41 : APPARATUS USED TO MEASURE SNOW ABLATION, 1979



FIGURE 42 : ABLATION MEASURING DEVICE, SITE 4, JUNE 1, 1979  
 Note absence of disturbance to snow surface immediately below measuring rope.

Position on rope (m)	Thickness of snow measured 1/6/79 (m)	Total ablation measured to 9/7/79 (29 measurements) (m)	Difference (m)	% error
7	1.19	1.09	-0.10	- 8.4
8	1.06	0.98	-0.08	- 7.5
9	0.98	0.83	-0.15	-15.3
10	0.87	0.82	-0.05	- 5.7
11	1.01	0.96	-0.05	- 5.0
12	0.94	0.90	-0.04	- 4.3
13	0.81	0.76	-0.05	- 6.2
14	0.71	0.67	-0.04	- 5.6
Average			-0.07	- 7.3
Standard deviation			0.04	3.5

TABLE 8: Comparison of successive measurements made with an ablation rope, and initial snow depths measured with a calibrated probe, upper rope, site 3, 1979

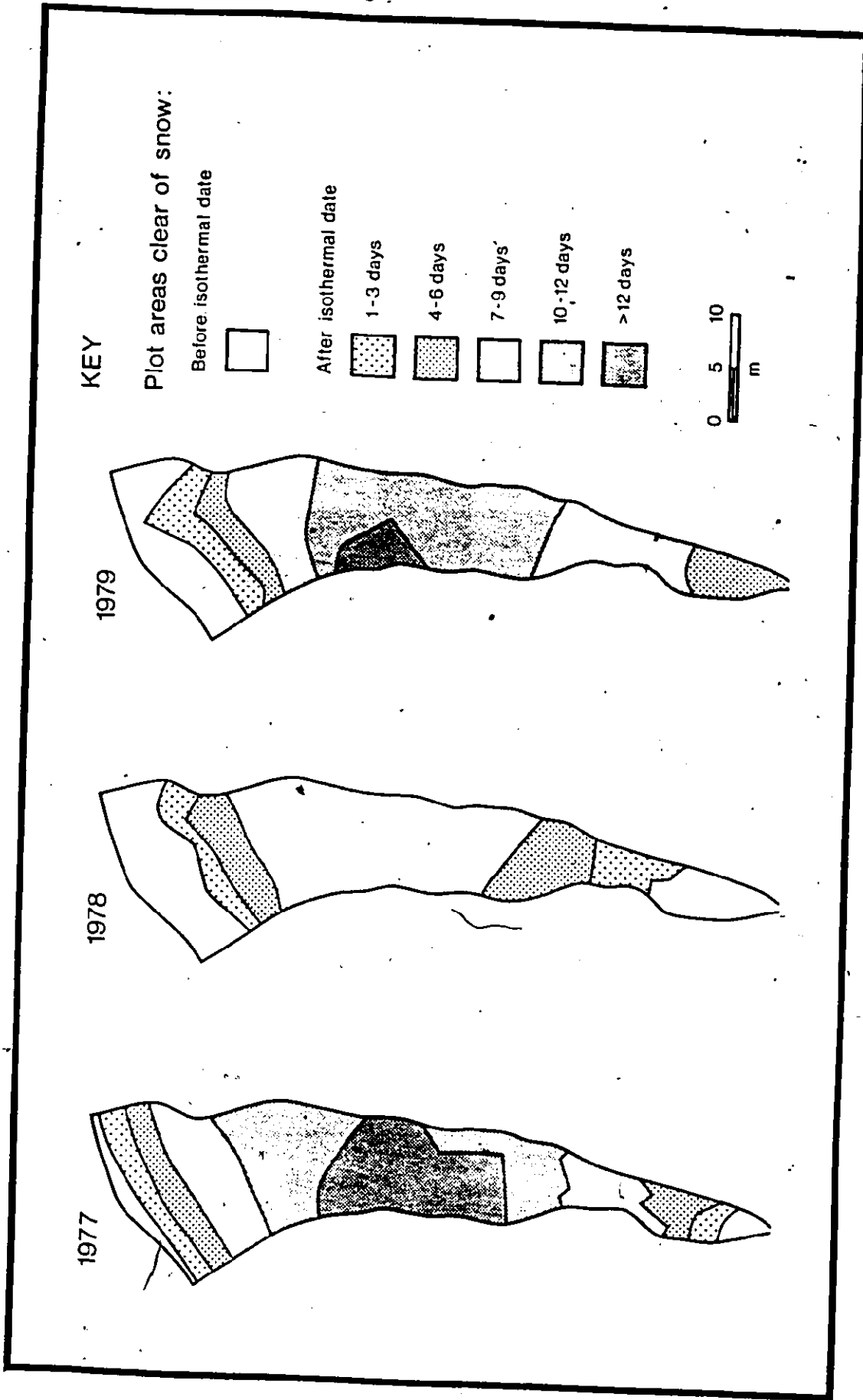


FIGURE 43 : MAPS OF PLOT 2 SHOWING CHANGES IN SNOW DISTRIBUTION DURING THREE SUCCESSIVE YEARS

this error may cause accuracy to decrease to  $\pm 50\%$  and outweighs all other technical problems. This point is discussed further in Chapter 5 (pp. 180-181).

Snow temperature profiles were monitored in 1977 and 1978 with CRREL snow thermometers (see Figure 40) accurate to  $\pm 0.1^\circ\text{C}$ , and in 1979 with a YSI tele-thermometer and thermistor probe (accurate to  $\pm 0.5^\circ\text{C}$ ) (Figure 44). The latter was easier to use since it did not require the creation of snow pits except in areas where the snow was more than 1 m thick. The temperature data obtained, were useful adjuncts to observations on the ripening process, and revealed when the snowpacks became isothermal at  $0^\circ\text{C}$ .

Precipitation was recorded with a number of raingauges during the three field seasons. A Weathermeasure automatic tipping-bucket raingauge (0.20 m diameter orifice) and recorder were stationed at site 1 throughout the whole period. This apparatus operated with increments of 0.25 mm of precipitation, so the maximum possible error is just less than this figure. Up to eight nonrecording gauges (0.15 m diameter standard and Taylor Clear-View types) were deployed at the study sites and at the base camp, and readings of these were completed within an hour of the cessation of rainfall. The usual problems of evaporation and gauge undercatch (Rodda, 1969) apply to these rainfall data, but results may also reflect spatial variation in rainfall amounts.

The method used to determine evaporation from snow in 1977 apparently incorporated substantial error. Two two-level, white polyurethane containers, 200 mm in diameter, and 150 mm deep, were filled with snow, weighed and embedded in the snowbank at site 2 (Figure 45A). Reweighing the following day allowed evaluation of the balance between

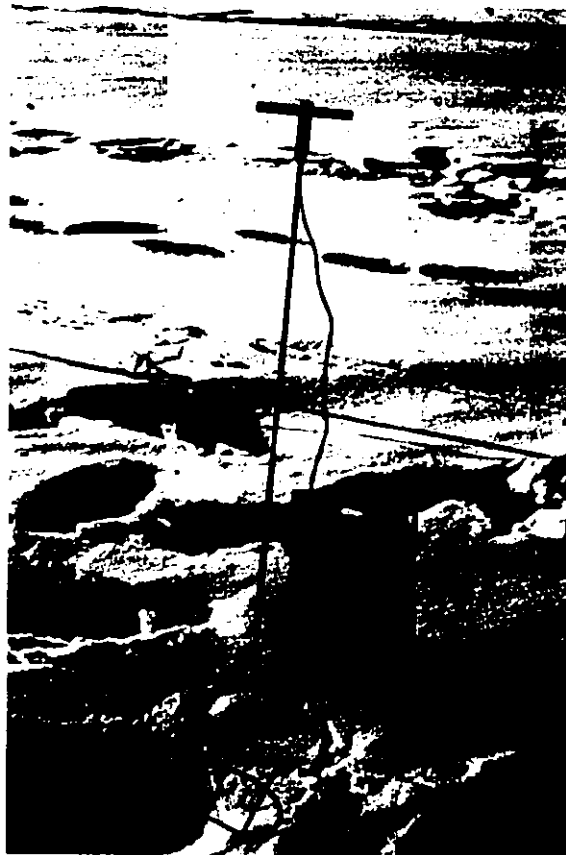


FIGURE 44 : YSI TELE-THERMOMETER

This device enabled measurement of snow temperatures to depths of 0.8 m without disturbance of the snowpack.

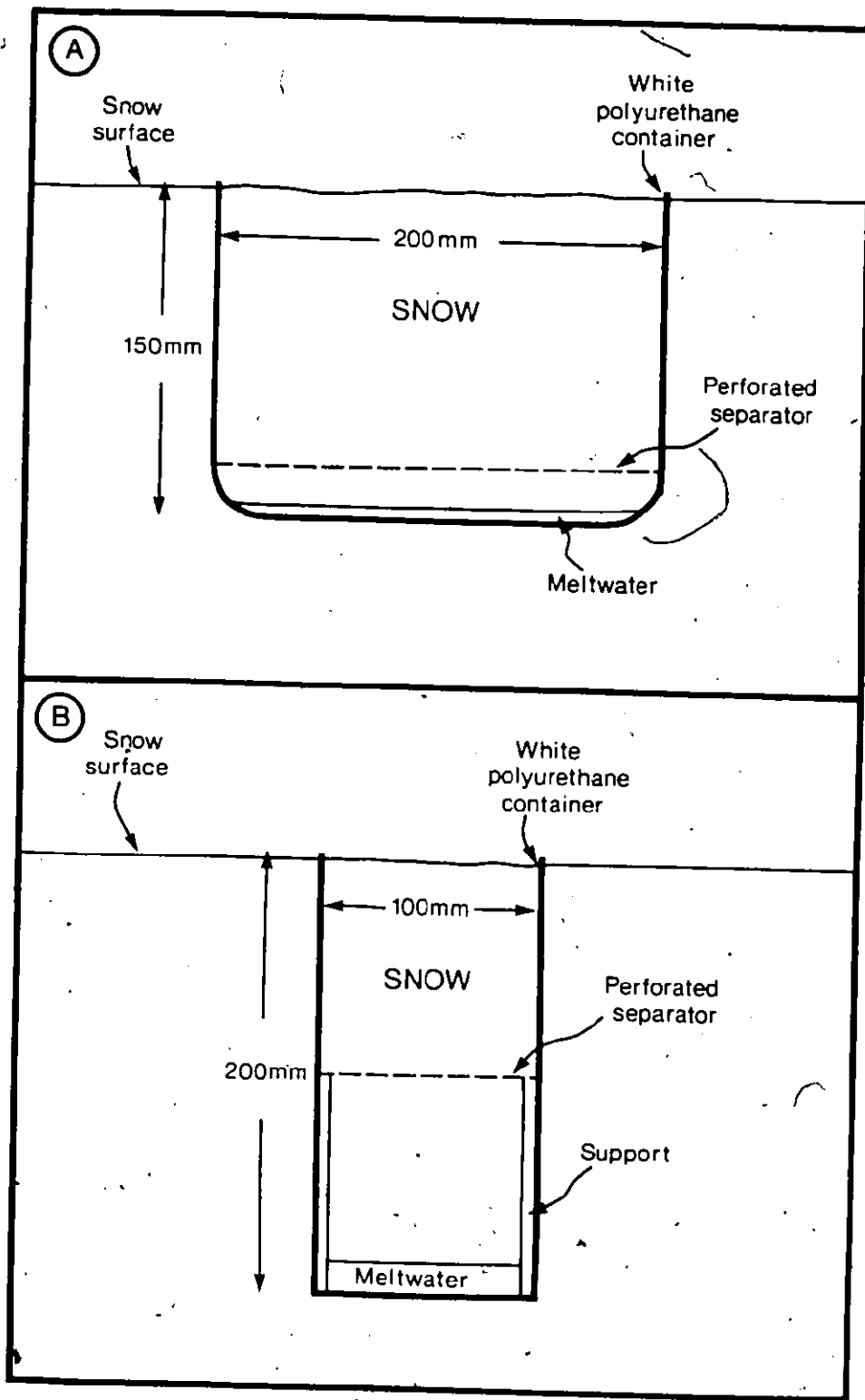


FIGURE 45 : SNOW LYSIMETERS

A - Large type, 1977;

B - Small type, 1978 and 1979.

evaporative losses and sublimation gains. The two levels in the containers were connected by holes which permitted drainage of meltwater from the exposed snow. The lower level, however, proved to be too small to hold all the water, and by the time of next measurement, a mixture of snow and ice often was floating within the container. Clearly, this surface differed considerably from that of the surrounding snowpack. Furthermore, the triple-beam balance was difficult to operate in the field with such large weights and most results had to be discarded as too inaccurate to be useful. In 1978 and 1979, the two large containers were replaced with five smaller ones, 100 mm in diameter and 200 mm deep (Figure 45B). The separator between the two levels was half-way up the container, so that sufficient volume was always available for meltwater to drain freely. Weighing was carried out inside the base camp, eliminating problems with wind. Despite these precautions, results sometimes differed by 100%, but figures averaged from all five lysimeters are thought to be accurate to approximately  $\pm 20\%$ .

Evapotranspiration was measured directly only in 1978 and 1979. Lysimeters of two different sizes were used: small, 73 mm in diameter and 108 mm deep, and large, 84 mm in diameter and 110 mm deep (Figures 46 and 47). Groups of three lysimeters were deployed at the runoff sites, one lysimeter in the unsaturated zone upslope of the snowbank, and the other two downslope in the saturated zone. The unsaturated lysimeter was allowed to dry naturally while the others were kept near saturation through the addition of water approximately equal to the weight lost the previous day. Most readings were made at intervals of 24 hours, but in some cases, longer times elapsed.

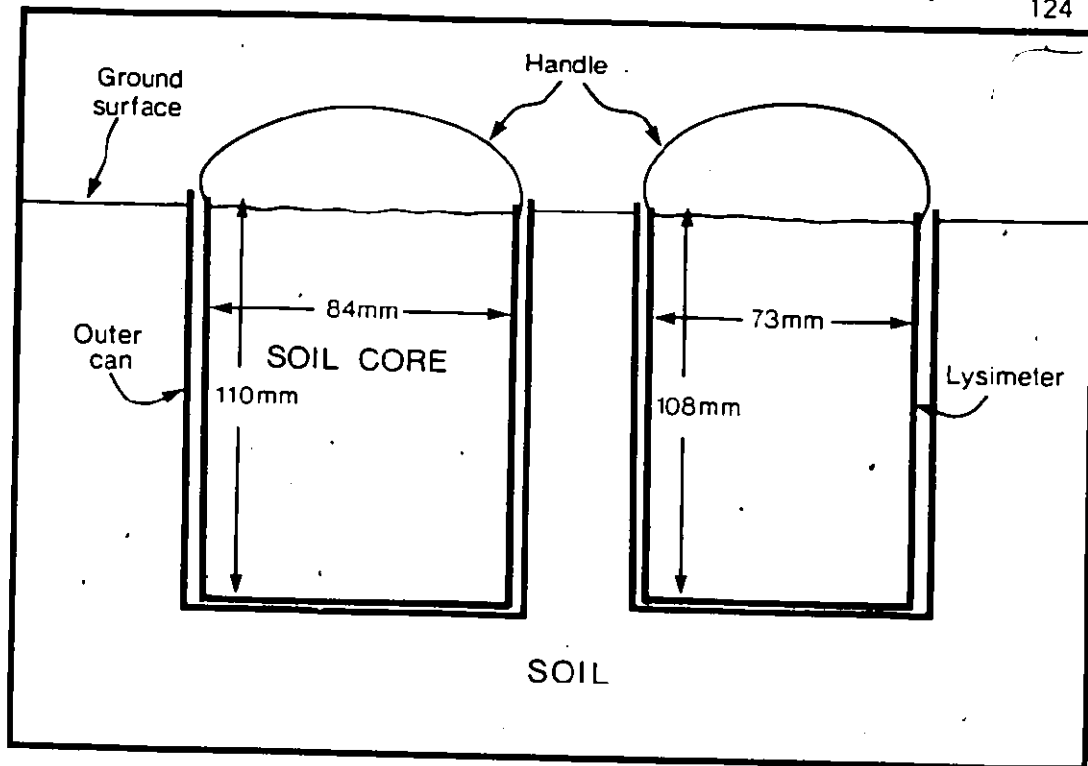


FIGURE 46.: LARGE AND SMALL SOIL LYSIMETERS

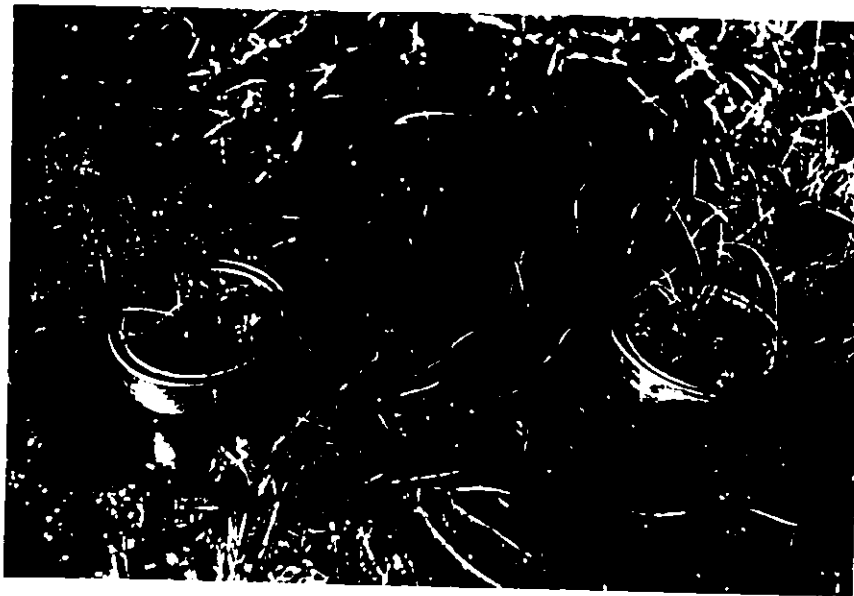


FIGURE 47 : SMALL SOIL LYSIMETERS IN USE, SITE 3, JULY 3, 1978  
Note saturated condition of surrounding surface caused by upslope ablating snowbank.

The lysimeters contained reconstructed cores of soil or undisturbed cores of soil and growing vegetation. A null hypothesis, that no significant differences existed between data collected from saturated lysimeters at the same location was proposed and tested with a Student *t*-test on the standard error of the difference. Data pairs ( $n=40$ ) from the lysimeters at plot 2 in 1978 were compared, and the null hypothesis could not be rejected at the 99% confidence level.

Surface flow was monitored simply in 1977. Water flowing over a runoff plot entered a copper box sealed to the ground surface with silicon sealant (Figures 19 and 48). The water travelled down a 25 mm diameter rubber pipe into a collecting container (60 L capacity). The collecting vessel had been calibrated, so that the depth of water within, measured with a metal metre rule, could be used to calculate volumes to  $\pm 0.01$  L.

The time between emptying the container or between taking successive depth readings was recorded, allowing direct calculation of discharge rates. This method was accurate but unsatisfactory because manual monitoring was required and it proved impossible to record "night-time" data, or flow at more than one site at a time.

During 1978 and 1979, the collecting method remained the same, but the pipe and water were led into a small weir, consisting of a wooden frame with polythene sides and base, and a narrow V-shaped copper weir slot. A stilling well and Leupold-Stevens Type-F water-level recorder (1:1 ratio and 24-hour movement) were installed in each weir (Figures 20 and 49). The weir slot ( $15^\circ$ ) was hand-cut and discharge was determined empirically with a stage-discharge curve (e.g. Figure 50).

The stage-discharge curve was plotted from numerous measurements made of the volumes of water issuing from the weir in short time periods.

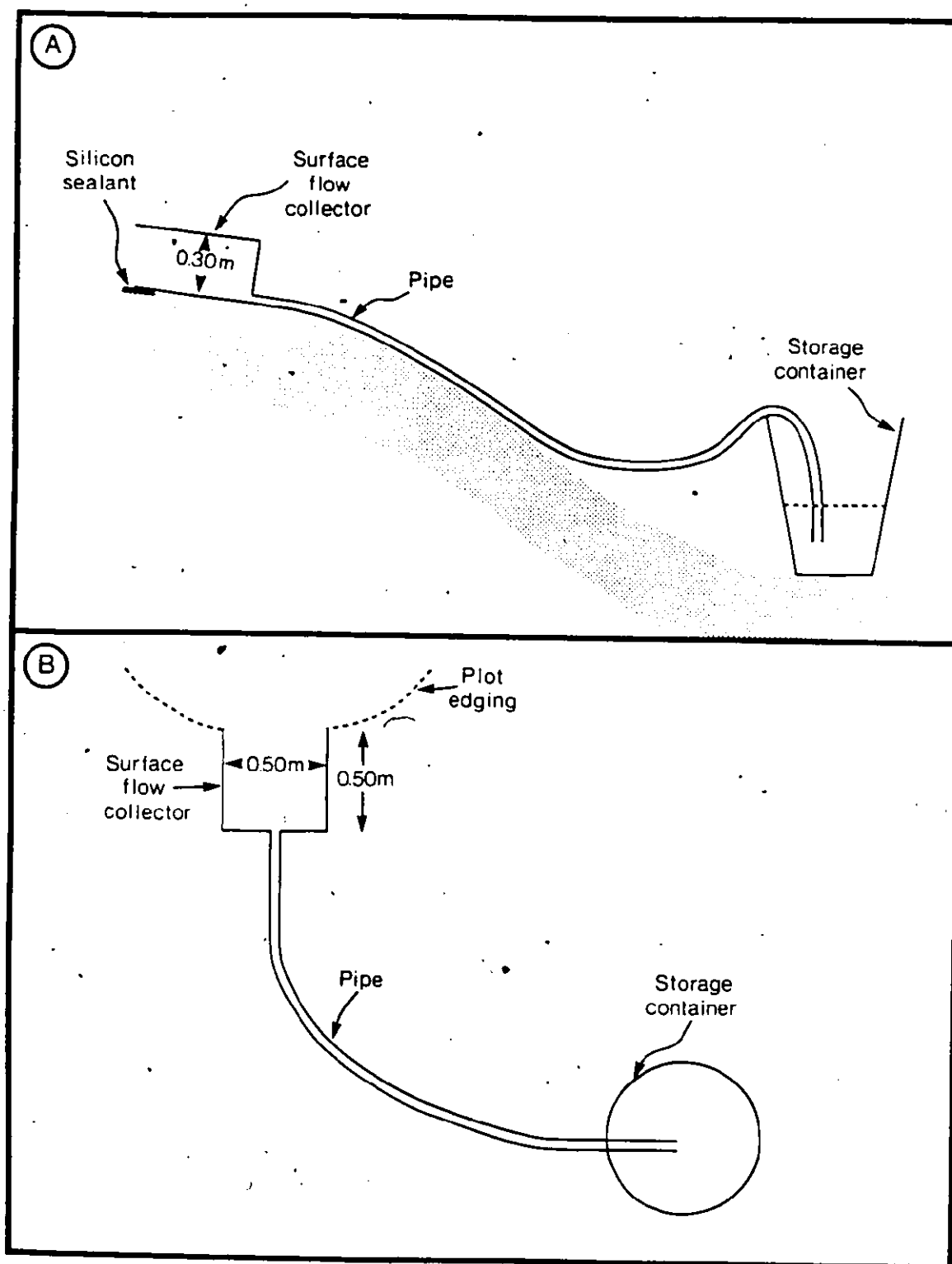


FIGURE 48 : SURFACE FLOW COLLECTION APPARATUS, 1977

A - Cross-slope elevation;

B - Plan view.

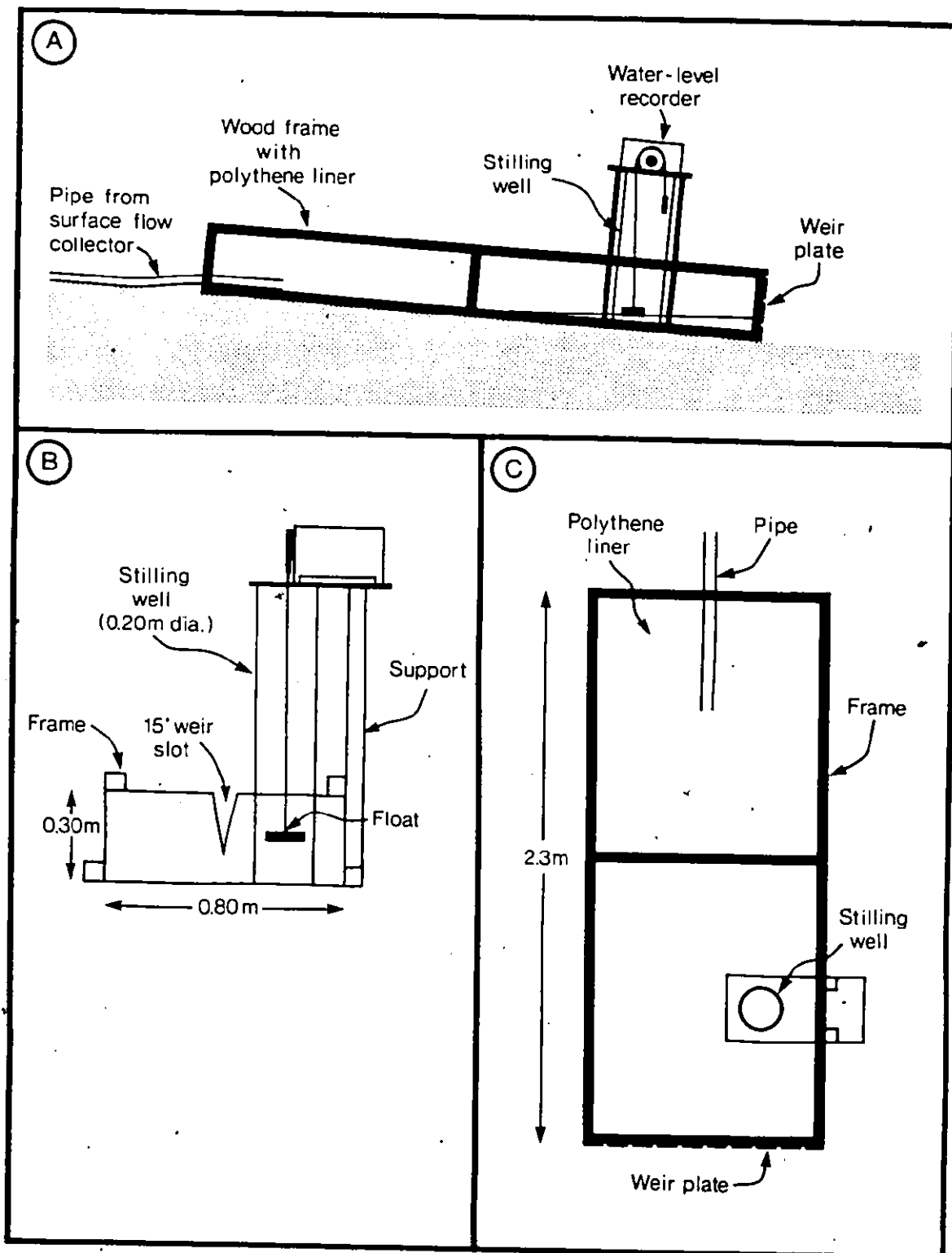


FIGURE 49 : SURFACE FLOW WEIR, 1978-1979

- A - Cross-slope elevation;
- B - Up-slope elevation;
- C - Plan view.

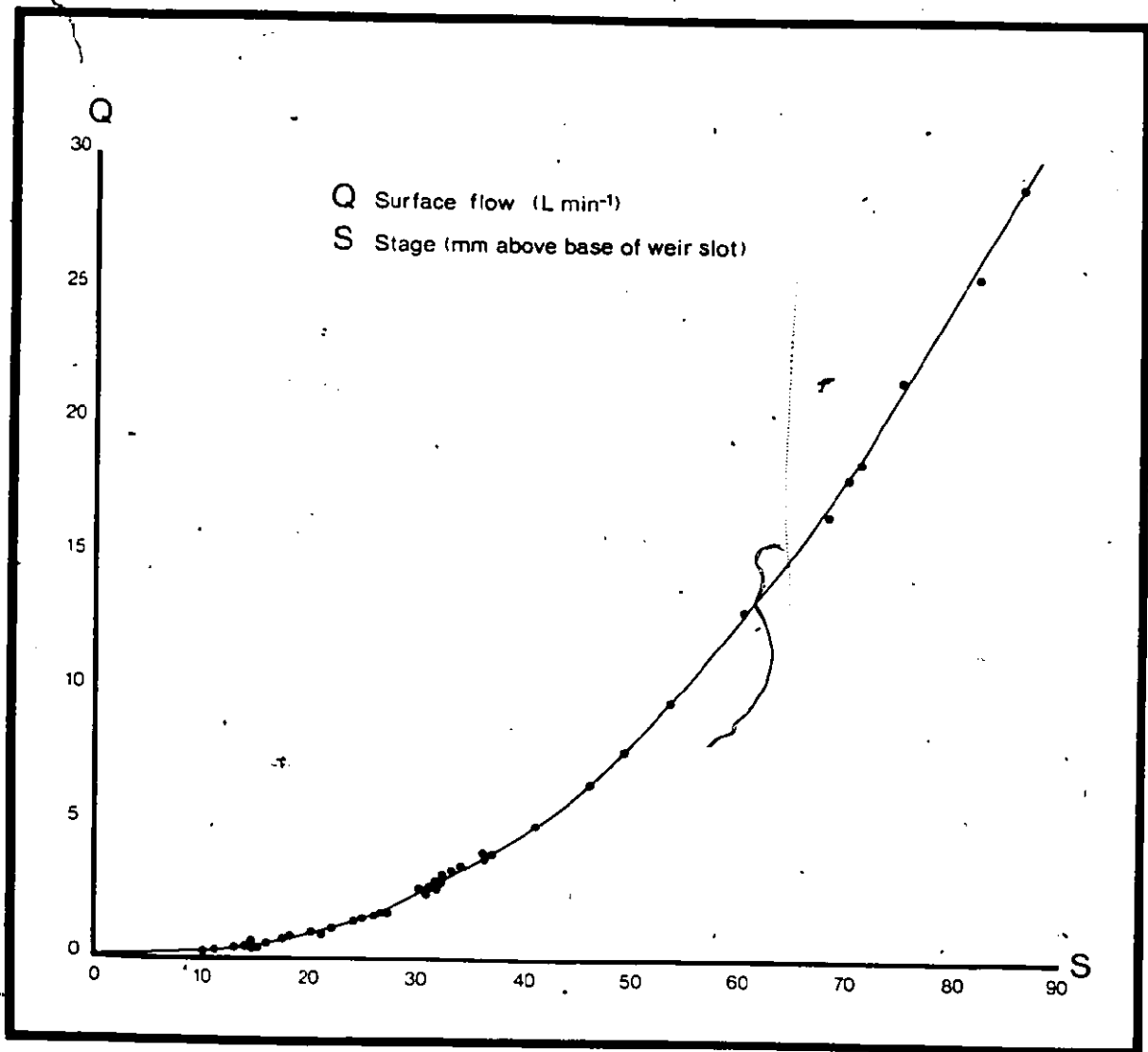


FIGURE 50 : STAGE - DISCHARGE CURVE, PLOT 3 WEIR, 1978

At high discharges ( $20 - 30 \text{ L min}^{-1}$ ), the container utilised to collect the flow filled in only five or six seconds, so errors associated with timing are greatest at high flows. Moreover, the nature of the stage-discharge relationship for a V-notch weir means that at high stages, small errors in reading the chart record are transformed into substantial errors in the discharge record. These problems can be overcome in part by taking a greater number of samples at higher discharges. For example on Figure 50, the points at the lower end of the graph represent one sample of thirty-seconds of flow, while the upper points are an average of five or six samples of six seconds each. A further accuracy problem also was encountered. The straightening of the line at the upper end of the stage-discharge curve (Figure 50) is the result of temporary subsidence of vegetation underlying the stilling well as the water level within the weir (and consequent downward pressure) increased at higher discharges. For this reason, the curve was constructed as an eye-determined line of best-fit.

Additional problems in the use of weirs were small leaks from the polythene, estimated at  $0.1 - 0.2 \text{ L min}^{-1}$  at all sites, which represent less than 5% losses for most discharges, and less than 1% of the seasonal total. Further, icing of the weir slots occasionally occurred during the early part of the ablation season in 1978, causing spurious rises in the stage curves.

In addition to surface discharge monitoring, the velocities of flow and its depth were recorded on a number of occasions. Average velocity on the runoff plots was measured with Rhodamine B dye, and spot measurements were made by timed photography of very thin ( $< 1 \text{ m}$ ) silver foil tracers, a method developed by Savat (1975). The foil tracers were inserted into the flow upstream of the area to be photographed and as they entered the frame,

the shutter was opened for 0.25 or 0.50 seconds. Stripes were left on the film, corresponding to the distance moved by the tracers, and these were compared with a scale, also within the frame (Figure 51). The velocity of tracer movement corresponds to surface flow velocity. Problems derived from camera shake on the tripod during periods of high winds which prevented the use of some photographs for analysis.

Subsurface discharge was measured in the same way during all three field seasons. Monitoring was undertaken at runoff sites 1 and 3 in 1977, and at 3 and 4 in 1978 and 1979. The apparatus consisted of copper guttering welded to a brass frame and inserted into the ground across the slope (Figures 52-54). Flow was diverted by means of a polythene sheet in contact with the rear of the guttering, into plastic pipes which led downslope into storage containers (Figure 54). Discharge rates were calculated either by collecting and measuring all the flow, or by taking instantaneous discharge measurements for 3 - 5 minutes at intervals of approximately one hour.

In 1977, the subsurface flow apparatus was installed at site 1 during the ablation period, and a number of problems resulted. In particular, surface disturbance produced during initial excavation of the frozen ground reduced the potential validity of the results. More care was taken at site 3 and most disturbance was avoided by covering the area around the pit with boards before digging. At site 3, only three gutters were installed at the base of the slope, and the fourth was installed separately within the runoff plot, 0.06 m below the surface. The most successful installation, however, was at site 4 at the end of the 1977 field season (Figure 53). The pit was excavated in thawed, dry ground, and problems associated with pit collapse were avoided. In addition to installation

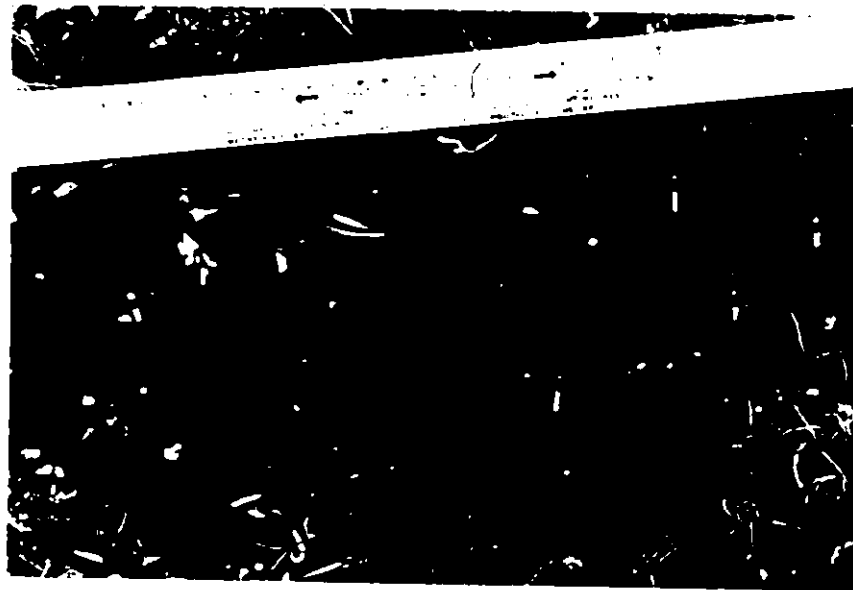
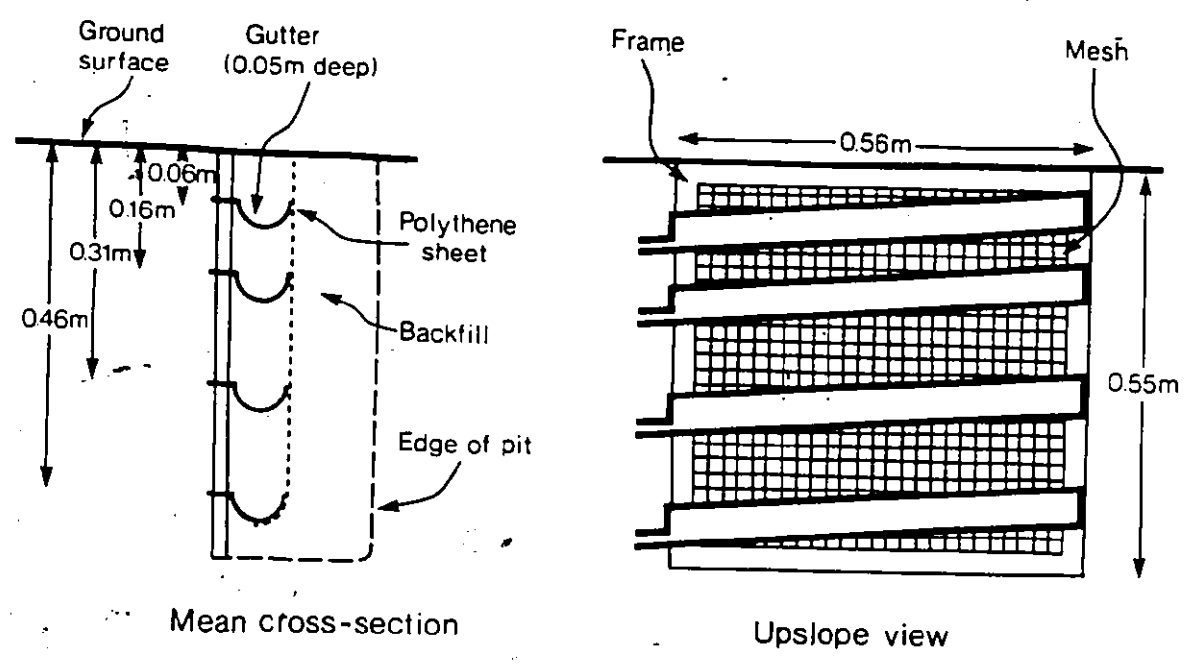


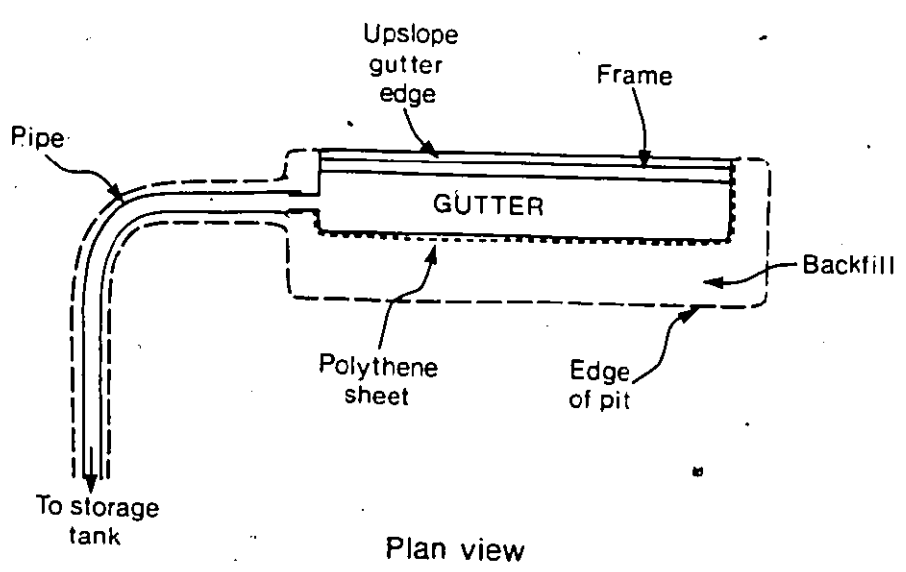
FIGURE 51 : SURFACE FLOW VELOCITY MEASUREMENT TECHNIQUE  
USING THIN SILVER FOIL

Arrowed foil travelled 55 mm in 0.25 seconds, indicating a  
velocity of  $0.22 \text{ m s}^{-1}$ . Other, shorter streaks in the photograph  
indicate lower surface velocities.



Mean cross-section

Upslope view



Plan view

FIGURE 52 : SUBSURFACE FLOW COLLECTOR

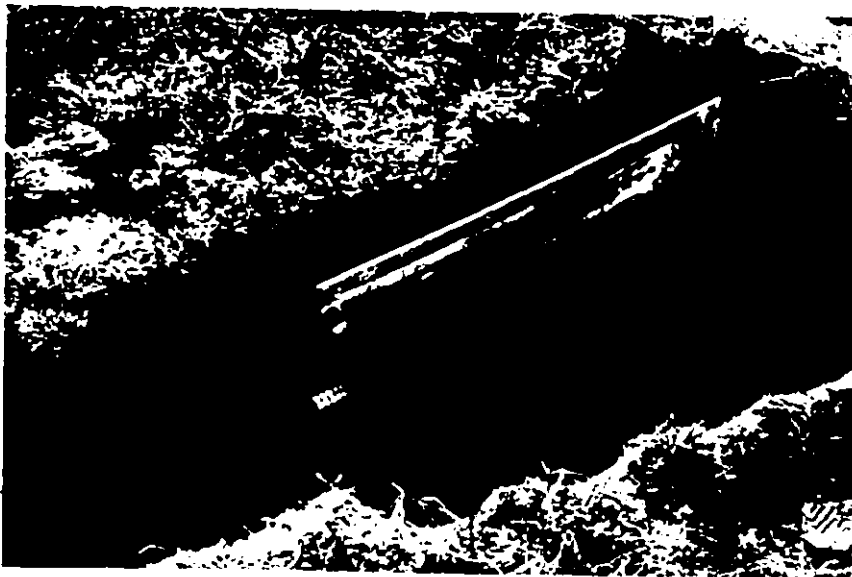


FIGURE 53 : SUBSURFACE FLOW COLLECTOR DURING INSTALLATION.  
AT PLOT 4, JULY 1977

Note lack of disturbance to upslope area, made possible by  
installation late in the field season.



FIGURE 54 : COLLECTION CONTAINERS FOR SUBSURFACE FLOW,  
SITE 3, JULY 23, 1979

Arrow indicates position of buried collector.

problems, however, other difficulties made themselves known later. Twists in pipes and shallow gradients prevented some of the gutters from functioning. Moreover, during thaw it was possible for flow to occur at the level of the guttering, but be prevented from collection by ice within the piping which was set approximately 0.10 m lower in the ground.

The major problem relating to the accuracy of the subsurface flow measurements involves the estimation of the contributory area. If active layer water-tables are low, it may be possible to use frost-table elevations as an indication of flow directions, but this is less likely to be accurate in an active layer close to saturation. Distortions of flow are known to occur in the vicinity of open pits (Knapp, 1973; 1974) but they have not been examined in the area around the type of apparatus used. In the absence of piezometric data, initially it must be assumed that the contributory area on a planar slope is directly upslope. Errors in this assumption can be assessed using data from site 4 where surface and subsurface discharge were investigated at the same position on the slope (see p. 188).

#### Sediment measurement techniques

The following types of sediment movement were examined:

- (1) aeolian sediment inputs;
- (2) suspended sediment in surface wash;
- (3) dissolved sediment in surface and subsurface wash.

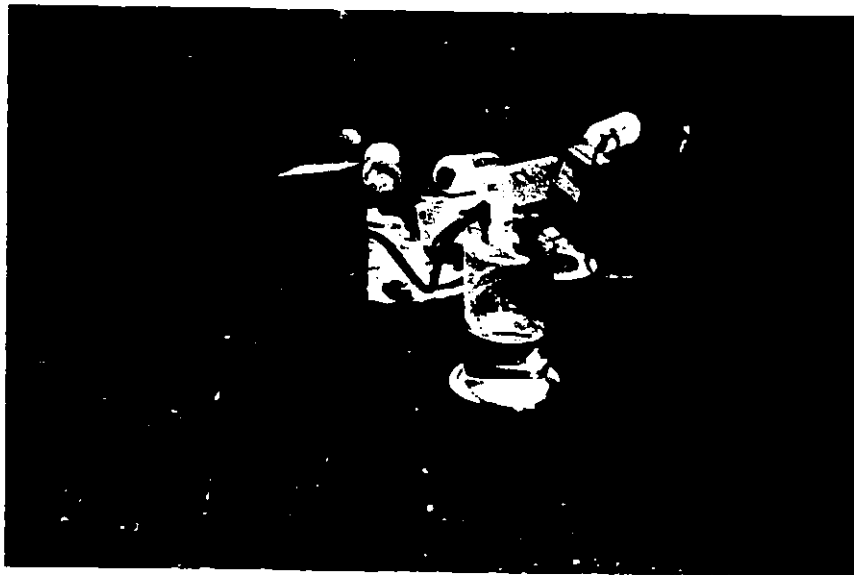
Aeolian sediment concentrations were observed on the snow at all sites during the three field seasons, but concentrations varied considerably from one year to the next. Two sampling techniques were employed. The first method examined the surface concentrations in  $\text{g m}^{-2}$ , by delimiting  $1 \text{ m}^2$  of snow and scraping approximately 10 mm of snow and sediment from it into polythene storage bags. Generally, the maximum

concentration close to the runoff plot was sampled. After returning to the base camp, the snow was melted and the sediment was filtered out through pre-weighed filter papers (see below). Overall accuracy is estimated to be  $\pm 10\%$ . This method was used most frequently because the majority of aeolian particles appeared to be on the snow surface. The second method involved stratified sampling within the pack, or integrated core sampling. Used in conjunction with the first method, these two enable assessment of the timing of aeolian activity during the snow accumulation period. Core samples were treated in the same way as those from the surface snow, but prior to filtering, the volume of each was measured to give the volumetric concentration of particles within the snow. In both methods, accuracy was limited by the filtration method (see pp. 137-138) so that concentrations of less than  $4 \text{ mg L}^{-1}$  could not be determined.

The measurement technique employed for suspended sediment in surface wash was derived from studies in fluvial geomorphology. It involved filtration of a 1 L sample through a pre-weighed filter paper using an Østrem filter pump (Figure 55). After drying in the laboratory, the filter paper was reweighed and the additional weight due to the sediment calculated.

The basic assumption of this method is that the weight of the filter paper remains constant. This was not the case. Problems arose from the extremely low levels of sediment concentration in surface wash, below the level of discrimination of the technique.

During the 1977 and 1978 experiments, filters were weighed before use, and after return to the laboratory they were dried and reweighed. Of necessity, a correction factor had to be applied relating to (1) changes in the moisture content of the filter paper, (2) potential loss of fibres



**FIGURE 55 : ØSTREM FILTER PUMP, 1977**  
Chemtrix conductivity meter is visible in background.  
Note, most filtration was undertaken at the base camp,  
not at the runoff plots.

during the filtration process, and (3) potential changes in the composition of the paper during drying at 105°C. The correction factor of 0.250 g was derived from a sample of five unused papers in 1977. After the 1978 field season, however, 24 filter papers used as controls showed a wide degree of variation, with a mean weight loss following drying of 0.093 g and a standard deviation of 0.030 g. Using the mean correction factor, many of the filter papers appeared to have "lost" weight during use. Clearly this was incorrect and the problem was presumed to lie in the correction estimate. It was considered too inaccurate for the very small amounts of sediment involved. This cast doubt on the validity of the earlier results which might have shown a similar variation had a larger control sample been used.

Tests subsequently were run in which 25 filter papers were dried for 48 hours at 105°C before weighing, then were dipped in distilled water, redried for 48 hours and reweighed. In both weighings a balance accurate to  $\pm 0.005$  g was used, and the papers were weighed within 10 seconds of their removal from the oven. This control group showed losses of 0.002 - 0.010 g with a mean of 0.007 g and a standard deviation of 0.002 g. The losses were ascribed partly to losses during washing when some fibres came loose from the filter paper, and partly to slight oxidation during the drying process.

In 1979, all filter papers were subjected to pre-drying and pre-weighing before use, and a correction factor of 0.007 g was applied. Some filters still showed slight "negative" sediment loads but this was to be expected. If concentrations of less than 4 mg L<sup>-1</sup> exist in the sample, it is probable that a small percentage of the filter papers will experience losses greater than 2 standard deviations of 0.002 g, causing negative

results. The measurement process, therefore, is accurate to approximately  $\pm 4 \text{ mg L}^{-1}$  and loads which appear negative are classed with others in a group of  $< 4 \text{ mg L}^{-1}$ .

Dissolved solids in surface and subsurface wash were monitored using three techniques. A Chemtrix type-70 conductivity meter with automatic temperature compensation was used to measure the conductivity of samples in  $\mu\text{siemens cm}^{-1}$  compensated to  $25^\circ\text{C}$ . A factor of 0.65 was used to convert  $\mu\text{siemens cm}^{-1}$  into total dissolved solids, expressed as  $\text{mg L}^{-1}$  (Rainwater and Thatcher, 1960). Error in the instrument can be assumed to be less than 1% full scale deflection, which at most concentration levels is  $\pm 2 \mu\text{s cm}^{-1}$ . The second method measured total, calcium, and magnesium hardness (expressed in  $\text{mg L}^{-1} \text{CaCO}_3$ ), using the Schwarzenbach titration procedure on 50 mL samples which were less than 24 hours old. Accuracy in this method normally is  $\pm 0.1 \text{ mg L}^{-1}$ , but under field conditions more realistically is  $\pm 1.0 \text{ mg L}^{-1}$ . The third method involved laboratory analysis of selected samples (undertaken by T. Gallie, Department of Geography, University of British Columbia), using a Perkin-Elmer 373 atomic absorption spectrophotometer with an acetylene-oxygen flame for Mg, Na and K, and  $\text{NO}_2$  flame for Ca.  $\text{SiO}_2$  was analysed colorimetrically on a Bausch and Lomb Spectronic-70 using the standard "moly-blue" method (Rainwater and Thatcher, 1960). Comparison of the three methods is shown below (Table 29, p. 226).

CHAPTER 5

HYDROLOGIC ASPECTS OF SLOPEWASH

## Annual water balance

### (a) General characteristics

The importance of each of the hydrologic processes operating at the four runoff plots can be assessed better if it is examined first within a regional context. The water balance of the Thomsen River upland area can be estimated using snow and evaporative data, with stream runoff calculated as a residual.

The basis of the water balance assessment is the successive mapping of snow-free areas during the summer of 1979 (see Figure 15). This was accomplished by 15 km ground traverses of the upland (covering an area of 7 km<sup>2</sup>), during which the edges of snow-free areas were marked on air photographs (scale 1:13,000). Errors in the method are thought to be within 5% of the calculated snow-free area since checks were possible by comparison of successive maps. Snow cover was complete on May 29, 1979, and maps were produced for June 10, June 12, June 18, June 27, and July 6. Based on observations of the snow cover remaining on July 25, it is estimated that the upland was completely free of snow on August 5, 1979.

For each date, the upland area can be divided into three zones:

- (1) snow-covered areas, presumably subject to evaporative losses;
- (2) snow-free areas located upslope of snow-covered areas (e.g. ridges and interfluves), presumably experiencing low rates of evapotranspiration;
- (3) snow-free areas downslope of snowbanks (e.g. valley floors), presumably experiencing high evapotranspiration rates. The changes in the proportions of the upland area that fall within these divisions are shown in Figure 56. Estimates are made for an additional day, July 3, on the assumption that the saturated areas reached a maximum on that date (Table 9).

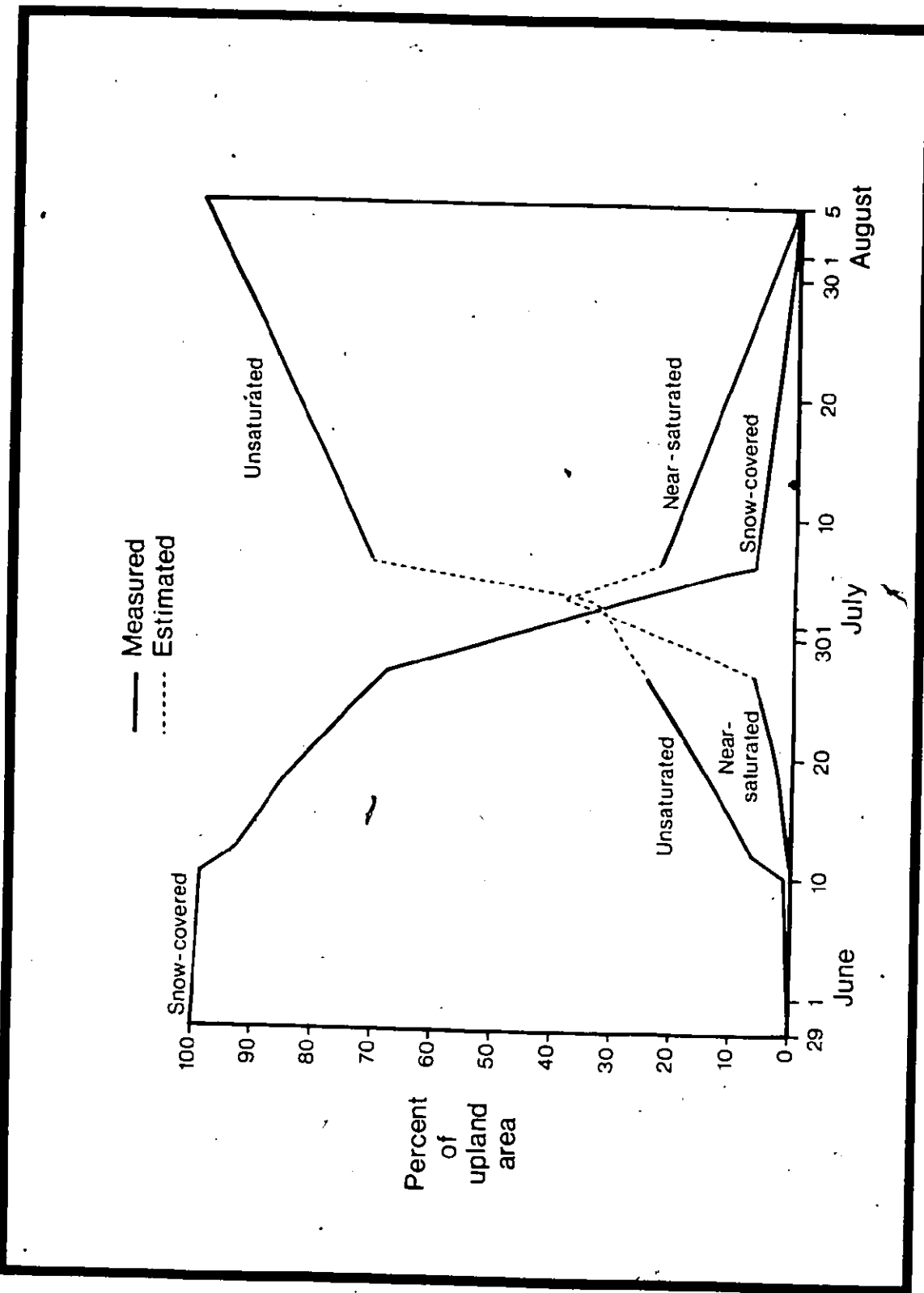


FIGURE 56 : CHANGES IN THE PROPORTIONS OF SNOW-COVERED, UNSATURATED AND NEAR-SATURATED AREAS, THOMSEN RIVER UPLAND, MAY 29 - AUGUST 5, 1979

Date	Percent snow-covered (evaporation rate: 0.56 mm d <sup>-1</sup> )	Percent unsaturated (evaporation rate: 1.04 mm d <sup>-1</sup> )	Percent near-saturated (evaporation rate: 3.24 mm d <sup>-1</sup> )	Average evaporation rate (mm d <sup>-1</sup> )
May 29	100.0	0.0	0.0	
June 10	98.9	1.1	0.0	0.56
June 12	92.8	6.7	0.6	0.59
June 18	84.9	13.0	2.1	0.64
June 27	67.9	25.2	6.7	0.77
July 3	28.0 <sup>a</sup>	33.0 <sup>a</sup>	39.0 <sup>a</sup>	1.31
July 6	6.9	70.7	22.4	1.63
August 5	0.0	100.0 <sup>a</sup>	0.0	1.27

<sup>a</sup> Estimated data

TABLE 9: Evaporative area changes in the Thomsen River upland, May 29–August 5, 1979

The three designated zones are assigned average evaporation rates based on lysimetric data. In reality a continuum of rates exists between ground denoted saturated and unsaturated, but for practical purposes a single value for each state, established over a wide range of meteorological conditions can be used. The snow evaporative rate applied is  $0.56 \text{ mm d}^{-1}$ , being the average of 245 lysimeter days, derived mainly from pre-isothermal snow during 1978 and 1979. This value is greater than others recorded in Arctic areas (see p. 81). It is possible that additional evaporation was promoted by the lysimeters themselves, following radiation absorption. The rates of ground evapotranspiration used in Table 9 are  $1.04 \text{ mm d}^{-1}$  for unsaturated areas, and  $3.24 \text{ mm d}^{-1}$  for areas at or near saturation. These values are derived from 395 and 74 days of lysimetric data respectively, recorded during melt and post-melt periods in 1978 and 1979. These figures lie within the range of evapotranspiration rates mentioned earlier (see Table 5, p. 83) and suggest that small lysimeters are an acceptable measurement technique for this process. Equally, it is clear that the division of the upland into three areas is justified by the considerable differences in values for the three rates. Average evaporation rates, weighted for the areas falling into each category, are shown for the periods between snow mapping in Table 9.

The snow-free area mapping, in combination with measured ablation and density data, permits an estimate to be made of the initial water equivalent of the snow in the upland. Average total ablation at the four sites between successive map dates, is multiplied by average upland snow-covered area for the same time periods. The sum of these volumes, multiplied by average initial snow density at sites 1 and 3 (representing interfluvial and snowbank locations respectively), results in an approximate

value for the initial water equivalent of the upland snowpack. This is estimated to be 174 mm for 1979 (Table 10). Since rain during the melt season was 3.6 mm, and assuming no change in soil moisture content between the pre-melt period and after the cessation of melt, stream runoff can be calculated as a residual. For 1979, stream runoff is estimated to be 107.4 mm (Table 10), representing 60% of the total precipitation. This is less than the "typical" value of 150 mm for stream runoff in polar desert areas suggested by Wedel (1979). Since the drainage density of the upland is approximately  $2.75 \text{ km km}^{-2}$  (see p. 34), unit-slope discharge calculated with equation (5) (p. 85), is  $19.5 \text{ m}^3 \text{ m}^{-1} \text{ a}^{-1}$ .

The estimated percentage of precipitation leaving the upland as runoff (Table 10) is less than that recorded on Cornwallis Island where it varies between 67% and 81% (Marsh and Woo, 1979). This may be partly the result of (1) the low rainfall total, and (2) the particularly lengthy ablation season in 1979, which sustained high evapotranspiration losses. For these reasons, a second set of water balance data was generated using the same snow input data, but increasing rainfall to 14.5 mm (the average during the 1977 - 1979 ablation seasons) and setting the date when the area became snowfree to July 28. This date was obtained by averaging the number of days between site 1 and the rest of the upland becoming snow-free for 1977 - 1979, and adding this to the 1979 date for site 1. Evapotranspiration totals are reduced by these changes, and runoff increases to 130.1 mm, 69% of total precipitation (Table 10). Unit slope discharge increases to  $23.6 \text{ m}^3 \text{ m}^{-1} \text{ a}^{-1}$ , a figure reasonably similar to others calculated for Bathurst Island (see p. 86).

Unfortunately, it is not possible to assess the probability of occurrence of the snow cover used for the calculations. A comparison of the water equivalents of snow accumulated at sites 1 and 2 (derived from snow

	INPUTS		OUTPUTS		Unit Slope discharge ( $m^3 m^{-1} a^{-1}$ )
	(mm)	(%)	(mm)	(%)	
1979	Snow: 174.1	98	Evaporation from snow: 16.5	9	19.5
	Rain: 3.6	2	Evapotrans- piration: 53.8	31	
	Total: 177.7		Stream runoff: 107.4	60	
Average 1977- 1979	Snow: 174.1	92	Evaporation from snow: 16.4	9	23.6
	Rain: 14.5	8	Evapotrans- piration: 42.1	22	
	Total: 188.6		Stream runoff: 130.1	69	

TABLE 10: Water balance components, Thomsen River upland

course and density measurements) with winter snowfall totals at Sachs Harbour reveals no obvious correlation. It appears that more snow accumulated in the Thomsen River area during the winters of 1977 - 1978 and 1978 - 1979, than in 1976 - 1977, but it is not possible to state which, if any, of these was an average season.

(b) Runoff plots and inter-year variability

The water balance of a runoff plot is obtained by equating inputs of snow and rainfall over a fixed area, with outputs of evaporation and water passing downslope through a vertical plane at the exit of the plot.

Plot 2 remained in the same position for all three study years and provides the best data set for analysis of the inter-year variability of the water balance components. In addition, it is typical of many snowbank locations in the Thomsen River area. Initial snow data were collected at the site either at the end of May or at the beginning of June, and data collection continued until the cessation of snowmelt (Table 11). To standardize the collection periods and render them comparable, the dates when the snow became isothermal at  $0^{\circ}\text{C}$  are used to divide the data. In 1977, for example, snow data collection commenced on the same day that the snow became isothermal. In 1978, collection began one month before the snowbank reached  $0^{\circ}\text{C}$ .

The values of water inputs and losses shown in Table 11 are calculated in a number of ways. The initial and isothermal snow water equivalents are derived by multiplying the volume of snow on the plot by average density values for the respective dates. The rainfall data are abstractions from the gauge records. The evaporative losses are calculated from the average evaporation rates for snow, unsaturated and near-saturated ground (see p. 143), weighted for the areas of the plot falling into these

	DATES	INPUTS		LOSSES						RESIDUAL							
		Snow on plot mm	Rain mm	Snow evaporation mm	Unsaturated evaporation mm	Moisture saturated evaporation mm	Flow through snow mm	Surface flow mm	mm	\$							
1977	Post-isothermal period June 5	260.2	98	6.5	2	4.0	1	5.7	2	4.6	2	0.0	0	67.5	26	184.9	69
	Pre-isothermal period June 19																
1978	Post-isothermal period May 25	233.9	100	0.0	0	14.8	41	3.4	9	1.0	3					16.7	47
	Pre-isothermal period June 25																
1979	Post-isothermal period June 25	198.0	99	1.2	1	2.3	1	3.5	2	5.0	3	0.0	0	22.4	11	166.0	83
	Pre-isothermal period July 4																
1979	Post-isothermal period May 29	317.6	100	0.0	0	16.3	50	3.5	11	0.0	0					12.7	39
	Pre-isothermal period June 30																
Av. 1977	Post-isothermal period June 30	285.1	99	3.6	1	3.7	1	5.2	2	4.9	2	201.0	69	7.4	3	66.5	23
	Pre-isothermal period July 13																
1979	Post-isothermal period	100				45			10	2							43
	Pre-isothermal period	99		1		1		2		2		23	13				58

TABLE 11: Water Balance, plot 2, 1977-1979

divisions. Flow through the snow is shown if the snow is isothermal and the plot exit still possesses a snow cover. It is calculated as the residual of water equivalent changes in the snowpack minus evaporative losses. Although surface flow was monitored continuously in 1978 and 1979, the figure for 1977 includes estimates for gaps in the record. Data to fill these omissions were predicted using simple regression of surface flow against lagged net radiation (see Lewkowitz, 1978, pp. 73 - 76). The residuals shown in Table 11 are calculated from total inputs minus total known losses for the pre-isothermal and post-isothermal periods.

Known losses during the pre-isothermal period are chiefly from snow evaporation. Ground evapotranspiration is of lesser significance because of the small areas that are snow-free. A major residual exists, averaging over 40%, and requires explanation. It is hypothesised that some parts of the snowpack reach 0°C before others and that water can issue from the snowbank even before the stated isothermal date. The values of flow through the snow, however, are calculated only for the post-isothermal period. These figures, therefore, probably are conservative estimates and much of the pre-isothermal residual is likely to be flow through the snow.

Average losses after the isothermal date are dominated not only by flow through the snow, but also by surface flow and a second large residual. Evaporative losses during the actual ablation period are relatively unimportant. Surface flow averages only 15% of the post-isothermal losses, despite its apparent significance in visual terms at the runoff plot. The large residual likely results from a combination of subsurface flow, storage within the active layer in preparation for subsequent evaporation, and possibly, measurement error.

The degree of inter-year variability in the absolute and percentage values of the components shown in Table 11 is striking. There is little relative variation in snow accumulation and rainfall values, but there are great changes in output values.

For example, flow through snow was zero in 1977 and 1978, but reached 201.0 mm in 1979, accounting for 69% of the post-isothermal losses. These and other variations are thought to be the results of differing meteorological conditions during both the snow accumulation and ablation periods.

In 1977, the first year of study, a particularly early thaw occurred (see Table 3, p. 18) and only a few hours of records exist for the period prior to the snow becoming isothermal at plot 2. The snowbank at this time was thin both at the crest and at the base of the slope. A maximum thickness of 1.2 m was reached in its central area (Figure 57). The surface flow collector was sited at the lower end of the snow profile and collected discharges as the snow ablated. Between June 9 and June 14, 1977, above-average air temperatures and near-continuous sunshine caused rapid ablation. The rate of meltwater production exceeded the threshold necessary to overcome subsurface and evaporative losses and generate surface flow. The surface runoff coefficient following isothermal conditions was 26%.

In contrast, the 1978 and 1979 thaw seasons commenced very late. Moreover, the snow distributions on plot 2 differed considerably from that of 1977 (Figure 57). Snow depths of 0.4 m and 0.9 m existed at the location of the surface flow collector in 1978 and 1979 respectively, so that these amounts of ablation were required before the collector position was clear of snow. In 1978, the necessary clearance occurred by evaporation and settling

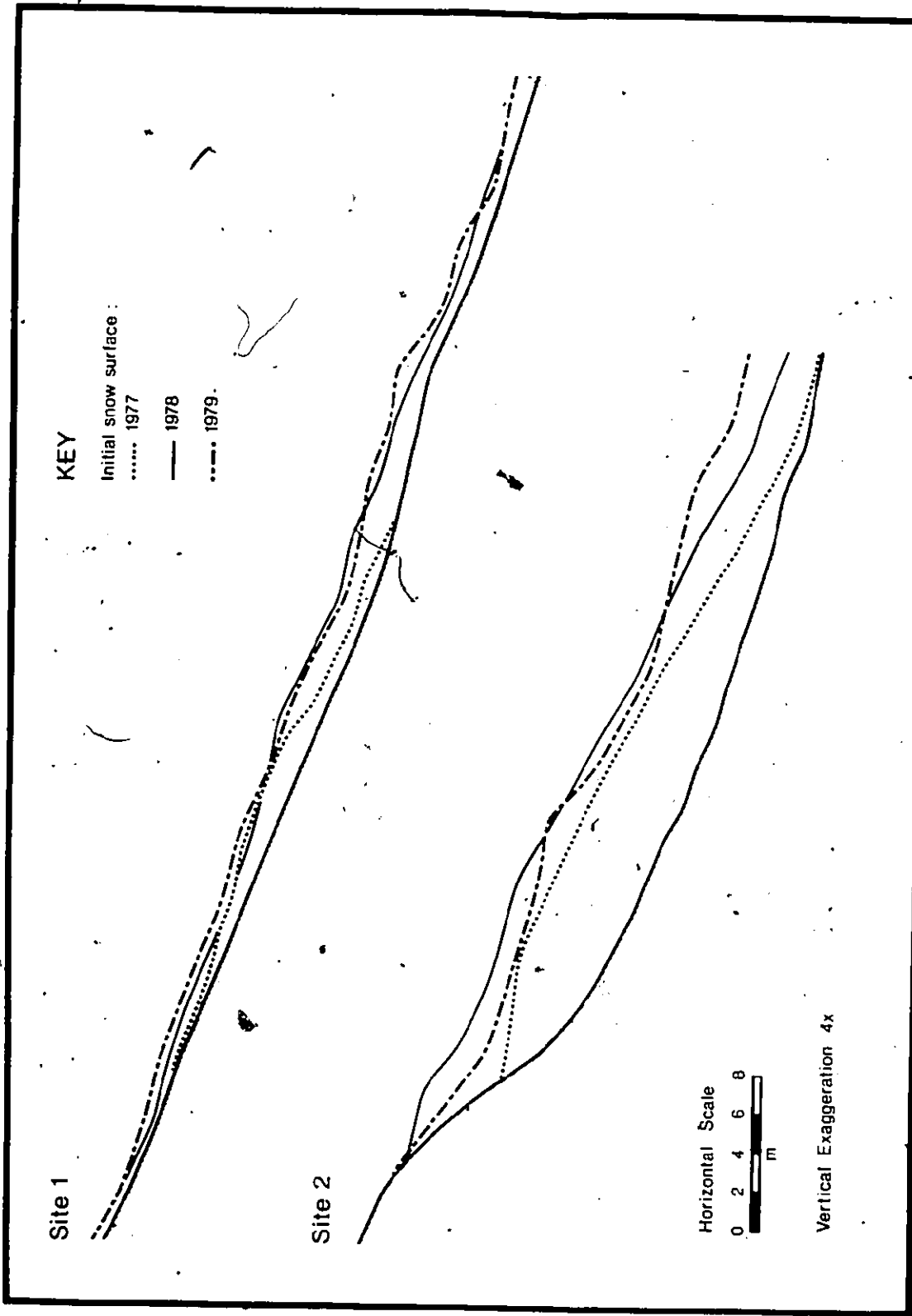


FIGURE 57 : SLOPE PROFILES AND SNOW THICKNESSES, SITES 1 AND 2, 1977-1979.

before the isothermal date, but in 1979, a large volume of water is thought to have left the plot as downslope flow through the snow (Table 11).

An additional variation attributed to the changes in snow distribution concerns the amount of snow remaining on the plot when surface flow was first generated. This volume varied by a factor of three, from 79.8 mm in 1979 to 260.2 mm in 1977. The post-isothermal residual values also illustrate the effect of the greater accumulation in the lower area of the plot in 1979. Those for 1977 and 1978 are quite similar, both in magnitude and percentage value, while that for 1979 is much less, as a result of the loss of water through the snow.

To summarize, it is clear that not only the volume of winter snow accumulated at a snowbank site is important, but also the distribution of snow on the slope. In the limiting case, a uniform thickness of snow will produce no surface and subsurface wash per se, but will be depleted solely by evaporation and flow through the snow.

It is also instructive to consider the inter-year contrasts and the general characteristics of the water balance at an interfluvial site. Site 1 is of this type. Complete data sets exist only for 1977 and 1979, so a three-year average has not been calculated (Table 12). No rain was recorded during either of the two ablation periods and snow is the only input. In both years, the snowpack was of variable thickness (Figure 57) and disintegrated into discrete units as it melted. This led to relatively large values for near-saturated evapotranspiration since much of the snow-free ground on the plot was affected by melting snow upslope.

At site 1, surface flow was generated only during two short 3 - 4 day periods. The amount of snow present on the plot when generation first occurred varied by a factor of three, but the pattern displayed was the

	DATES	INPUTS			LOSSES					RESIDUAL								
		Snow on plot mm	Rain mm	Snow evaporation mm	Unsaturation- evaporation mm	Near saturated evaporation mm	Flow through snow mm	Surface flow mm	mm									
1977	Pre-isothermal period May 28 -- May 29	71.1	100	0.0	0	0.5	3	0.0	0	0.1	1						14.3	96
	Post-isothermal period May 29 -- June 4	56.2	100	0.0	0	1.4	2	3.3	6	4.6	8	0.0	0	6.4	12		40.5	72
1979	Pre-isothermal period May 29 4 -- June 28	85.2	100	0.0	0	16.2	35	0.5	1	1.5	3						27.9	61
	Post-isothermal period June 28 -- July 5	39.1	100	0.0	0	2.0	5	1.8	4	5.3	14	0.0	0	19.0	49		11.0	28

TABLE 12: Water balance, plot 1\*, 1977 and 1979

\*N.B. Boundaries of the runoff plots for the two years are different (see Figure 27)

opposite of that shown at plot 2 (i.e. the smaller amount remained in 1977). In 1977, surface flow generation began three days after the pack became isothermal. During these three days, air temperatures remained positive, but radiation inputs were low. Snowmelt proceeded too slowly to exceed subsurface and evaporative losses and thereby generate surface flow. Little snow was left when weather conditions changed and sunshine became continuous for the period June 1 - 3, 1977. In contrast, while standing water was visible on the plot for seven days before runoff was generated on June 29, 1979, continuous water loss was prevented by heat deficits in the snow at "night" as air temperatures fell below 0°C. As in 1977, however, surface flow finally was generated during a period of high radiation inputs coincident with increased air temperatures. The large residuals in Table 12 in both the pre-isothermal and post-isothermal periods are attributed to subsurface flow and active layer recharge in the periods prior to and during surface flow production.

In summary, data from plot 1 indicate that surface flow production at an interfluvial site is dependent on a combination of high air temperatures and radiation inputs. Had the meteorological conditions observed during the initial part of the isothermal periods prevailed, the melt thresholds necessary to generate surface flow would not have been exceeded.

The magnitude of water volumes leaving plots 1 and 2 as subsurface flow during snowmelt can be estimated by comparing the residuals shown in Tables 11 and 12 with lysimetric data for the period following ablation at the sites. This comparison is accurate, however, only if evapotranspiration data are available up to the freezeback period, and if it is assumed that the overall water content of the active layer does not change between thaw and freezeback. This length of record is available only for site 2 in 1978.

The residual in the post-isothermal water balance for 1978 at plot 2 is 166.0 mm (Table 11). Total water loss from near-saturated lysimeters after the end of snowmelt (July 4) to new snow cover (August 14) was 22.2 mm, but unsaturated lysimeters actually gained water (4.8 mm) over this period as a result of rain in July and light snow on every day in August. The overall loss from the plot, weighted for the areas falling into the two categories is 5.1 mm. In this case, therefore, evapotranspiration explains little of the residual figure, and assuming small measurement errors, the great majority of the 166.0 mm must have left the plot by subsurface flow. This is a distinct possibility at plot 2 since the surface flow collector is sited on moss, and presumably water leaves the plot through the vegetation mat.

While exact figures cannot be calculated, it is useful to compare the partial evaporation records for 1979 with the residuals in the water balances. The residual at plot 2 for 1979 is 66.5 mm. Using the weighting technique outlined above, only 7.6 mm of water evaporated from the plot between July 13 - July 20. Similarly, at plot 1 in 1979, 14.1 mm of water evaporated from the surface between July 5 - July 20. The post-isothermal residual for this plot, however, is only 18.1 mm (Table 12). At first sight, this implies that only 4.0 mm of water could have left the plot as subsurface flow, a great contrast to the plot 2 figures. Because this is only a partial summer record, however, this implication may be incorrect. It is possible for additional water to have left the plot as subsurface flow during and immediately following melt if the subsequent evaporation is the result of ground ice thaw.

In summary, it appears that the greater part of each post-isothermal residual shown for plot 2 (Table 11) is the result of subsurface flow, and not evapotranspiration. At site 1, where limited data indicate

smaller residuals, a larger proportion may be the result of post-snowmelt evapotranspiration.

(c) Inter-site variability

Data from plot 2 enable examination of inter-year variability. It is pertinent to consider, however, the degree to which this site is representative of other snowbank locations in the study area.

Table 13 shows the water balances of plots 3 and 4 for 1978 and 1979. They may be compared with the data for plot 2 (Table 11). In all cases, snow is the most important hydrological input, but there are large differences among the winter accumulations on the runoff plots. One effect of these is the duration of the snow ablation period. For example, the snow on plot 2 ablated eight days earlier than on plot 3 in 1977, and a similar pattern was displayed in 1978. This difference can have important effects on the water balance figures if a longer-lasting snowbank is subject to a period of particularly high, or low, energy inputs after other snowbanks have ablated. A second difference in the data shown in Tables 11 and 13, concerns the apparent increases in water on the plot between pre-isothermal and isothermal snow conditions shown in the latter. The increases at plot 4 can be explained as flow downslope from the upper parts of the snowbank, since this plot does not run all the way to the slope crest. The increase at plot 3 in 1979, however, cannot be attributed to this process, and may be measurement error. In this case, the earlier figure may be an underestimate, the result of an inability to sample the snow density at the base of the site 3 snowpack with only a 2 m snow tube.

Figures in Table 13 indicate that total evaporative losses constitute only a small percentage of total water outputs at a snowbank

	DATES	INPUTS				LOSSES						RESIDUAL	
		Snow on plot mm %	Rain mm %	Snow evaporation mm %	Unsaturation mm %	Unsaturation mm %	Unsaturation mm %	Flow through snow mm %	Surface flow mm %	Snow remaining on plot when monitoring ceased mm	%		
PLOT 3	1978	Pre-isothermal period May 26 -- June 14	710.2 100	0.0 0	10.8 17	0.7 1	0.2 <1				52.8 82		
		Post-isothermal period June 14 -- July 15	645.7 >99	1.2 <1	9.2 1	10.5 2	11.3 2	0.0 0	525.7 81		90.2 14		
	1979	Pre-isothermal period May 29 -- June 30	894.5 N.D.	0.0 0	17.2 N.D.	0.2 N.D.	1.5 3				-106.9 N.D.		
		Post-isothermal period June 30 -- July 23	984.0 >99	3.6 <1	8.0 1	7.3 1	0.5 <1	234.2 24	242.0 25		78.6 8	417.0 42	
PLOT 4	1978	Pre-isothermal period May 27 -- June 14	912.0 N.D.	0.0 0	9.7 N.D.						-138.1 N.D.		
		Post-isothermal period June 14 -- July 16	1040.4 >99	1.2 <1	13.3 1	3.1 <1	14.1 1	518.2 50	303.0 29		189.9 18		
	1979	Pre-isothermal period May 29 -- July 1	454.3 N.D.	0.0 0	16.7 N.D.						-68.8 N.D.		
		Post-isothermal period July 1 -- July 24	1006.4 >99	3.6 <1	12.3 1	1.5 <1	0.9 <1	0.0 0	283.7 28		508.5 50	203.1 20	

TABLE 13: Water balance, plots 3 and 4, 1978 and 1979

location. Moreover, percentage loss appears to be larger at small snowbank sites (up to 13% at plot 2) and smaller at large snowbank sites (3 - 6% at plots 3 and 4).

The 1978 results for the three snowbank sites show that surface flow was least important in absolute and percentage terms at plot 2, and most important at plot 3 where it made up 81% of the water equivalent of the post-isothermal snow. Although incomplete, since monitoring ended before all snow had ablated, the 1979 data display similar trends.

The post-isothermal residuals shown in Tables 11 and 13 generally exhibit a trend opposing that of the surface flow figures. The highest percentage value in 1978 is for plot 2, while the lowest is for plot 3, although the absolute value is higher at plot 3. These characteristics suggest that larger amounts of water may be lost to subsurface flow and active layer recharge at large snowbank sites, but that percentages lost by this route are lower. Presumably, the threshold ablation rate necessary to exceed subsurface flow losses is reached more frequently, or exceeded by higher margins at large snowbank locations.

In terms of the upland water balance (pp. 144-145), the data for the four runoff plots indicate that all contribute above average quantities of water to streams (surface and subsurface flow, calculated as inputs minus all evaporative losses). As expected intuitively, the largest snowbank sites produce higher proportions of total runoff (up to 96% of the original inputs) while plot 1 produces the lowest value (as low as 68%). Since the average stream runoff coefficient for the upland as a whole is between 60 and 69% (Table 10, p. 145), these figures support the hypothesis that many parts of the upland contribute little or no water to the streams and that they lose their snow cover directly to the atmosphere.

At a still larger areal scale it is instructive to compare the plot surface runoff coefficients with others obtained in the discontinuous permafrost zone (Landals and Gill, 1973). Table 14 shows that a wide range of values were obtained in both areas, but that those from Thomsen River are generally lower than those farther south, in spite of higher snowpack water equivalents. This trend is probably the result of more rapid melt rates in the Yellowknife area, which allow greater exceedance of threshold saturation values. For example, Landals and Gill (1973) recorded more than 30% of total annual surface runoff on a single day at long-lasting snowbanks. In contrast, the day with the maximum runoff volume at plot 3 in 1978, (July 1-2, see below, Table 18, p. 177) produced only 16% of the surface runoff for that year. If the reasons for these differences have been correctly identified it can be hypothesised that at latitudes greater than those of the Thomsen River area, still lower surface runoff coefficients are produced. In such areas, potential daily energy inputs are likely to be lower, the probability of exceeding the ablation rate necessary to generate surface runoff is reduced, and when the threshold is exceeded, the marginal value is less.

In summary, it appears that: (1) at interfluvial sites and small snowbanks surface flow may constitute a relatively unimportant form of water loss; (2) large snowbank sites experience greater loss by surface flow than small snowbank sites; (3) the actual configuration of the snowbank plays an important role in determining the amount of snow left on the plots when the collector position becomes snow-free; (4) increased winter snow accumulation does not necessarily correspond to greater volumes of surface flow; and (5) runoff coefficients in the study area are lower than in a subarctic area, but may be higher than in locations further north.

Plot No.	1	2	3	4	5	6	7	8
	Plot characteristics	Bare rock No.1-south facing	Bare rock No.2-north facing	Vegetated ridges 10% Jackpine cover; ground cover in other areas; some bare ground	Vegetated ridges	Muskeg depression No.5-recently burned	Muskeg depression No.6-diverse plant species cover	Mixture of regoliths - organic material, bedrock, sands, vegetation diverse
Initial pack water equivalent (mm)	46.0	79.5	79.5	74.0	79.5	61.0	66.0	94.5
Surface runoff coefficient (%)	1	82	68	72	37	56	77	50
Plot No. and year	1 1977	1 1979	2 1977	2 1978	2 1979	3 1978	4 1978	
Plot characteristics	See Table 4 and Figures 19-34							
Initial pack water equivalent (mm)	71.1	85.2	260.2	233.9	317.6	710.2	912.0	
Surface runoff coefficient (%)	9	22	26	10	2	74	33	

TABLE 14: Comparison of plot surface runoff coefficients, Yellowknife and Thomsen River, N.W.T.

### The snowmelt period

The dominance of snowmelt as the source of slopewash justifies a more detailed investigation of snow ablation processes.

#### (a) Snow metamorphism

Snow metamorphism encompasses changes in temperature, structure and density of the snowpack, both before and during the isothermal period. An understanding of these changes is important since they influence the amount of water left on the plot when the collector position is exposed, the timing and duration of surface flow, lag times between snowmelt and surface flow discharge, and many other factors. In this section, discussion is confined to (1) snow temperature changes, (2) snow density changes, and (3) the formation of ice layers.

Late-winter minimum temperatures recorded at the base of the snowpack in 1978 and 1979 were  $-15^{\circ}\text{C}$ . During these years, the changes towards isothermal conditions could be fully observed.

The time taken for snow to reach  $0^{\circ}\text{C}$  depends both on specific site factors, such as snow thickness, albedo, and surface slope, and on seasonal meteorological factors, notably incoming radiation and air temperature. Heat is transferred vertically in the snowpack by the propagation of a surface melt-wave and its refreezing at depth. Examples of some of the influences on snow warming can be seen in the progress of the snow on the upper part of site 3 towards the isothermal condition in 1978 and 1979 (Figure 58).

In Figure 58, there is a clear correlation between increases in air temperature and snow temperature, but the graphs reveal an exaggerated response in 1978 compared with 1979. For example, the change in mean daily air temperature between June 10 and June 11, 1978, from  $-3^{\circ}\text{C}$  to  $+1^{\circ}\text{C}$ , correlates with a 0.45 m penetration of the  $0^{\circ}\text{C}$  isotherm into the snow. In

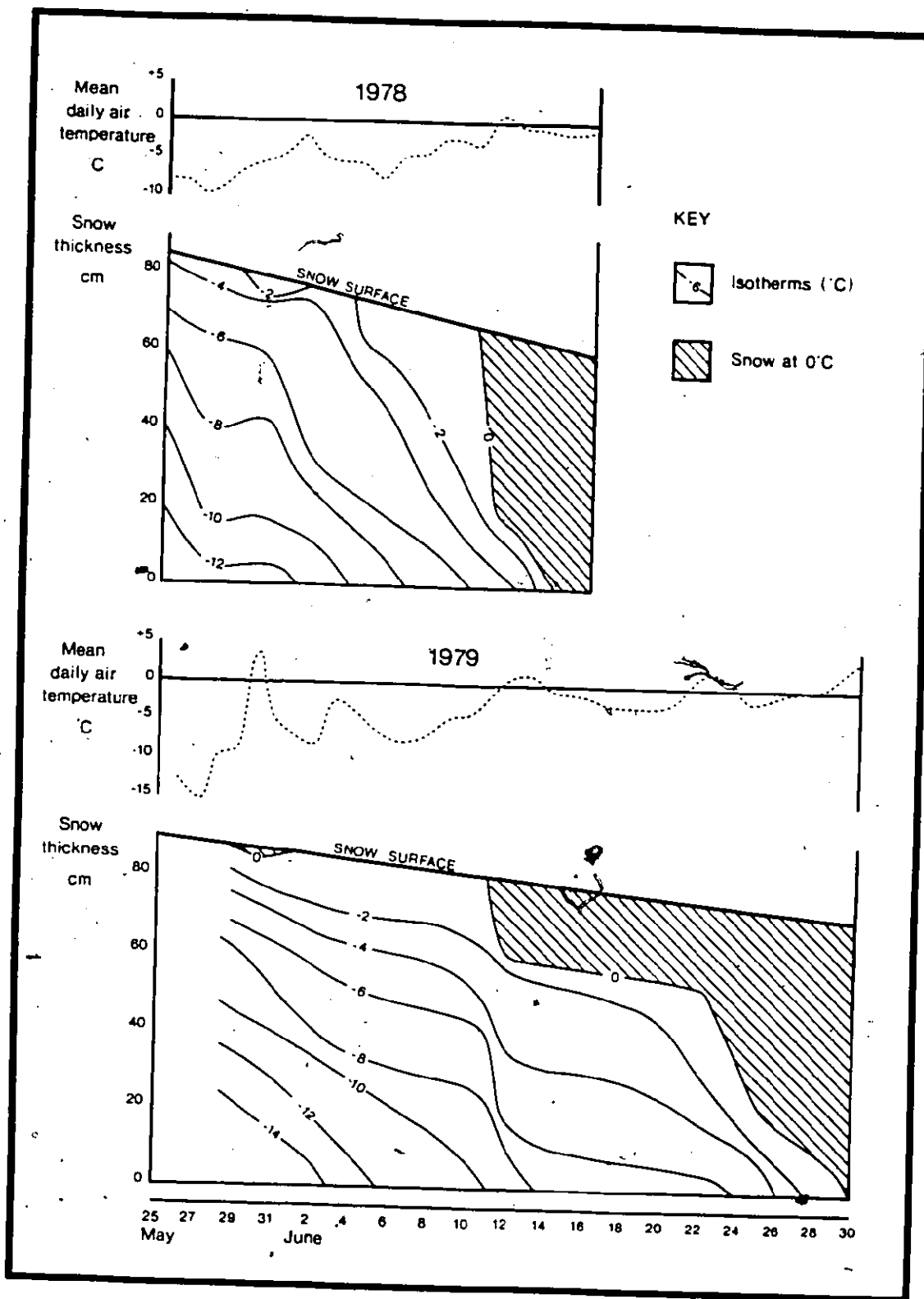


FIGURE 58 : SNOW ISOOTHERMS, UPPER PART OF SITE 3, 1978 AND 1979

contrast, a similar change between June 11 and June 12, 1979 is linked with only a 0.20 m isotherm penetration (Figure 58). The heightened response in 1978 is thought to be the result of extensive aeolian sediment concentrations on the snow surface that were not present in 1979. Presumably, the sediment lowered the albedo of the snow surface, allowing greater change in snow temperature as the response to similar energy inputs.

The unpredictability of the progress of snow warming also is illustrated by Figure 58. On May 30, 1979, all indications were that the snow was beginning a rapid warming. A change in meteorological conditions, however, caused the snow surface to refreeze and remain at temperatures below 0°C for a further 10 days.

Table 15 shows average snow temperature gradients at all four sites in 1978. Values range from 4 to 23°C m<sup>-1</sup> and are consistently greater for the shallow pits, probably as a result of lower thermal conductivities associated with lower snow densities. The changes in temperature gradients can be linked to the warming of the snow (Table 15).

As a result of thermal conductivity differences, and variable snow thicknesses and slope angles, snow temperature profiles at the runoff sites vary both within a single site and between locations (Figure 59). At site 2 on June 19, 1979, for example, the snow was isothermal to a depth of 0.2 m at the top of the slope, but to a depth of 0.4 m in the middle and lower parts of the snowbank. As discussed above (p. 148), this variability may have consequences relating to the water balance of the runoff plots.

Associated with temperature changes in the snowpack are increases in snow density. Densities measured at the runoff sites in 1979 are shown in Table 16, and similar patterns were observed in the other two years. In general, densities increase from the end of winter through to the final disappearance of the snowbank. Linear regression of snow densities at sites

DATE	SNOW TEMPERATURE GRADIENTS ( $^{\circ}\text{C m}^{-1}$ )		TEMPERATURE CHANGES
	Shallow Pits (<0.5m)	Deep Pits (>0.5m)	
May 25	23	16	Surface becomes colder or is stable; base warms
May 28	18	9	Surface cools; base stable
May 30	13	10	Surface warms faster than base
June 1	23	14	Surface colder; base warmer
June 4	19	10	Little change in gradient, but pack warms by 2-3 $^{\circ}\text{C}$
June 8	20	12	Little change in gradient, surface of snow becoming isothermal
June 11	20	9	Surface of snow isothermal; gradients decrease as base nears 0 $^{\circ}\text{C}$
June 14	6	4	

TABLE 15: Average snow temperature gradients, 1978

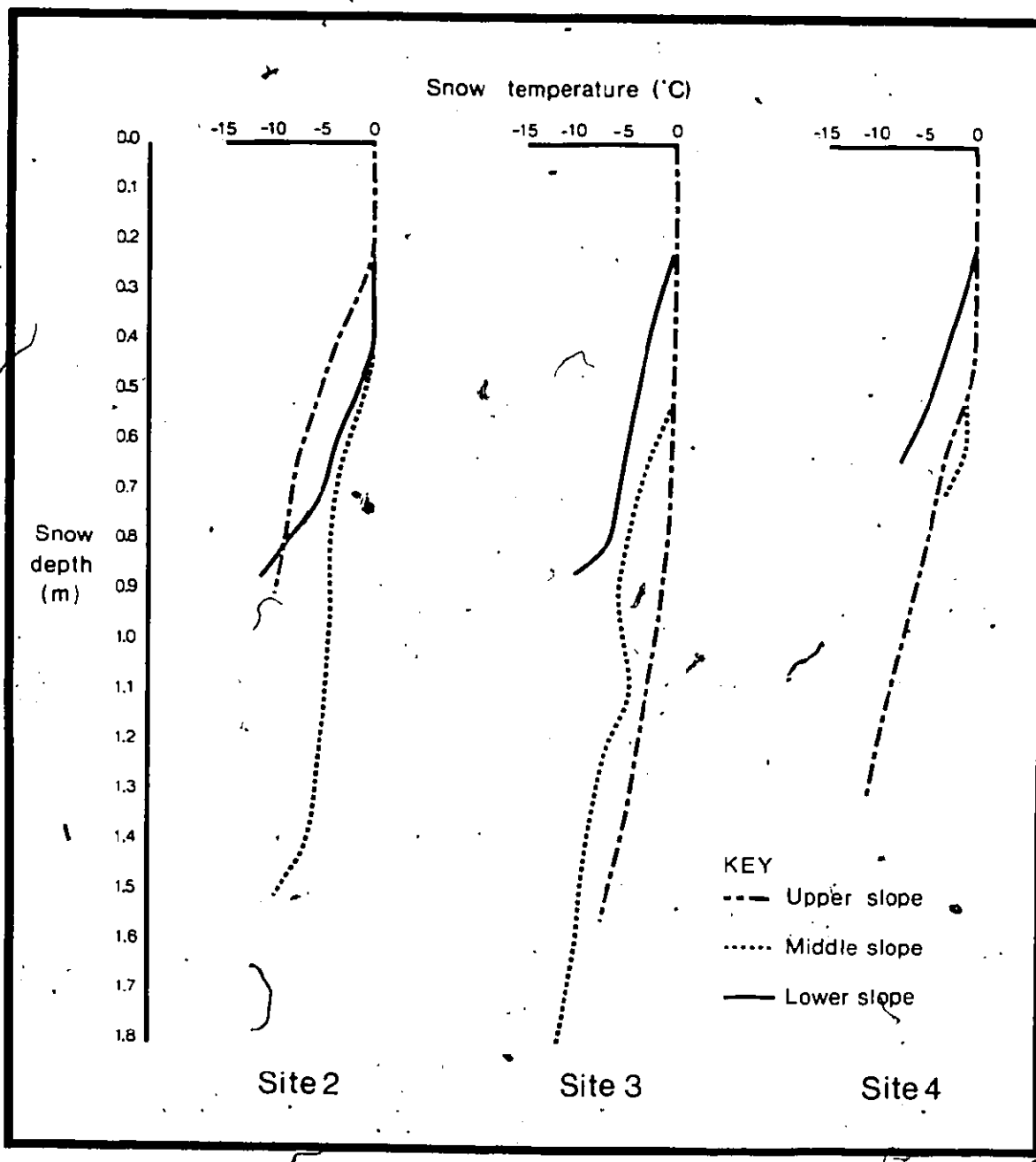


FIGURE 59 : VERTICAL SNOW TEMPERATURE PROFILES, JUNE 19, 1979

DATE	SNOW DENSITY (Mg m <sup>-3</sup> )			
	Site 1	Site 2	Site 3	Site 4
May 26			0.397 (12)	
June 1	0.359 (4)	0.406 (3)	0.374 (6)	0.400 (1)
June 6	0.336 (9)	0.367 (12)	0.440 (8)	0.433 (12)
June 17	0.347 (10)	0.431 (8)	0.447 (14)	0.432 (11)
June 26	0.348 (6)			
June 27		0.432 (6)	0.490 (8)	0.491 (9)
July 6		0.534 (12)		
July 12			0.589 (15)	
July 20				0.625 (12)
July 21			0.627 (12)	0.593 (9)
July 22			0.623 (13)	0.623 (9)
Sample size in brackets				

TABLE 16: Snow density changes, 1979

3 and 4 against time (Figure 60) shows a relationship significantly different from zero at less than the 0.1% rejection level ( $H_0$  is that  $\rho = 0$ ;  $n = 139$ ;  $r = +0.73$ ;  $t = 12.59$ ;  $t$  from tables = 1.98 so  $H_0$  rejected). Since densities were not measured at these sites late in the 1978 field season, the density - time relationship was used in ablation and water balance calculations.

In the month of May in both 1978 and 1979, snow pits at the plots revealed no visible ice layers, presumably indicating the absence of mid-winter thaws. Ice layers develop during warming, however, when the snow surface reaches  $0^\circ\text{C}$  and the underlying snow remains at lower temperatures. Under these conditions, meltwater liberated at the surface percolates downwards and refreezes. Ice layers form when the daily surface melt-wave reaches its greatest downward penetration and is in contact with the snow beneath for some hours.

Usually, ice layers at the runoff sites were only 5 - 10 mm thick and occurred as a series of bands within a single pit. The maximum thickness recorded for a single ice layer was 80 mm. The snow temperature above the lowest ice layer typically was  $0^\circ\text{C}$ , but a steep temperature gradient ( $20 - 30^\circ\text{C m}^{-1}$ ) existed immediately below the ice.

Ice layers within snow may divert meltwater, but their effects are limited since they are progressively destroyed by continuing surface melt-water production and percolation. However, a single major ice layer widely observed in long-lasting snowbanks in the Thomsen River area develops at the base of the snow and is thought significant. This layer, usually between 0.1 m and 0.2 m thick, may have upset water balance calculations, either by channeling additional meltwater towards the runoff collector or by diverting flow away from the plot (Figures 61 and 62). Presumably, the layer itself

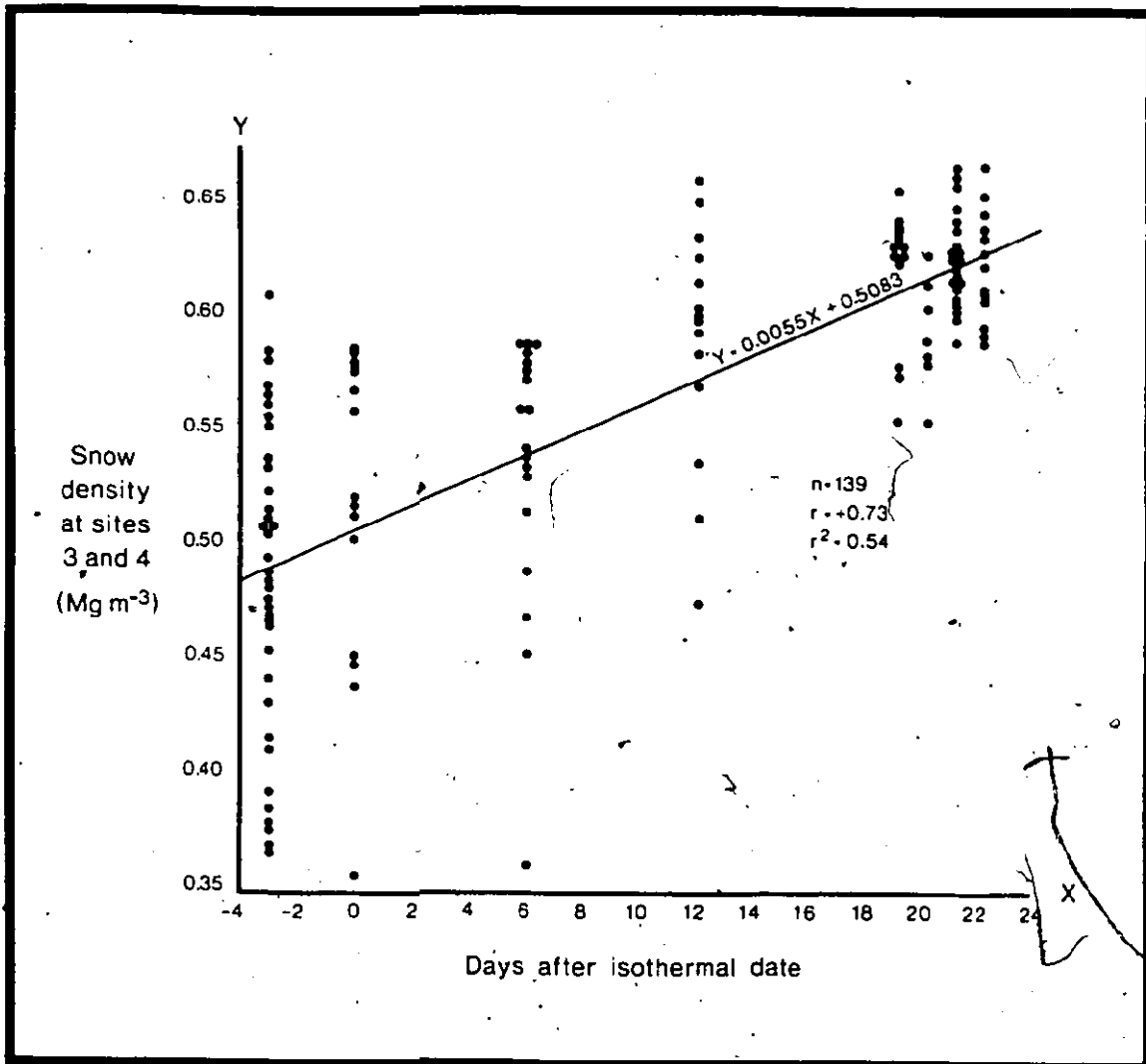


FIGURE 60 : CHANGES IN SNOW DENSITY WITH TIME,  
SITES 3 AND 4, 1978 AND 1979



FIGURE 61 : SITE 3 SNOWBANK, UPSLOPE VIEW, JULY 23, 1979  
Basal ice layer is exposed at edge of snowbank.



FIGURE 62 : SITE 3 SNOWBANK, DOWNSLOPE VIEW, JULY 23, 1979  
Basal ice layer in foreground. Note drainage over the ice and  
extensive surface flow on the slope.

forms as the result of refreezing of some part of the daily melt-wave, but whether the water is derived from vertical percolation, downslope basal flow, or both, is uncertain. The presence of the basal ice layer also has important implications in relation to the geomorphic effectiveness of slopewash.

(b) Ablation rates

Three parameters of snow ablation can be distinguished. First, surface melt ( $S_t$ ) is the mean change in elevation of the snow surface from one day to the next. It is expressed in water equivalent terms, i.e.

$$S_t = \frac{\sum_{x=1}^{x=n} (H_{x,t-1} - H_{x,t})}{n} \cdot \frac{(D_t + D_{t-1})}{2} \quad \dots(8)$$

where:

- $S_t$  is the mean surface melt rate on day  $t$  ( $\text{mm d}^{-1}$ );
- $H_{x,t}$  is the elevation of a point  $x$  on the snow surface on day  $t$  (mm);
- $n$  is the number of points sampled;
- $D_t$  is the mean snow density on day  $t$  ( $\text{Mg m}^{-3}$ ).

Equation (8) does not quantify the depth of water released from the snow, however, if  $D_t$  is much greater than  $D_{t-1}$ . In this case, an additional term ( $W_t$ ) must be introduced to account for snowpack densification.

$$W_t = S_t \cdot (D_t - D_{t-1}) \cdot H_t \quad \dots(9)$$

where:  $W_t$  is the mean depth of water released from the snowbank on day  $t$  ( $\text{mm d}^{-1}$ );  
 $H_t$  is the mean remaining snow thickness on day  $t$  (mm);  
 all other terms are as defined in equation (8).

The third snow ablation parameter is the actual volume of water released ( $Q_t$ ) at a runoff plot on a particular day:

$$Q_t = W_t (A_t + A_{t-1}) \times 0.5 \times 10^{-3} \quad \dots(10)$$

where:  $Q_t$  is the volume of water released from the snow, available for evaporation or runoff ( $\text{m}^3 \text{d}^{-1}$ );  
 $A_t$  is the area of snow on the plot on day  $t$  ( $\text{m}^2$ );  
 $W_t$  is as defined in equation (11) ( $\text{mm d}^{-1}$ ).

The first parameter,  $S_t$ , reflects energy exchange and not internal snowpack processes. Rates for 1978 and 1979 vary from 4.2  $\text{mm d}^{-1}$  to 161.3  $\text{mm d}^{-1}$  (Figure 63). In 1978, surface melt rates paralleled the meteorological changes without obvious lags at a daily time scale. Peak rates at sites 2, 3 and 4 occurred between July 1 and July 4, several days prior to a marked air temperature increase. Since incoming radiation values were similar during both July 1-4 and July 6-7, the later, lesser rates may reflect lower wind velocities averaging 8.4  $\text{m s}^{-1}$  on July 2-3 and only 2.8  $\text{m s}^{-1}$  on July 6-7.

In 1979, maximum recorded surface melt rates were not as great as in 1978, and a single period of high values is not apparent (Figure 63).

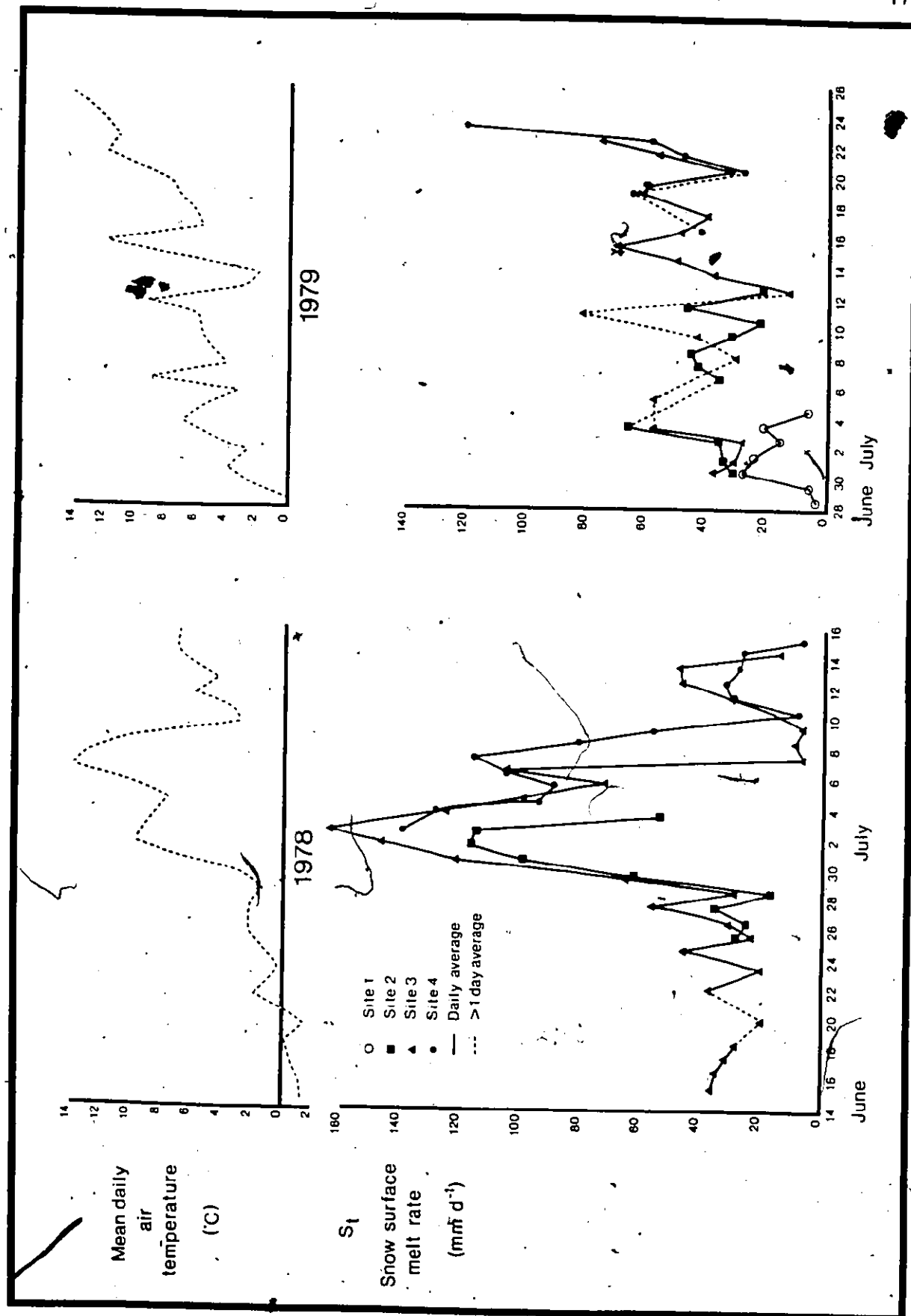


FIGURE 63 : SNOW SURFACE MELT RATES, 1978 AND 1979

Values equal to the 1978 maxima may have recurred after the end of the ablation records since air temperatures reached a three-season maximum of  $19.6^{\circ}\text{C}$  on July 25, 1979. Later maxima are not considered likely, however, for a number of reasons. First, due to reduced solar declination, radiation is not as effective in inducing melt in late July as it is earlier in the month. Second, as a result of the presence of sediment at the surface, the snow albedos at sites 3 and 4 were lower in 1978 than in 1979. Third, the highest air temperatures in 1979 were accompanied by near-calm wind conditions, which reduced the potential effects of warm air in turbulent energy exchange. It is hypothesised, therefore, that the period of cloudless and warm weather necessary to produce high melt rates, occurred too late in 1979 to induce rates matching those of 1978.

The rates of melt recorded at the runoff sites on any particular date are similar but not equal (Figure 63). In 1978, surface melt at site 2 on most days proceeded more slowly than that at site 3; site 4 experienced greater values. The record is less complete for 1979 and the only obvious generalisation is that melt at site 1 was slower than elsewhere. Inter-site differences likely are caused by variations in slope aspect, snow surface angle, and snow albedo.

Using the threshold concept of surface flow production, the most important ablation events are the annual maxima. Table 17 shows that  $W_{\text{max}}$  (the maximum depth of water released) always occurred on the same day as  $S_{\text{max}}$  (maximum surface melt).  $S_{\text{max}}$ , however, did not necessarily coincide with  $Q_{\text{max}}$  (maximum water produced by the snowbank), because the snowbank area had become depleted. Examples of the two possibilities of non-coincidence and coincidence occurred at plot 2, in 1978 and 1979 respectively (figure 64). Simultaneity, or its absence among

LOCATION AND YEAR		MAXIMUM SURFACE MELT RATE ( $S_{max}$ )		MAXIMUM WATER RELEASE RATE ( $W_{max}$ )		MAXIMUM VOLUME OF WATER RELEASED ( $Q_{max}$ )		COINCIDENCE
		Dates	mm d <sup>-1</sup>	Dates	mm d <sup>-1</sup>	Dates	m <sup>3</sup> d <sup>-1</sup>	
Plot 1	1977	3/6-4/6	53.2	3/6-4/6	53.2	30/5-31/5	3.21	No
	1979	30/6-1/7	28.2	30/6-1/7	28.2	30/6-1/7	9.17	Yes
Plot 2	1977	13/6-14/6	70.5	13/6-14/6	70.5	11/6-12/6	19.14	No
	1978	1/7-2/7	117.3	1/7-2/7	117.3	30/6-1/7	26.53	No
	1979	3/7-4/7	65.9	3/7-4/7	62.4	3/7-4/7	23.65	Yes
Plot 3	1978	2/7-3/7	131.0	2/7-3/7	128.8	30/6-1/7	21.09	No
	1979	11/7-12/7	80.2	11/7-12/7	73.6	11/7-12/7	12.90	Yes
Plot 4	1978	3/7-4/7	161.3	3/7-4/7	158.4	3/7-4/7	8.40	Yes
	1979	23/7-24/7	121.1	23/7-24/7	119.0	23/7-24/7	4.97	Yes

TABLE 17: Maximum daily ablation parameters, 1977-1979

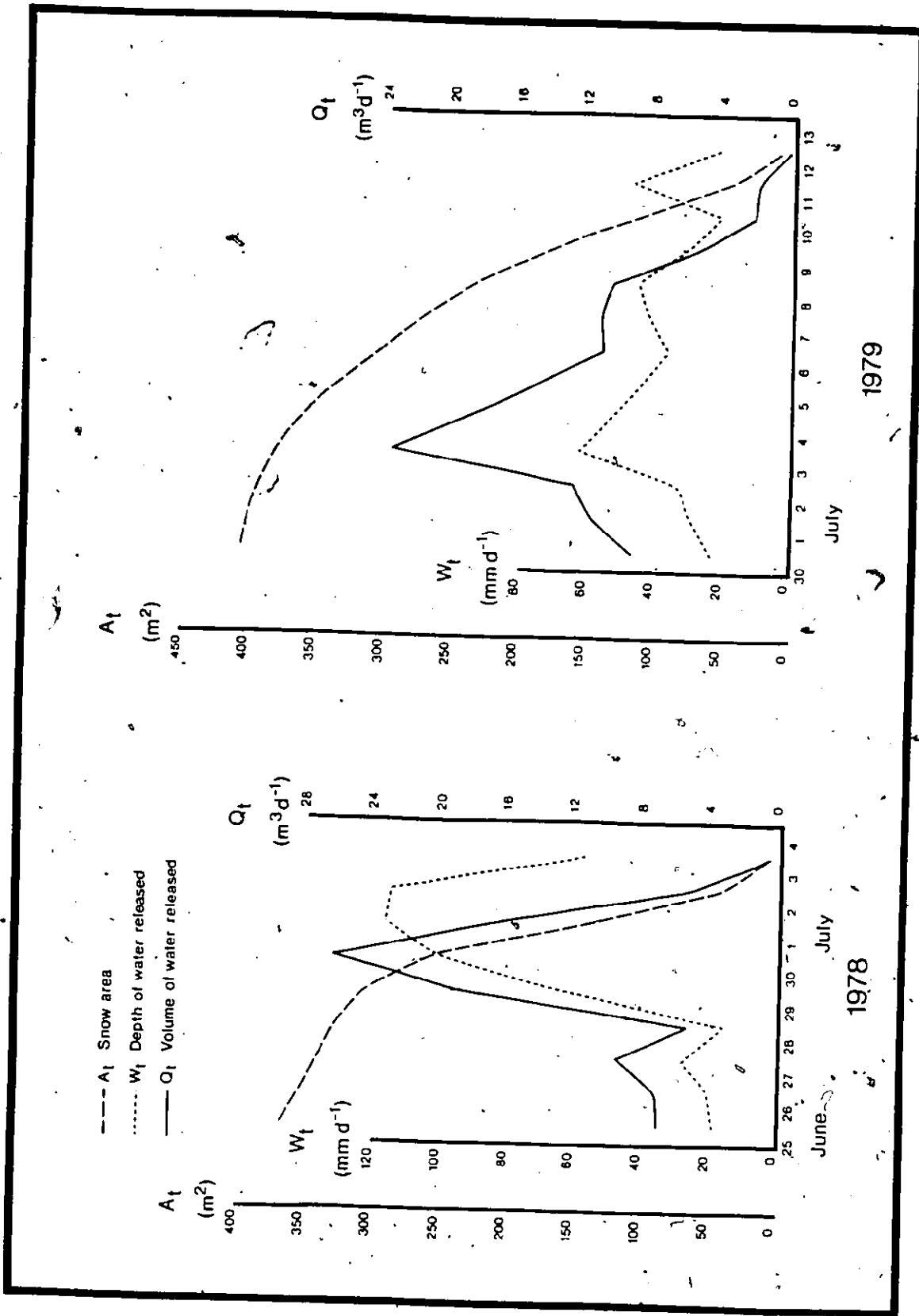


FIGURE 64 : SNOW ABLATION PARAMETERS, PLOT 2, 1978 AND 1979

the three ablation parameters, is largely a matter of chance, depending on thaw season weather conditions. If high radiation and turbulent energy inputs are received soon after the snow becomes isothermal, peak melt rates and water production may coincide. If a greater amount of time elapses, say 10 - 15 days, coincidence remains possible, but is less likely since area reduction continues even at slow rates of surface melt.

(c) Surface runoff coefficients

Surface runoff coefficients are calculated by comparing the volume of surface flow leaving the runoff plot with the water production of the snowbank on it. These two sets of data are not temporally in phase, however, and a correction factor must be applied.

The lag time between melt and surface flow is the time taken for meltwater to percolate vertically through the snowpack and then move downslope to the collector. The optimum lag for a particular day's flow can be assessed by regressing surface flow discharge rates against a lagged index of energy inputs (e.g. net radiation) and maximising the value of the correlation coefficient. The value obtained using this method is a compromise, however, because the lag varies during the day (see pp. 77 - 78), being greatest at low rates of melt and least at the time of peak melt. At plot 3 on June 25, 1978, for example, these lag times were 8 hours and 2.75 hours respectively. The simplest solution is to make an estimate of the lag for each particular ablation measurement, by visually comparing graphs of recorded net radiation and surface flow. On days with varying inputs of radiation resulting from changing cloud cover, the rises and falls of energy inputs are assumed to occur simultaneously at the runoff plots. On cloudless days, however, site topography and orientation become

important, and the times of rise and peak are assessed using output values from a computer model (Fuggle, 1970; Garnier and Ohmura, 1968).

The periods for which surface runoff coefficients can be calculated are determined by the frequency of ablation data collection (usually daily), since a continuous record of surface flow exists. Surface runoff coefficients calculated from the 1978 field data are reasonably typical and are presented in Table 18. Values range from 0 - 100%, and some apparently exceed the latter figure. With few exceptions, the higher coefficients occur on days with higher meltwater production. This supports the concept that surface flow generation occurs only after certain threshold values of snowbank water production are exceeded (see p. 157).

Further information concerning the nature of surface runoff generation is obtained by regressing daily surface runoff against meltwater production at the plots (Figure 65). There is a strong relationship between the two, with correlation coefficients ranging from +0.747 (plot 2) to +0.942 (plot 1) (Table 19). T-tests reject the null hypothesis that these values are not significantly different from zero at less than the 0.1% level (Table 19). The regression lines apparently explain between 55.8% and 88.8% of the variance of the dependent variable. These may be inflated values, however, since calculation of the Durbin-Watson 'd' statistic (Durbin and Watson, 1950; 1951) results in a rejection of the null hypothesis of no positive serial autocorrelation in the data (i.e.  $d < D_L$  (tables)) to less than the 1% significance level (Table 19). For this reason, no further inferential tests on the values of the slope and intercept coefficients in the regression equation were undertaken.

Dates	PLOT 2			PLOT 3			PLOT 4		
	Melt-water production ( $m^3 d^{-1}$ )	Surface runoff ( $m^3 d^{-1}$ )	Surface runoff coefficient (%)	Melt-water production ( $m^3 d^{-1}$ )	Surface runoff ( $m^3 d^{-1}$ )	Surface runoff coefficient (%)	Melt-water production ( $m^3 d^{-1}$ )	Surface runoff ( $m^3 d^{-1}$ )	Surface runoff coefficient (%)
June 15-June 16				7.14	0.27	3.8			
June 16-June 17				6.97	0.29	4.2			
June 17-June 18				5.61	0.33	5.9			
June 18-June 19				5.17	1.33	25.7			
June 19-June 21				3.21	1.31	40.6			
June 21-June 23				7.14	2.62	36.7			
June 23-June 24				2.89	1.16	40.0			
June 24-June 25				9.18	3.79	41.3			
June 25-June 26	8.14	0.00	0.0	3.76	1.60	42.5			
June 26-June 27	7.07	0.00	0.0	6.11	1.49	24.4			
June 27-June 28	9.83	0.19	2.0	10.77	5.94	55.2			
June 28-June 29	5.75	0.01	0.2	4.74	3.82	80.5			
June 29-June 30	19.23	3.08	16.0	11.16	8.30	74.4			
June 30-July 1	25.97	5.44	20.9	20.26	19.10	94.2			
July 1 -July 2	17.46	2.92	16.7	19.91	21.58	108.4			
July 2 -July 3	5.28	0.11	2.0	17.03	15.99	93.9			
July 3 -July 4	0.58	0.00	0.0	11.72	11.49	98.1	6.57	6.39	97.3
July 4 -July 5				8.34	8.21	98.3	4.53	3.87	85.4
July 5 -July 6				5.53	8.34	150.8	3.79	3.18	84.0
July 6 -July 7				6.97	6.80	97.6	3.95	1.72	43.5
July 7 -July 8				0.33	5.39	1648.1	3.96	0.24	6.1
July 8 -July 9				0.43	2.13	490.5	2.50	0.40	15.9
July 9 -July 10				0.21	1.28	605.2	1.49	0.00	0.0
July 10-July 11				0.50	1.39	281.2	0.19	0.07	38.4
July 11-July 12				0.66	0.68	102.1	0.70	0.04	6.2
July 12-July 13				0.73	0.12	16.4	0.70	0.0	0.0
July 13-July 14				0.33	0.01	1.6	0.53	0.0	0.0
July 14-July 15				0.01	0.00	0.0	0.33	0.0	0.0
July 15-July 16				0.01	0.00	0.0	0.03	0.0	0.0

TABLE 18: Surface runoff coefficients, plots 2-4, 1978

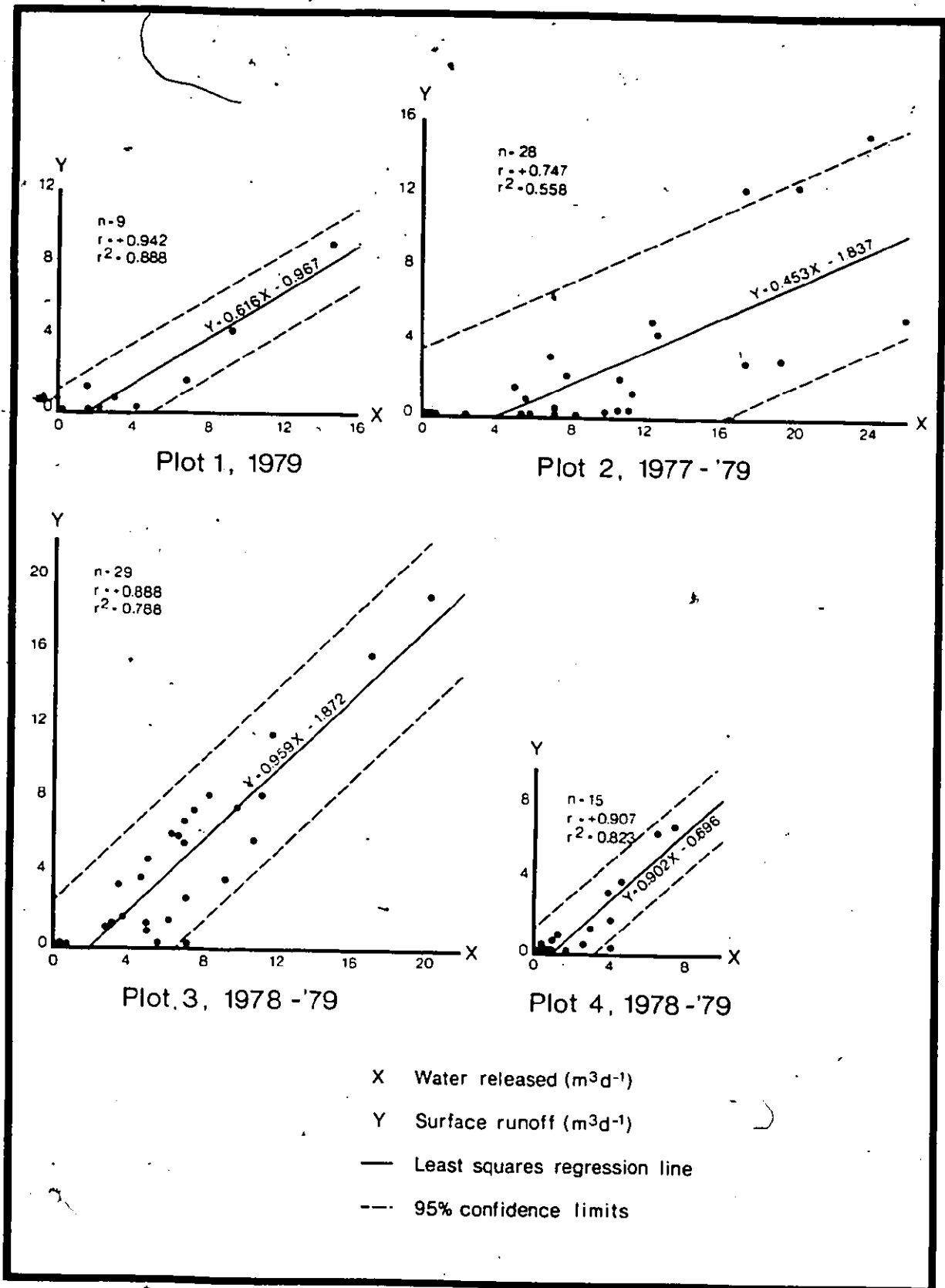


FIGURE 65 REGRESSION OF DAILY SURFACE RUNOFF VS. DAILY MELT WATER PRODUCTION, PLOTS 1 - 4  
 Note: runoff coefficients greater than 100% not included.

	Number of Points (N)	Correlation coefficient (R)	R <sup>2</sup>	Regression equation <sup>a</sup>	Intercept on X-axis	T <sub>Calc</sub>	T <sub>Tables</sub> (0.1% level)	Durbin Watson Statistic (d)	D <sub>Tables</sub> (Tables Level)
Plot 1	9	+0.942	0.888	Y = 0.616X -0.967	1.57	11.27	5.41	N.D. <sup>b</sup>	N.D.
Plot 2	28	+0.747	0.558	Y = 0.453X -1.837	4.06	5.73	3.75	0.64	1.10
Plot 3	29	+0.888	0.788	Y = 0.959X -1.872	1.95	10.02	3.65	0.80	1.12
Plot 4	15	+0.907	0.823	Y = 0.902X -0.696	0.77	7.78	4.32	0.77	0.81

<sup>a</sup> X = Snowbank meltwater production (m<sup>3</sup>d<sup>-1</sup>)  
<sup>b</sup> Y = Surface runoff (m<sup>3</sup>d<sup>-1</sup>)  
<sup>b</sup> = N < 15

TABLE 19: Regression parameters for daily surface runoff vs. daily meltwater production

Regression line intercepts on the x-axes (Table 19) are the average levels of daily meltwater production required to generate surface flow. These "threshold" values range from  $0.77 \text{ m}^3$  (Plot 4) to  $4.06 \text{ m}^3$  (Plot 2). The gradients of the regression lines represent the average surface runoff coefficients for meltwater production in excess of the threshold values. Those for plots 3 and 4 are close to unity, while those for plots 1 and 2 are considerably lower (Table 19). The former suggest that once the threshold values are exceeded, increased water loss to subsurface flow and evapotranspiration is slight.

While the regression lines describe the major trends, a number of other factors may influence the relationships since a considerable portion of the variance in the data remains unexplained.

First, some of the variance in the regression residuals is probably the result of the time-scale enforced by the ablation measurements. The relationship between snowmelt and surface flow can change over the course of a few hours in response to changing energy inputs, but the time-scale used is a daily one. An hypothetical example illustrates the point. A day with continual melt at relatively low rates contrasts with another day in which very low melt rates are interrupted by a period of rapid meltwater release. The meltwater production totals are similar in both cases (i.e. x-axis values constant), but in the first, subsurface flow and evapotranspiration remove all water produced, while in the second some surface flow is generated (i.e. y-axis values vary).

Second, measurement errors may introduce variation about the regression line. Coefficients greater than 100% were not included in the statistical analysis but provide some clues as to the potential sources of errors in the data that were used. The runoff coefficients which exceed

100% at plot 3 in 1978 (Table 18) may be separated by the July 7 date. The existence of the earlier group may relate to inaccurate assessment of lag times, and long snow storage recessions, since the overall runoff coefficient for the period, June 27 - July 7, is less than 100%. Incorrect allocation of runoff cannot explain the anomalies recorded after July 7 since total calculated surface runoff exceeds total meltwater production. Here, the source of error probably is the presence of the basal ice layer in the snowbank which made snow depth measurements increasingly difficult during the final stages of ablation.

Third, the presence of serial correlation in the regression residuals suggests other factors at work, namely antecedent moisture conditions, and lowering of the frost-table. These are discussed below.

During the periods of peak snowmelt, surface flow typically becomes continuous at the plots. Thus, active layer moisture deficits do not develop in areas downslope of the snowbanks. At other times, however, and particularly during the first few days of surface flow generation, hydrographs are discontinuous and moisture deficits must be satisfied each day before surface flow commences. This mechanism is thought to be responsible, for the low surface runoff coefficients between June 15 - June 18, 1978 at plot 3 (Table 18). To test this hypothesis, daily surface can be regressed against both meltwater production on the same day, and an index of the four previous days' production in the form of:

$$AMI = \sum_{t=1}^{t=4} Q_t(0.5)^t \quad \dots(11)$$

where: AMI is an antecedent moisture index;  
 $Q_t$  is the volume of meltwater produced by the snow on day  $t$  ( $m^3$ );  
 $t=1$  is the previous day, and  $t=4$  is four days previous.

The introduction of a second independent variable increases the correlation coefficient values obtained at all plots relative to those from simple linear regression (Table 20). The correlation is regarded as spurious at plots 1 and 4, however, because the slope coefficients for the moisture index are negative, probably indicating a partial correlation of the AMI with other factors. At plots 2 and 3, on the other hand, the slope coefficients are positive as expected, and in view of the larger sample sizes, they are more likely to be correct.

Lowering of the frost-table allows an increase in the rate of subsurface discharge and should be negatively correlated with surface runoff production. Regression analysis, after the introduction of frost-table depth at the surface flow collector position as an additional variable, generally fails to support this hypothesis (Table 21). At plots 1-3, the slope coefficients for the variable are positive. This is thought to be the result of partial correlation with other time-dependent variables. One of these, especially at plots 2 and 3, may be antecedent moisture. At plot 4, subsurface losses appear to be more important and work against the trend of the AMI. The thickening of the unfrozen soil layer from 0.10 - 0.25 m between July 5-6 and July 7-8, 1978, for example, apparently was responsible for reducing runoff coefficients for similar meltwater production from 84% to 6% (see Table 18, p. 177). As a result, the expected trend is illus-

	Regression equation <sup>a</sup>	N	R	R <sup>2</sup>
Plot 1	$Z = 0.613X - 0.045Y - 0.820$	9	+0.945	0.894
Plot 2	$Z = 0.379X + 0.224Y - 3.083$	28	+0.791	0.626
Plot 3	$Z = 0.793X + 0.205Y - 2.151$	29	+0.904	0.816
Plot 4	$Z = 1.101X - 0.331Y - 0.377$	15	+0.921	0.849
<sup>a</sup> X = Meltwater production ( $m^3 d^{-1}$ ) Y = AMI (see equation 11) ( $m^3$ ) Z = Surface runoff ( $m^3 d^{-1}$ )				

TABLE 20: Multiple regression equations: daily surface runoff vs. daily meltwater production and antecedent moisture index

	Regression equation <sup>a</sup>	N	R	R <sup>2</sup>
Plot 1	$Z = 0.631X + 10.337Y - 4.239$	9	+0.949	0.900
Plot 2	$Z = 0.455X + 13.998Y - 4.689$	28	+0.774	0.599
Plot 3	$Z = 0.907X + 4.734Y - 3.783$	29	+0.910	0.829
Plot 4	$Z = 0.841X - 1.974Y - 0.068$	15	+0.910	0.829

<sup>a</sup> X = Meltwater production ( $m^3 d^{-1}$ )  
Y = Frost-table depth (m)  
Z = Surface runoff ( $m^3 d^{-1}$ )

TABLE 21: Multiple regression equations: daily surface runoff vs. daily meltwater production and frost-table depth at the surface flow collector position

trated at this plot, and the frost-table depth is negatively correlated with surface runoff.

In all cases, the data used to develop the equations in Tables 20 and 21 show positive serial autocorrelation. The introduction of additional variables, therefore, failed to eliminate the trends in the residuals and no further inferential tests on the regression coefficients were undertaken.

In summary, a strong positive relationship exists between meltwater production and surface runoff production at the four runoff plots. At plots 3 and 4, once threshold values of water production are reached, almost all excess water leaves the plot as surface flow. At plots 1 and 2, even after the threshold necessary to generate surface flow is exceeded, evaporation and subsurface losses continue to increase. Antecedent meltwater production has some influence over the threshold value at plots 2 and 3. At plot 4, the level of the frost-table appears important through its effect on subsurface flow rates.

(d) Generation of subsurface flow

At daily time scale, subsurface losses at the runoff plots are best examined as residuals of the water-balance equation. These can be converted into discharge per unit width of slope to aid inter-site comparison (Figure 66). A number of pertinent points emerge.

First, the relative importance of subsurface flow in the water balance at each plot varies on a daily basis, since it may constitute virtually all of the water loss (e.g. plot 2, June 25 - 26, 1978) or only a very small percentage (e.g. plot 4, July 3 - 4, 1978). This variation is to be expected as the corollary of the processes of surface flow production. Second, the relative importance of subsurface flow varies from site to site, mainly as a result of the variation in snowbank dimensions. For example,

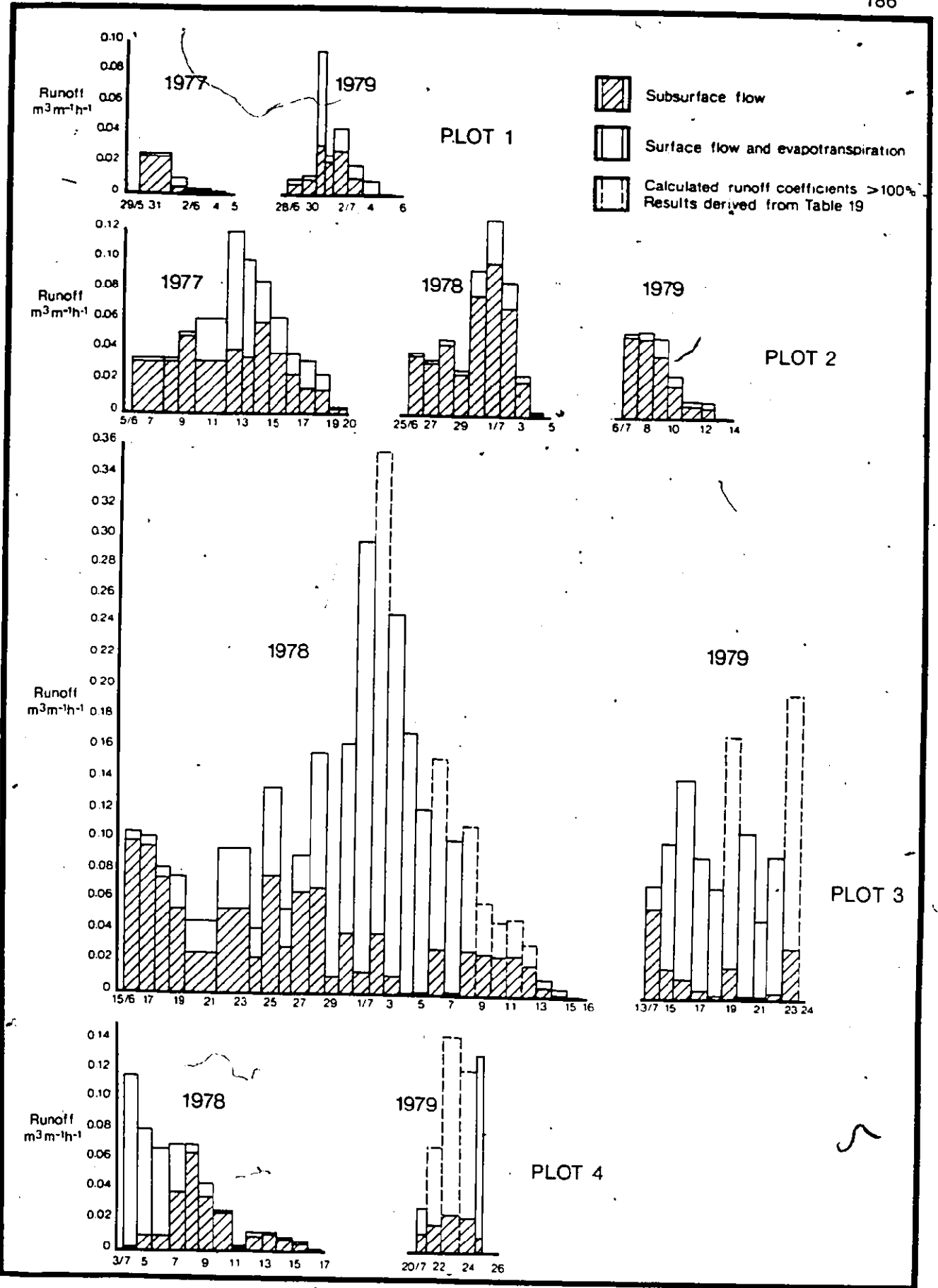


FIGURE 66 : MELT WATER PRODUCTION AND SUBSURFACE FLOW RATES, PLOTS 1 - 4, 1977-1979

the size of the snowbank dimension orthogonal to the slope contours at plot 3 results in a very high production of meltwater per unit width of slope. The regolith cannot absorb this water supply, and most leaves the plot as surface flow with a consequent diminution in the importance of subsurface flow. Third, as discussed above, effects of frost-table lowering during the ablation season are identifiable at plot 4 but cannot be distinguished at the other plots. Fourth, differences between subsurface flow production at the same plot in different years also indicate the potential importance of the position of the frost-table, itself influenced by the initial distribution of snow on the plot. For example, at plot 2, similar seasonal maximum values of meltwater production, on June 11-12, 1977 and June 30 - July 1, 1978 ( $0.119 \text{ m}^3 \text{ m}^{-1} \text{ h}^{-1}$  and  $0.129 \text{ m}^3 \text{ m}^{-1} \text{ h}^{-1}$  respectively), produced differences of 2.4 times in subsurface flow rates in spite of similar antecedent moisture conditions (Figure 66). In 1977, the thaw depth at the collector was 0.11 m, while in 1978 it was 0.23 m. Presumably, the extra depth allowed more water to leave the plot as subsurface flow.

Data from the subsurface flow collectors are interesting adjuncts to the water balance residuals, but for the most part are less useful since monitoring was discontinuous. Two periods of 24-hour monitoring, July 4 - 5 and July 8 - 9, 1978, were undertaken at plot 4 to record diurnal variations. At this site, the subsurface flow collector is situated beneath the surface flow collector, so that after calculation of the upslope contributory width, direct comparisons of the surface and subsurface unit slope discharges can be made.

The residuals from the water balance equation for the two days in question (see Figure 66) are much lower than the total flow collected. Calculations show the collected subsurface flow to be an overestimate of the slope discharge by a factor of 3.17 on July 4 - 5 and 4.20 on July 8 - 9. These figures suggest an upslope contributory width for the subsurface flow collector of approximately 3 - 4 times its actual width (i.e. 1.5 - 2.0 m), either due to the alteration of lines of hydraulic potential in the neighbourhood of the collector, or due to concentration of flow by the configuration of the frost-table. The flow rates shown in Figure 67 have been corrected by the scaling factor, so that realistic comparisons of surface and subsurface discharges can be made.

July 4 - 5, 1978 was the second day of monitoring at plot 4 and the unfrozen layer at the position of the collectors was less than 0.10 m thick. The topmost gutter thawed and began functioning during the period, but the other three remained below the frost-table. The subsurface flow is of minor importance compared with surface flow, shows little variation and no diurnal regime (figure 67). In contrast, by July 8 - 9, the snowbank was much depleted and unable to maintain continuous surface flow. The second gutter at a depth of 0.16 m had thawed by this time, and was collecting much greater discharges than the upper one. These observations suggest increased substrate permeability at a depth of 0.12 - 0.18 m, a view supported by the negative slope coefficient in the multiple regression (Table 21, p. 184). The decline in surface flow from 17.00 h on July 8 (Figure 67), is paralleled by a more gradual recession in subsurface flow at the lower level. If a lag is present, it cannot be detected because of the interval of the subsurface flow measurements. Although at very low rates, the flow in the upper gutter also shows a decline during the hours of slow snowmelt.



Both levels of subsurface flow show rises the following morning, but the rise at the 0.16 m depth precedes the other by 1.0 - 1.5 hours. This probably indicates that the active layer is saturated from the frost-table upwards.

These two days of measurements, if typical, suggest that when surface flow is continuous, subsurface flow shows little or no diurnal variation. Furthermore, discontinuous surface flow indicates desaturation of the active layer, and a decline in subsurface discharges. It is evident that during days of rapid snowmelt, surface flow is much more important than subsurface flow. At lower rates, and with a deepening frost-table, the subsurface flow rates increase and may dominate the daily hydrological regime.

Information about subsurface flow rates following the final ablation of the snowbanks is available only from the subsurface flow collectors. Continued movement is expected from (1) water left in the thawed layer moving downslope after snowmelt is completed, (2) water from precipitation events, and (3) water from ground ice thaw as the frost-table falls.

The decline in volumes of subsurface flow collected at plot 4 in 1978 is very rapid, while that at site 3 is more gradual (Figure 68). At both sites, cessation of flow is progressive, from the topmost gutter downwards. Flow continues at site 3 after it ceases at site 4, either because the collector is located at the base of a longer slope, or as a result of lower permeability values.

Revitalisation of subsurface flow follows precipitation events and subsequent recession is again faster at site 4. If the contributory widths of slope for the collectors are the same as during the ablation period, it is evident that the subsurface flow afterwards is much less important than

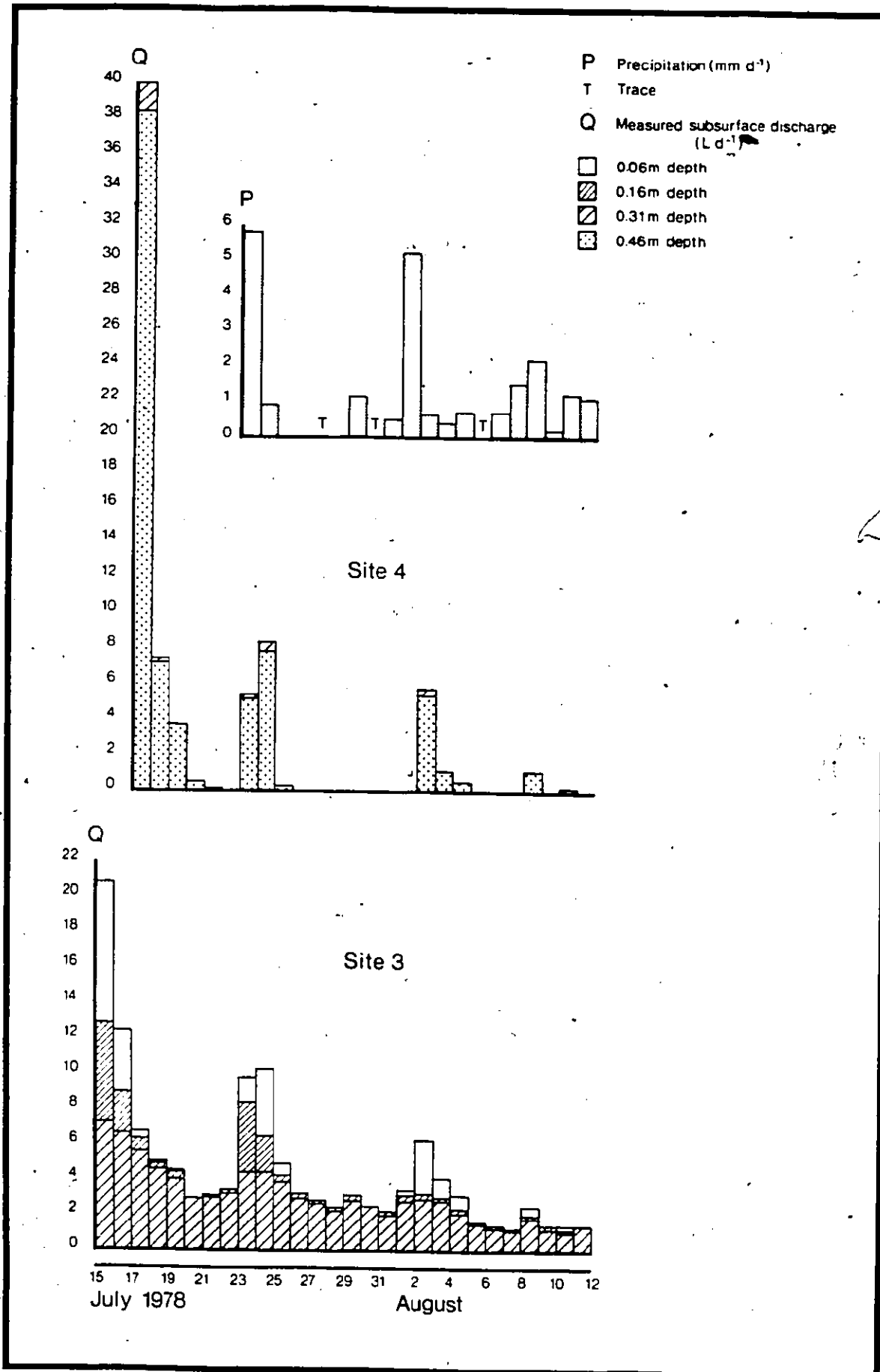


FIGURE 68 : DECLINE OF SUBSURFACE FLOW AFTER THE CESSATION OF SNOWMELT; SITES 3 AND 4, 1978

during snowmelt. Total flow at all gutter levels after the end of snowbank ablation was  $0.008 \text{ m}^3$  and  $0.132 \text{ m}^3$  at site 3 from June 28 - July 27, 1977 and July 15 - August 16, 1978 respectively, and was  $0.075 \text{ m}^3$  at site 4 from July 17 - August 16, 1978. These figures can be compared with, for example,  $0.188 \text{ m}^3$  collected at site 3 and  $1.980 \text{ m}^3$  at site 4 in 24 hours during the snowmelt period (July 8 - 9, 1978).

The importance of active layer thaw to subsurface flow can be estimated using a computer program to calculate equation (1) (p. 67). For example, for site 3, using a value for available water (i.e. water lost at suctions of less than 1.5 MPa) of 0.25 (an average for silty-clay subsoils (Hall et al., 1977)), and the recorded retreat of the snowbank for 1978, maximum ground ice thaw occurred on June 15 (Figure 69). After the end of snowmelt, the maximum quantity of ice thawed as available water was  $0.32 \text{ m}^3 \text{ m}^{-1} \text{ d}^{-1}$ , but only one-third of this is likely to have flowed downslope rapidly (i.e. was held above field capacity). Even so, this figure is considerably greater than that recorded at the 0.31 m depth at site 3, and it appears quite feasible, therefore, that the low sustained rates at this level were caused by water release following ground ice thaw.

#### (e) Daily cycles of snowmelt runoff

Eighty-five daily hydrographs of surface flow were recorded at the runoff plots during the three field seasons. No two are identical, but a number of recurrent features exist.

The typical plot surface flow hydrograph consists of (1) a steeply sloping rising limb, (2) a peak, and (3) a more gently sloping falling limb. On a day with cloudless skies, the limbs are relatively smooth curves, but the peak is sharp (e.g. July 1 and 2, Figure 70). Similar smooth curves are extant on a day with overcast skies (e.g. June 29, Figure 70) but snowmelt

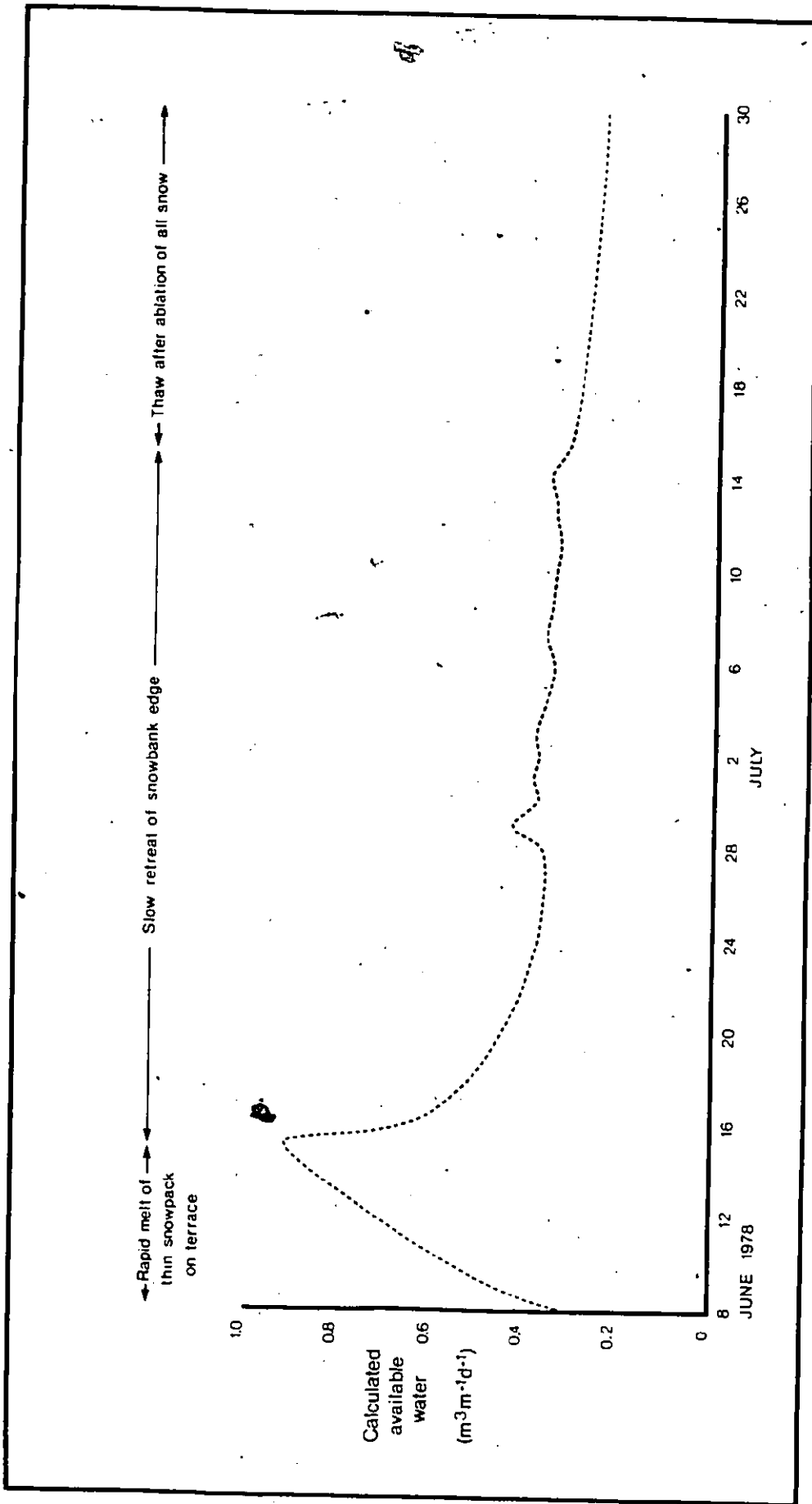


FIGURE 69 : WATER AVAILABLE FROM GROUND ICE THAW, SITE 3, 1978  
 Values derived from equation (1).



rates are lower and the volume of surface flow produced is smaller. Moreover, the maximum surface flow rate occurs several hours later in the day and the hydrograph is less sharply peaked. On intermittently cloudy days, changes in energy inputs are reflected in a variable hydrograph shape. The double peak on June 30, 1978, for example (Figure 70), represents the response to a double peak in net radiation produced by the passage of a bank of altostratus clouds. Depending on the amount of snow left on the plot and the rate of melt, hydrographs are continuous or discontinuous. Discontinuous flow prevailed (1) at the interfluve site where the snow was thin, and (2) at the snowbank sites at the beginning of the ablation season when melt rates were low, and at the end, when snow thicknesses were much reduced. Continuous flow was usual at the snowbank locations during the middle part of the ablation season when semilogarithmic recession curves extended into the following day (Figure 70). All these observations are consistent with snowmelt hydrological theory (see pp. 77 - 79).

The initial rise in a surface flow hydrograph represents the arrival of the first melt flux or shock front of the day, usually derived from the area of snow closest to the downslope snow edge. For a homogeneous snowpack of uniform thickness and slope, theory predicts a rapid increase in the rate of water leaving the pack, as shock fronts which have developed in the unsaturated zones of more distant areas emerge from the basal saturated layer. The snowpack at plot 3 is not of uniform thickness or slope, however, and while this does not preclude the occurrence of a single diurnal peak, it is thought to affect the shape of the hydrograph.

Melt fluxes simultaneously leaving the surface of a pack of non-uniform thickness do not arrive at the downslope edge after lag times proportional to their upslope distance, because horizontal velocities are up

to two orders of magnitude greater than vertical ones. This was demonstrated during an experiment with rhodamine dye at site 3 on July 24, 1979. Dye was placed on the melting snow at five points across the snowbank, each a measured horizontal distance from the bottom snow edge and each on top of a measured snow thickness (Figure 71). The first dye trace to emerge at the downslope snow edge came from the top of the snowbank where the dye had been positioned directly on the exposed basal ice layer (Table 22). Multiple regression of time to emerge against vertical and horizontal distance travelled, shows that the vertical velocity of the melt flux averaged 32 times less than the horizontal velocity ( $0.19 \text{ mm s}^{-1}$  and  $6.00 \text{ mm s}^{-1}$  respectively). Variance about the regression line (which explained 80.2% of the data scatter) probably is the result of snow crystal heterogeneity and varying slope angles.

The dye experiment shows that the first shock-fronts to arrive at the runoff collector are derived both from the upslope and downslope areas of the snowbank where snow thicknesses are least. It is thought that this variation prevents the rapid build-up of a wave in the basal saturated layer of the snow and results in a more gentle initial rise in the surface flow hydrograph. This hypothesis is supported by comparison of the hydrograph shapes typical of the four runoff plots (Figure 72). The location with the most uniform snow thickness, plot 1, exhibits the steepest rising limb. Plot 2, which has a large, lower snowbank area of relatively uniform thickness, shows the next steepest rising limb. Plots 3 and 4, possessing greater variation in snow depths are characterised by more gently rising curves.

Examination of the hydrographs recorded at the four plots reveals considerable variation in the times at which rises and peaks occur

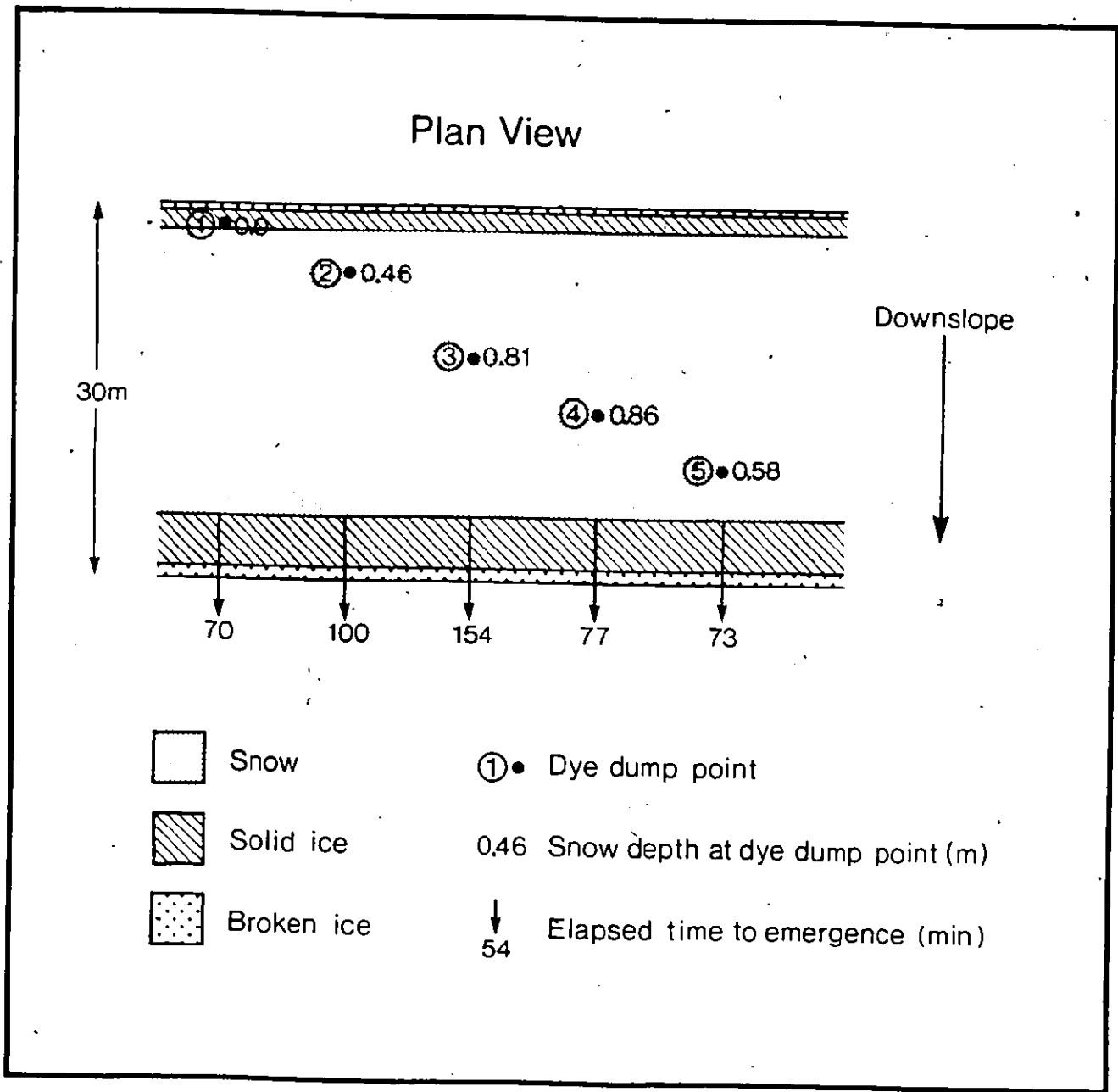


FIGURE 71 : INTRA-SNOW FLOW SPEEDS, PLOT 3, JULY 24, 1979

	X	Y	Z
Dye dump point	Distance from downslope snow-bank edge (m)	Depth of snow (m)	Elapsed time to emergence (min)
1	28.0	0.00	70.0
2	24.6	0.46	100.0
3	17.8	0.81	154.0
4	13.4	0.86	77.0
5	9.0	0.58	72.5
<p>REGRESSION EQUATION<sup>a</sup>: <math>Z = 2.77X + 88.45Y - 3.71</math></p> <p><math>R = +0.896</math></p> <p><math>R^2 = 0.802</math></p> <p><math>N = 6</math></p> <p><math>t_{calc} = 9.05</math>, which is greater than <math>t_{tables}</math> at 0.1% rejection level.</p> <p><sup>a</sup> includes point (0, 0, 0)</p>			

TABLE 22: Intra-snow flow speeds, plot 3, 1335h July 24, 1979

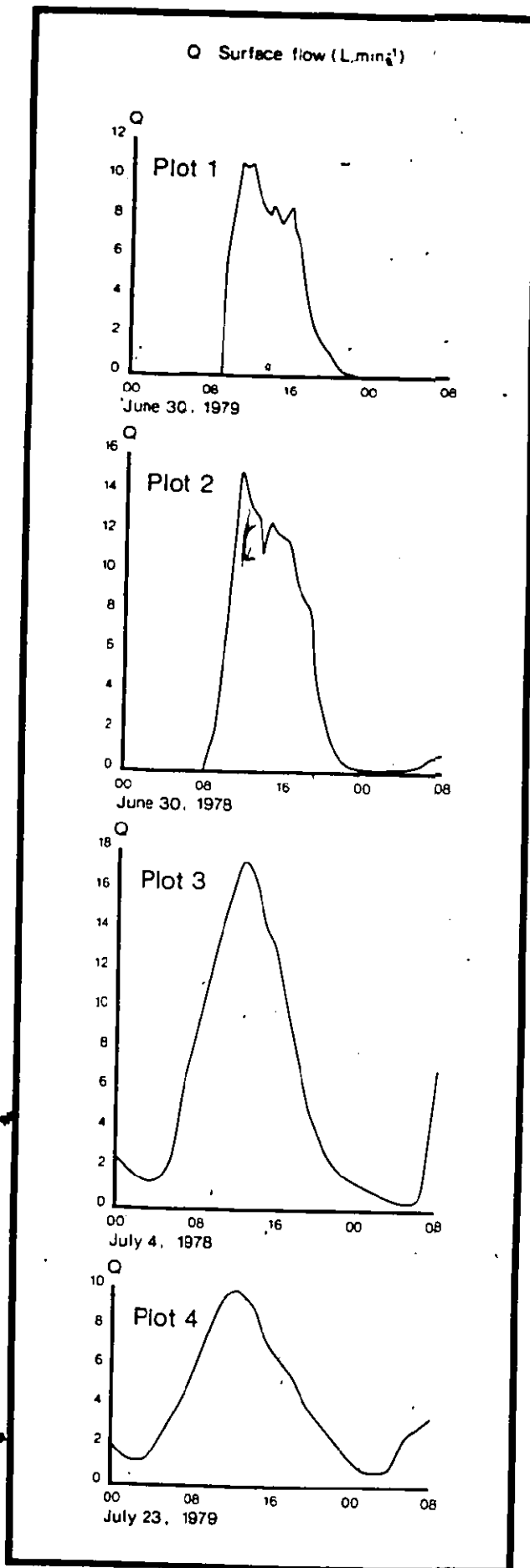


FIGURE 72 : SURFACE FLOW HYDROGRAPHS, PLOTS 1 - 4

(Figure 72). Factors influencing this variation include (1) orientation relative to radiation inputs (i.e. slope angle and aspect), (2) actual meteorological conditions (cloud cover, air temperature, wind speed, etc.), (3) the rapidity of snowmelt (fluxes of greater magnitude travel faster), (4) snow depth (the time for vertical movement in the snow is the major element of the lag), and (5) snow crystal size (which controls permeability). Even successive hydrographs, in which most of the snow factors remain constant, show variability in time of rise and peak. Over the five days shown in Figure 70, for example, times of rise at plot 3 varied from 0415h to 1045h, and times of peak from 1200h to 1845h.

As the ablation season progresses, a trend exists for lag times between energy inputs and surface flow response to decrease, as a result of diminishing snow thickness and increasing snow crystal size. Data for plot 3 in 1978, the longest recorded succession of hydrographs, show that the times of hydrograph rise, of the point mid-way to peak, and of hydrograph peak, occur earlier in the day later in the season (Figure 73; Table 23). There is little change in the average number of hours elapsing between rise and peak, but the peak itself averages 9.25 hours after the time of maximum incoming radiation at the beginning of the ablation season, and only 0.60 hours after it at the end of the season. In addition, the steeper slope of the mid-point regression (Figure 73) indicates that the shape of the hydrograph changes as the ablation season progresses, to a steeper initial limb and a broader peak. This change also was noted in plot runoff hydrographs recorded near Yellowknife, N.W.T. (Landals and Gill, 1973). Considerable variance exists about the regression lines, but this is expected since no attempt has been made to standardize for varying meteorological conditions and rates of surface melt.

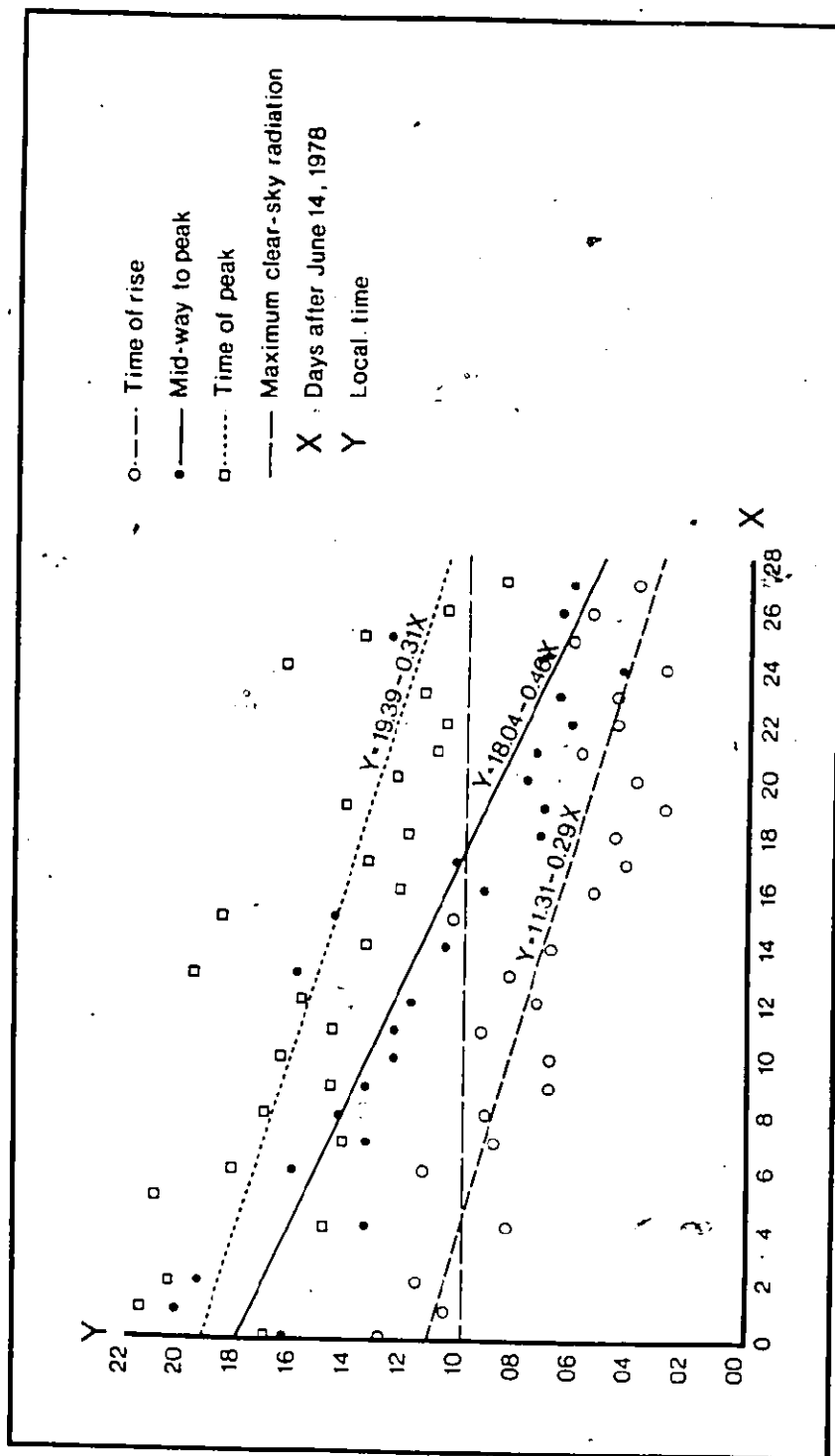


FIGURE 73 : CHANGES IN TIMING OF SURFACE FLOW HYDROGRAPHS, PLOT 3, 1978

	Regression equation <sup>a</sup>	N	R	R <sup>2</sup>	t <sub>calculated</sub> <sup>b</sup>
PEAK	$Y = 19.39 - 0.31X$	26	+0.747	0.558	5.729
MID-POINT	$Y = 18.04 - 0.46X$	25	+0.863	0.744	8.528
RISE	$Y = 11.31 - 0.29X$	25	+0.834	0.695	7.551

<sup>a</sup> X = Number of days since surface flow first generated (June 14=0)  
Y = Time of rise/mid-point/peak (hours local time)  
<sup>b</sup> Values significant at less than the 0.1% rejection level

**TABLE 23:** Linear regression of time of surface flow hydrograph rise, mid-way point, and time of peak, vs. days after surface flow generated, plot 3, 1978

The trend of decreasing lag times exhibited at plot 3 is present at all plots in all years, but correlation coefficients elsewhere are not significant at less than the 5 % rejection level. Introduction of water released per unit area of the snowbank per day ( $W_+$ ), as a second independent variable, produces additional significant correlation coefficients at the 5 % level at plot 1 in 1979 and plot 4 in 1978 (Table 24). Additional regressions were not undertaken for the other locations because of small sample size.

The recession limbs of the surface flow hydrographs represent the emergence of slow-moving melt fluxes developed after the time of peak surface melt. Slow drainage of the pack continues until the arrival of the shock-front of the succeeding day's melt. Recession limbs at any one location are expected to become steeper during the ablation season as both vertical and horizontal flow paths decrease. The steepening trend is exemplified in the 1978 data set from plot 3 (Figure 74). Obvious steepening began after July 2 following a number of days of particularly rapid snow loss. Similar trends were developed at the other plots.

A final variation in the shape of surface flow hydrographs is caused by the generation of surface flow by rainfall. This occurred on only two occasions during the three field seasons, both in 1977, illustrating the relative unimportance of rainfall compared with snowmelt in generating surface flow. Moreover, it is probable that without antecedent snowmelt to produce high active layer moisture contents, surface flow would not have been generated during these precipitation events.

The first rain-induced flow event occurred on June 15, 1977, when 6.25 mm of rain fell in 5 hours, with a maximum intensity of 2.5 mm  $h^{-1}$ . Plot 2 was being monitored during this period. The recorded

Location and year	N	Simple regression equation <sup>a</sup>	R	R <sup>2</sup>	t-test significance level	Multiple regression equation <sup>b</sup>	R	R <sup>2</sup>	t-test significance level
Plot 1, 1979	5	$Y=13.94-0.91X$	+0.770	0.600	10%	$Z=15.90-0.68X-0.13Y$	+0.960	0.922	2%
Plot 3, 1978	26	$Y=19.39-0.31X$	+0.747	0.558	<0.1%	$Z=19.88-0.33X-0.01Y$	+0.759	0.576	<0.1%
Plot 3, 1979	11	$Y=19.13-0.69X$	+0.906	0.820	<0.1%	$Z=19.57-0.67X-0.01Y$	+0.885	0.783	<0.1%
Plot 4, 1978	6	$Y=12.74-0.43X$	+0.768	0.590	10%	$Z=16.17-0.70X-0.03Y$	+0.849	0.721	5%

<sup>a</sup>X = number of days since surface flow first generated; Y = time of peak (hours local time).  
<sup>b</sup>X = number of days since surface flow first generated; Y = water released per unit area per day; Z = time of peak (hours local time).

TABLE 24: Linear and multiple regression, time of peak surface flow

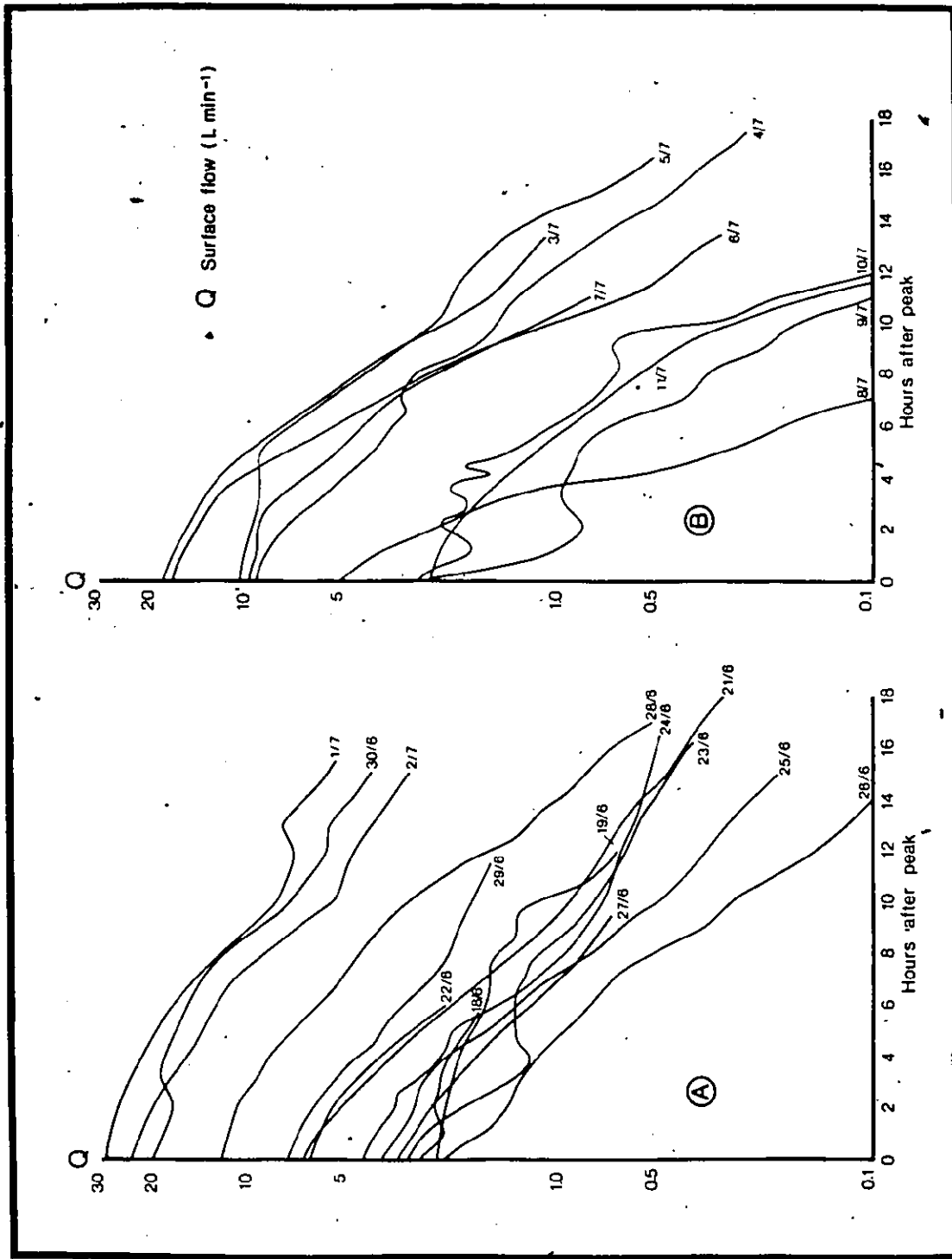


FIGURE 74 : SURFACE FLOW HYDROGRAPH RESSION, PLOT 3, 1978

A - Recession curves, June 18 - July 2;

B - Recession curves, July 3 - July 11.

hydrograph can be separated by regression into a peak attributable to rainfall, and "baseflow" attributable to snowmelt (Figure 75). The response to rainfall is almost immediate and there is an equally rapid recession as the intensity decreases. These points suggest that the runoff was derived from areas with short flow paths to the surface flow collector. The volume of rain falling on the saturated area downslope of the snowbank accounts for 92 % of the hydrograph peak. Since the runoff peak represents only 20 % of the total rainfall falling on the plot, however, most was lost as infiltration into the unsaturated areas of the plot, and to water storage within the snowbank.

On June 23, 1977, the greatest 24-hour rainfall event recorded in the three field seasons occurred between 0000 h - 0300 h. Approximately 7.5 mm of precipitation fell and surface flow was generated at plot 2. In view of the low rainfall intensity value, the runoff is thought to have been saturated overland flow rather than the Hortonian type. All snow had ablated at the plot three days prior to the storm. In contrast, snowmelt at plot 1 had ceased on June 4, and neither this storm nor the earlier one of June 15 produced surface runoff at this location. Other precipitation events during the three field seasons, including a 7.2 mm storm on July 23 - 24, 1978, failed to generate surface flow at the four runoff plots.

The explanation of the lack of surface flow response to rainfall may lie in the rapid lowering of the frost table that follows snow ablation, and in the soil moisture deficits that develop within the active layer as a result of continual evapotranspiration and subsurface flow. Rainfall appears able to generate surface flow only on an area saturated by snowmelt (June 15, 1977), or on one very recently saturated (June 23, 1977). Consequently, the low magnitude rainfall events typical of the high Arctic have the highest probability of generating surface flow in those areas

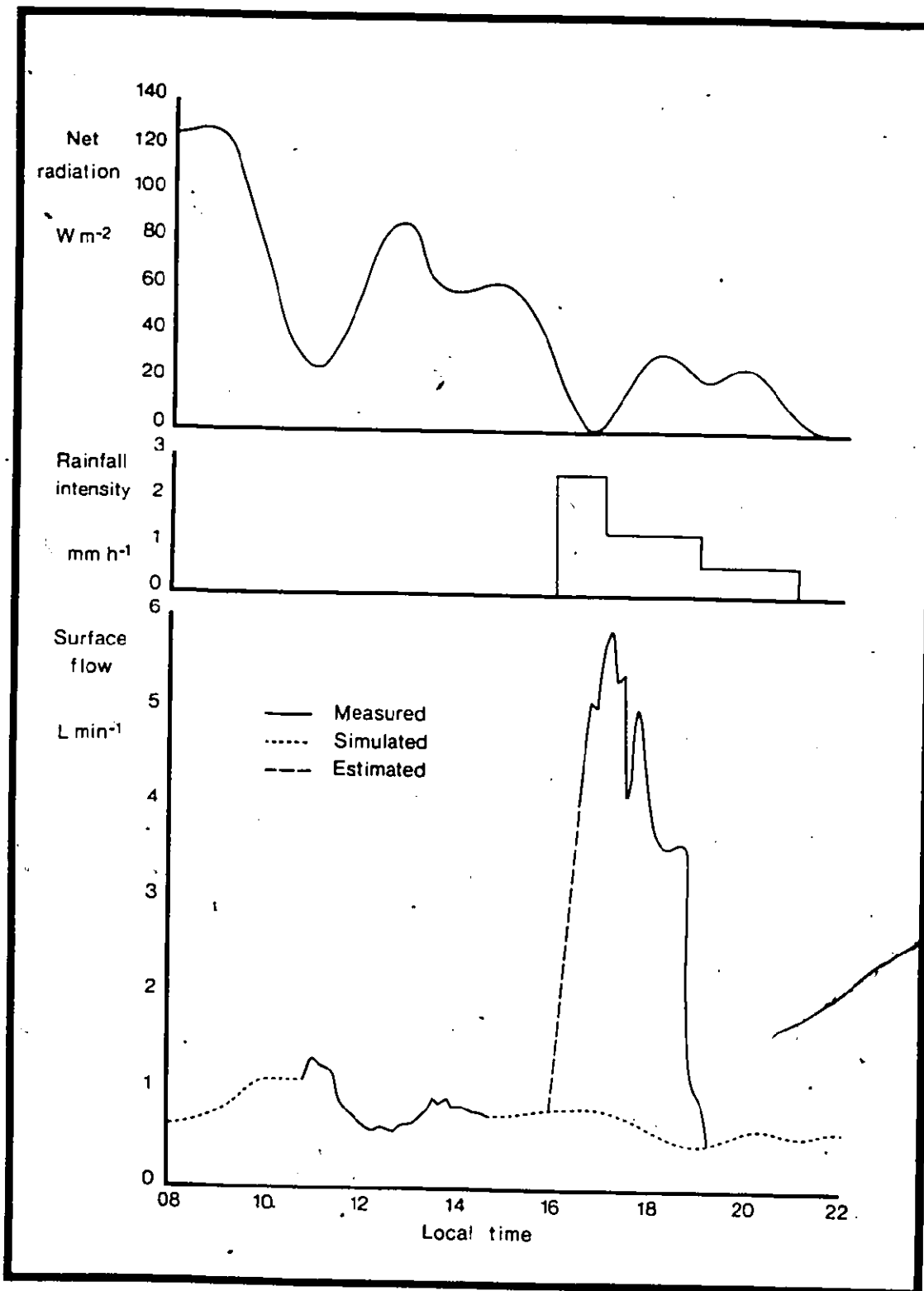


FIGURE 75 : RAINFALL AND SURFACE FLOW, PLOT 2, JUNE 15, 1977

Simulation of surface flow used the following equation:

$$\text{Surface flow} = 0.005 R_n + 0.48$$

where:  $R_n$  is lagged 2 hours,  $N = 25$ .

subjected to snowmelt inputs for long periods (i.e. snowbanks themselves and the areas downslope of them). Clearly, it would be rare for rainfall-induced surface flow to be produced at interfluvial locations where snowmelt lasts for only four or five days of the year.

It follows that the partial area contribution theory is applicable to high Arctic catchments. During the majority of precipitation events, the only parts of a catchment to produce surface flow are those where snowbanks still exist, or where they have recently existed. The areas contributing to stream runoff vary from one storm to another, depending on timing relative to the progress of catchment snow ablation. Only in very exceptional rainfall events (e.g. Cogley and McCann, 1976), and aided by the presence of impermeable permafrost, are other areas of the catchment likely to produce surface flow.

In summary, most surface flow hydrographs recorded at the runoff plots are explicable in terms of snowmelt theory. Lag times between energy inputs and surface flow are shorter at locations with thin snowpacks than they are at locations with deep snow accumulations. Lag times generally decrease during the ablation season as the pack is reduced in thickness and snow crystal sizes increase. On a daily basis, however, considerable variation exists in the lags at a single site because of the importance of the surface melt rate in influencing the shock-front speed in the unsaturated zone. On the falling limbs, recessions are longest at thick snowpacks where they may extend for 40 - 50 hours after peak. The reduction in snow thickness during the season causes the recession limbs to become steeper. The overall shape of a snowmelt-derived hydrograph at snowbank locations becomes less asymmetrical and less peaked as the ablation season progresses. Surface flow generated by rainfall is rare, and relies on the

occurrence of rain during, or immediately following, snowmelt at a particular site. The partial area contribution and variable area concepts appear applicable to high Arctic catchments, but surface flow is most likely to be generated downslope of snowbank locations, rather than as in temperate regions, in areas along the stream banks.

CHAPTER 6

EROSION AND SEDIMENT YIELDS

## Surface sediment transport

### (a) Suspended sediment concentrations

Following experimental techniques employed in 1977 and 1978 (see pp. 135 - 138), systematic measurements of suspended sediment concentrations in surface wash were made in 1979. Inferences about the earlier years may be drawn from these data.

A total of 78 samples were collected at the runoff plots in 1979, generally at one or two hour intervals on selected days. Since more than half of these samples possessed concentrations less than  $4 \text{ mg L}^{-1}$ , a prime conclusion is that suspended sediment concentrations in surface wash are low. Moreover, with the exception of plot 1 on June 30, 1979, no diurnal or seasonal cycles in sediment concentrations are apparent. Changes in concentrations depend on the availability of sediment and on the transport capacity of the flow. The absence of cycles at the three snowbank plots and the single cycle at plot 1, both imply that concentrations are weathering-limited rather than transport-limited.

More than half of the total season's surface runoff was generated on June 30, 1979 at plot 1, and a maximum discharge of  $10.7 \text{ L min}^{-1}$  was recorded (Figure 76). The concentrations of suspended sediment reached  $70 \text{ mg L}^{-1}$ , but peaked one hour before maximum runoff and subsequently declined in spite of increasing runoff rates. Similar patterns of sediment concentrations have been observed in surface wash generated by natural and simulated rainfall (e.g. Emmett, 1970; 1978; Lowdermilk and Sundling, 1950; Yair and Lavee, 1976), and indicate removal of the most easily transported particles in a relatively short time. Emmett (1970), however, recorded marked variations in concentrations on subsequent days, after his runoff

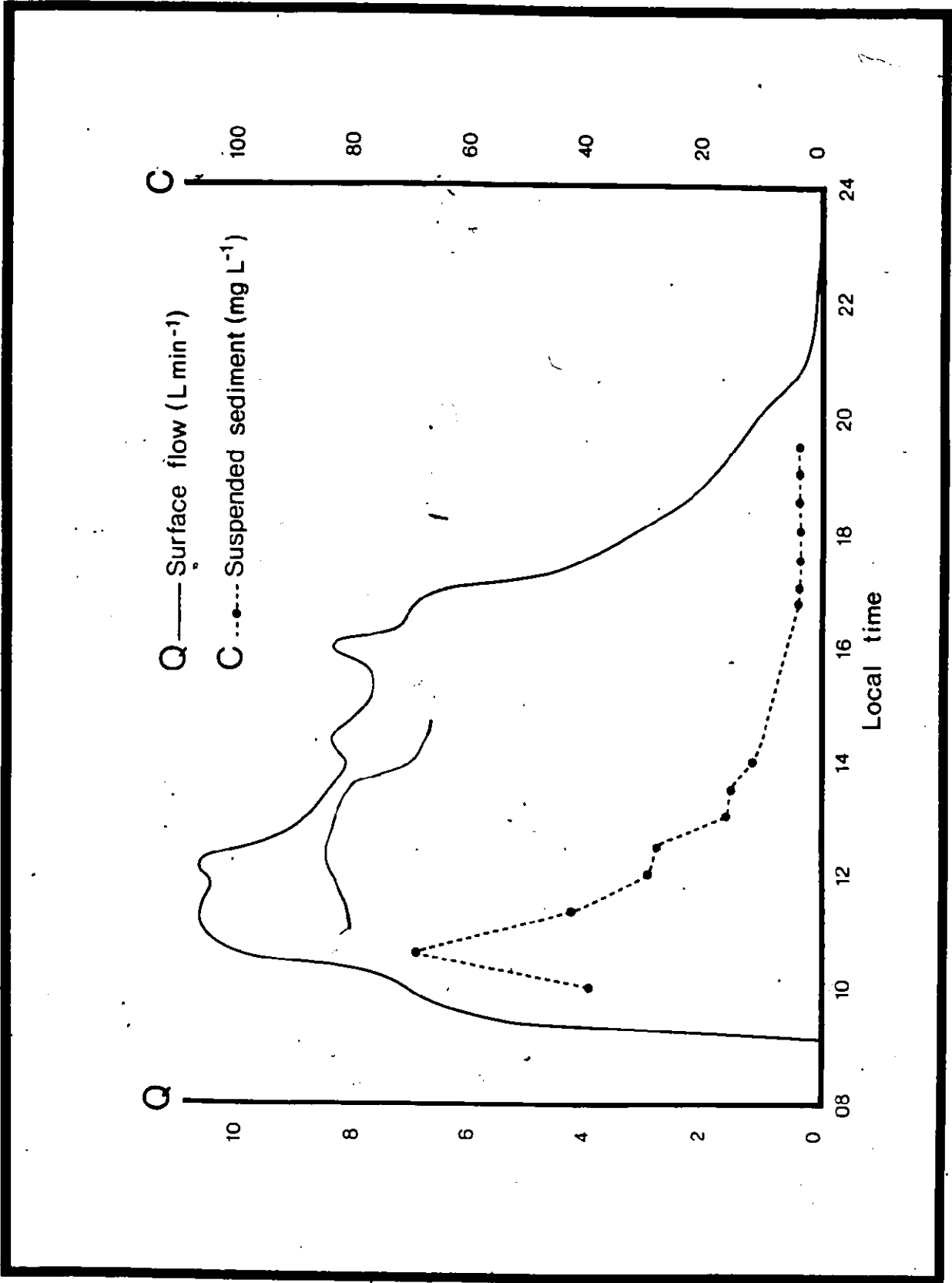


FIGURE 76 : SURFACE FLOW AND SUSPENDED SEDIMENT CONCENTRATIONS, PLOT 1, JUNE 30, 1979

plots had dried overnight. Such trends were not observed at plot 1 and a number of explanations are proposed. First, later discharges (maximum of  $4.7 \text{ L min}^{-1}$ ) may have been too low to pick up measurable quantities of sediment. Second, the drying process, to which Emmett (1970) attributed the renewal of sediment supply, may be less effective in Arctic areas since water release from the snow is near-continuous. Third, the absence of raindrop impact during the flow events at plot 1, may have resulted in an earlier washing-in of fines and the formation of a filtration pavement. Fourth, a mechanism peculiar to frozen ground may have been at work. Wedel et al. (1978) suggest that the initial thawing of the ground surface on Bathurst Island creates an uncohesive soil which requires several days to settle, and is particularly erodible over this period. This may have caused the high concentrations to develop at plot 1.

The absence of any observable cycles at the other three plots, in spite of unit discharges greater than those at plot 1, is thought to result from the presence of vegetation. The feedback mechanism is such that the wettest areas (i.e. those with the highest unit discharges) coincide with those possessing a near-continuous meadow tundra vegetation cover. Moreover, as a result of longer hydrograph recessions at these sites, little or no drying can take place between one day and the next.

A limited number of experiments on the hydraulic parameters of surface flow were undertaken. At peak flow at plot 1 on June 30, 1979, an area of flow selected as one of the most rapid, had surface velocities of  $12.3 \times 10^{-2} - 17.1 \times 10^{-2} \text{ m s}^{-1}$  (measured with the silver foil method (pp. 129 - 130)). Assuming average velocity to be half the surface velocity (Emmett, 1978) and water temperature to be  $5^{\circ}\text{C}$ , a few simple

measurements (flow depth, width and slope) enable calculation of the Reynolds' number ( $R_e$ ), Froude number ( $F_r$ ), Darcy-Weisbach friction coefficient ( $f_f$ ) and the Manning  $n$  (Table 25). Following Leopold et al. (1964, p. 154) and Pearce (1976a; 1976b), the Reynolds' numbers do not include the factor of four used by Emmett (1970), so that his figures must be reduced accordingly to render them comparable.

On June 30, 1979 at plot 1, the  $R_e$  values (369 - 514) indicate a flow falling into the laminar-transitional range, rather than strictly laminar as suggested by the presence of surface standing waves. Since sediment was being removed at this time, it is clear that erosion by thin flows does not require turbulence, a result that concurs with the findings of Emmett (1970) and Pearce (1976b). The  $F_r$  values averaged 0.24, slightly higher than the range measured by Emmett (1970), and much greater than the maximum value (0.09) recorded by Pearce (1976a). The value of 0.24 is still within the tranquil regime, however, and well below critical. Values of resistance to flow, shown by  $f_f$  and  $n$ , are in the low range of Emmett's measurements, probably because of greater flow depths. Values of  $n$  are in the high range in comparison with rivers (Gregory and Walling 1973, p. 129) indicating the increased resistance experienced in thin flows. Results of the same order were obtained for three areas of flow on plot 3 on July 11, 1978 (Table 25).

At both plots 1 and 3, depths of flow were one to two orders of magnitude greater than those typically found in laboratory experiments. The depths are indicative of topographic irregularities and surface roughness, and exemplify the potential dangers of transferring laboratory results to natural slopes. Moreover, comparison of average flow speeds measured with rhodamine dye, and the spot measurements made at the same time at plot 3

	Surface Velocity ( $\times 10^{-2} \text{ m s}^{-1}$ )	Mean Velocity ( $\times 10^{-2} \text{ m s}^{-1}$ )	Flow Depth (mm)	Hydraulic Radius (mm)	Reynolds Number	Froude Number	Darcy-Weisbach Coefficient	Manning n
PLOT 1 June 30, 1979	12.28	6.14	10.0	9.0	369	0.207	6.56	0.13
	13.88	6.94	10.0	9.0	417	0.234	5.13	0.12
	17.14	8.57	10.0	9.0	514	0.288	3.37	0.09
PLOT 3 July 11, 1978	10.53	5.27	5.0	4.2	147	0.260	10.35	0.15
	11.58	5.79	5.0	4.2	162	0.290	8.55	0.13
	11.58	5.79	5.0	4.2	162	0.290	8.55	0.13
	16.32	8.16	5.0	4.2	229	0.400	4.31	0.09
	16.81	8.41	5.0	4.2	235	0.410	4.06	0.09
	26.32	13.16	5.0	4.2	369	0.650	1.66	0.06
	9.60	4.80	5.0	4.1	131	0.240	12.15	0.16
	10.56	5.28	5.0	4.1	144	0.260	10.04	0.14
	21.60	10.80	5.0	4.1	295	0.540	2.40	0.07
	6.97	3.49	5.0	4.4	102	0.170	24.66	0.23
8.97	4.49	5.0	4.4	132	0.220	14.98	0.18	
	RHODAMINE	3.90						

TABLE 25: Hydraulic parameters of surface flow, plots 1 and 3

indicates the high degree of variation that is possible over relatively small distances in the field (Table 25).

In summary, all suspended sediment concentrations measured at the plots were low ( $<70 \text{ mg L}^{-1}$ ). The maximum value was recorded at plot 1 where an incomplete cover of vegetation permits sediment removal, and where a single diurnal concentration cycle was noted during sub-critical, non-turbulent flow. At the snowbank locations, under similar flow conditions, no such cycles were observed. The data suggest that suspended sediment transport is governed by sediment availability and not flow transport capacity.

(b) Suspended sediment amounts

The amounts of sediment removed in suspension can be assessed from the surface wash volumes and the sediment concentrations. Extraction of data from the sediment concentration and discharge records at plot 1 on June 30, 1979, for example, gives a total of 104.0 g of sediment removed. Records for the other four days of surface flow production at this site indicate a further 18.5 g of sediment leaving the weir slot. In addition, 54.5 g of sediment was trapped in either the weir or the collector at the end of the snowmelt runoff. The total of all these amounts is 177.0 g, giving an average concentration of  $21.3 \text{ mg L}^{-1}$  of sediment in the surface wash. Unfortunately, this type of analysis cannot be undertaken for the other plots, because variations in suspended sediment concentrations are too close to the level of discrimination of the measurement technique (see pp. 137 - 138).

If all concentrations in the category of  $<4 \text{ mg L}^{-1}$  are assumed to average  $2 \text{ mg L}^{-1}$ , the mean concentrations measured at plots 2, 3 and 4 in 1979 are  $2.6 \text{ mg L}^{-1}$ ,  $3.3 \text{ mg L}^{-1}$ , and  $4.0 \text{ mg L}^{-1}$  respect-

ively. With a standard deviation of  $2 \text{ mg L}^{-1}$  implicit in the correction factor, these values (and those of plot 1) are subject to errors of  $\pm 4 \text{ mg L}^{-1}$  at the 95% level. These figures are thought applicable to surface flow volumes measured in earlier years. When combined with known minima (i.e. sediment collected from the weirs), a range of probable sediment totals are produced (Table 26), and weights of sediment removed per  $\text{m}^2$  of plot surface can be derived. Best-estimates of these figures (Table 27) vary by an order of magnitude among the plots, with plot 2 showing the lowest rate of sediment loss, and plot 3 the highest.

The range of values of weight loss,  $0.02\text{-}1.74 \text{ g m}^{-2} \text{ a}^{-1}$ , can be compared with measurements made by A. Jahn (1961) on unvegetated slopes in Spitsbergen (see pp. 100-101). The maximum value recorded by Jahn,  $18 \text{ g m}^{-2} \text{ a}^{-1}$ , exceeds the maximum at Thomsen River by a factor of ten. Indeed, the former's minimum rates of loss of  $1 \text{ g m}^{-2} \text{ a}^{-1}$  are closer to the maximum recorded in the study area. It is probable that these differences result from the absence of feedback between sustained moisture supply and vegetation on the coarse regolith slopes studied by Jahn. Without this feedback a positive correlation would be expected between snowbank size and surface weight loss. This correlation existed in Jahn's study but not at Thomsen River. There, the interfluvial plot experienced more erosion than plot 2, the result of greater sediment concentrations in surface wash on the largely unvegetated surface. The potential of the regolith to support plant life, if supplied with sufficient moisture clearly is an important factor in the prediction of slopewash erosion rates over wider areas. It may be that with a shorter growing season in true polar desert areas, even under terrain conditions similar to those at Thomsen

Location	Year	Probable minimum concentration (mg L <sup>-1</sup> )	Probable maximum concentration (mg L <sup>-1</sup> )	Best-estimate concentration (mg L <sup>-1</sup> )	Surface flow volume (m <sup>3</sup> )	Probable minimum sediment weight loss (g)	Probable maximum sediment weight loss (g)	Best-estimate sediment weight loss (g)
Plot 1	1977	17.3 <sup>a</sup>	25.3 <sup>a</sup>	21.3 <sup>a</sup>	0.96	16.6	24.3	20.4
	1979	17.3	25.3	21.3	8.33	144.1	210.7	177.4
Plot 2	1977	0.1 <sup>b</sup>	6.6 <sup>a</sup>	2.6 <sup>a</sup>	26.39	5.6 <sup>c</sup>	174.2	68.6
	1978	0.2 <sup>b</sup>	6.6 <sup>a</sup>	2.6 <sup>a</sup>	11.74	2.5 <sup>c</sup>	77.5	30.5
	1979	0.1	6.6	2.6	3.90	0.8	25.7	10.1
Plot 3	1978	0.1 <sup>b</sup>	7.3 <sup>a</sup>	3.3 <sup>a</sup>	137.90	8.4 <sup>c</sup>	1006.7	455.1
Plot 4	1978	0.4 <sup>b</sup>	8.0 <sup>a</sup>	4.0 <sup>a</sup>	20.30	7.4 <sup>c</sup>	162.4	81.2

a inferred from 1979 results  
b sediment collected from weir/surface flow volume  
c sediment collected from weir at the end of the ablation season

TABLE 26: Calculated and inferred suspended sediment concentrations and weight loss from the runoff plots

Plot	Year	Area (m <sup>2</sup> )	Rate of sediment loss from the whole plot (g m <sup>-2</sup> a <sup>-1</sup> )			Area of plot affected by wash (%)	Rates of sediment loss from areas affected (g m <sup>-2</sup> a <sup>-1</sup> )		
			Probable minimum	Probable maximum	Best estimate		Probable minimum	Probable maximum	Best estimate
1	1977	149.6	0.11	0.16	0.14	40.0	0.28	0.41	0.34
	1979	437.7	0.33	0.48	0.41				
2	1977	524.7	0.01	0.33	0.13	29.7	0.04	1.12	0.44
	1978	524.7	<0.01	0.15	0.06				
	1979	524.7	<0.01	0.05	0.02				
3	1978	262.3	0.03	3.84	1.74	30.1	0.11	12.74	5.76
4	1978	67.0	0.11	2.42	1.21	34.8	0.32	6.97	3.48

TABLE 27: Suspended sediment losses averaged over all and affected areas of the runoff plots

River, surface wash denudation rates are greater due to a decline in the effectiveness of surface protection by vegetation. If this is the case, it modifies the inference drawn by French (1976, p. 142) which predicts the highest erosion rates for areas transitional between tundra and polar desert.

Within the boundaries of the runoff plots, only those areas downslope of melting snow are affected by surface flow and subject to sediment removal. These areas constitute 30-54% of the total plots (calculated from snow edge retreat records). Erosion rates averaged over these limited areas are 2-3 times greater than those averaged over the plots, with a range of best-estimate values of  $0.06-5.76 \text{ g m}^{-2} \text{ a}^{-1}$  (Table 27). The order of weight losses, with plot 3 experiencing the highest and plot 2 the lowest rates, remains the same as for the total areas. Annual variation in rates appears considerable with loss rates for the affected area at plot 2 varying by a factor of seven over three years of measurements. The main conclusion, however, is that even averaged over the plot areas affected by wash, surface sediment loss rates are low and less than those recorded in other polar areas.

Conversion of weight loss to surface lowering (i.e. denudation) requires an estimate of the surface bulk density of the materials. Except in solid rock, this value is less than the actual density of the constituent particles. On a silty sand, Pearce (1976b) measured lowering rates and the weight of sediment removed and found the bulk density of the eroded surface to be  $0.67 \text{ Mg m}^{-3}$ . This conversion factor is used to relate weight loss to rates of surface lowering by surface wash over the whole plots and affected areas (Table 28).

Plot	Year	Area affected (m <sup>2</sup> )	Rate of surface lowering ( $10^{-3}$ mm a <sup>-1</sup> ) <sub>23</sub> (bulk density assumed to be 0.67 Mg m <sup>-3</sup> )									
			Over the whole plot					Over affected areas				
			Probable minimum	Probable maximum	Best-estimate	Best-estimate average	Best-estimate maximum	Probable minimum	Probable maximum	Best-estimate	Best-estimate average	Best-estimate maximum
1	1977	59.8	0.16	0.24	0.21	0.41	0.42	0.61	0.51	0.82		
	1979	235.8	0.49	0.72	0.61		0.91	1.33	1.12			
2	1977	156.0	0.01	0.49	0.19		0.06	1.67	0.66			
	1978	193.3	0.01	0.22	0.09	0.10	0.01	0.60	0.24	0.33		
	1979	160.1	0.01	0.07	0.03		0.01	0.24	0.09			
3	1978	79.0	0.04	5.73	2.60	2.60	0.16	19.01	8.60	8.60		
4	1978	23.3	0.16	3.61	1.81	1.81	0.48	10.04	5.19	5.19		

TABLE 28: Rates of surface lowering by suspended sediment loss in surface wash over all and affected areas of the runoff plots

Best-estimate average rates of lowering range from  $0.10 \times 10^{-3}$  to  $2.60 \times 10^{-3}$  mm a<sup>-1</sup> over all the plots, and from  $0.33 \times 10^{-3}$  to  $8.60 \times 10^{-3}$  mm a<sup>-1</sup> over the wash affected areas (Table 28). These values can be compared with the global range of denudation rates by surface wash of  $0.1 \times 10^{-3}$  to  $>230.0$  mm a<sup>-1</sup> (Young, 1974, p. 69). The loss rates on the runoff plots are amongst the lowest in the world. They are even lower than  $50 \times 10^{-3}$  mm a<sup>-1</sup>, a figure predicted (using Sachs Harbour climatic data) from a nomograph in Carson and Kirkby (1972, p. 219, Figure 8.18). This is probably a reflection of the unexpected effectiveness of vegetation as a protective surface cover at high latitudes. Moreover, it is likely that the average erosion rate over the study area is less than those recorded at the runoff plots, since some parts of the landscape are not subjected to surface wash.

It is pertinent to compare the erosion rates measured in this study with others from nonpermafrost areas where snowmelt is an important part of the annual hydrological cycle. C. Thorn (1974) recorded surface wash erosion in the alpine environment of the Colorado Front Range, while A. Pearce (1976a; 1976b) conducted similar research in northern Ontario. Thorn's (1974) denudation rates of  $3.5 \times 10^{-3}$  to  $8.4 \times 10^{-3}$  mm a<sup>-1</sup>, recorded downslope of snowbanks, are of the same order as those at Thomsen River. They suggest that snowmelt-derived surface wash is ineffective as an erosional agent in both environments. In contrast, Pearce recorded denudation of up to 1 mm on unvegetated slopes during the snowmelt period, and up to 6 mm over the course of a year. These data emphasise the importance of sediment detachment by raindrops, but also reveal that snowmelt can be effective as an erosional agent if it is sufficiently rapid.

In summary, the movement of sediment particles in surface wash in the Thomsen River periglacial environment, even on areas immediately downslope of ablating snowbanks, appears to be a process of minor importance. Erosion rates are comparable to others measured in an alpine area, but are in the low end of the global range.

(c) Dissolved sediment concentrations

Measurements of the specific conductance of 255 samples of surface wash and 300 samples of subsurface wash were undertaken during the three field seasons. Of these samples, 67 were subjected to field titration for total hardness, and 33 were analysed for calcium and magnesium hardness (Schwarzenbach method). In addition, 8 samples were analysed independently using atomic absorption spectrophotometry and colorimetry (see p. 138). These data sets can be used to calculate the load of total dissolved solids (T.D.S.) in surface wash, and to identify the major constituent ions present in solution.

A high degree of correlation exists between total hardness and specific conductance ( $R = +0.997$ ,  $n = 67$ ;  $t = 103.9$ , significantly different from zero at better than the 0.1% rejection level) (Figure 77). Equally good correlations (significant at better than the 0.1% rejection level) are apparent between total hardness and calcium hardness ( $R = +0.995$ ,  $n = 33$ ;  $t = 55.5$ ), and total hardness and magnesium hardness ( $R = +0.993$ ,  $n = 33$ ;  $t = 46.8$ ) (Figure 78). It follows, therefore, that the specific conductance data can be used to predict both total hardness and its calcium and magnesium components. Moreover, the linear relationship implies that the proportions of other solutes in the load are fixed. This hypothesis is supported by the eight samples subjected to laboratory analysis (Table 29) which independently affirm the values of the hardness tests. The detailed breakdown of cations shows that despite the absence of carbonate bedrock in

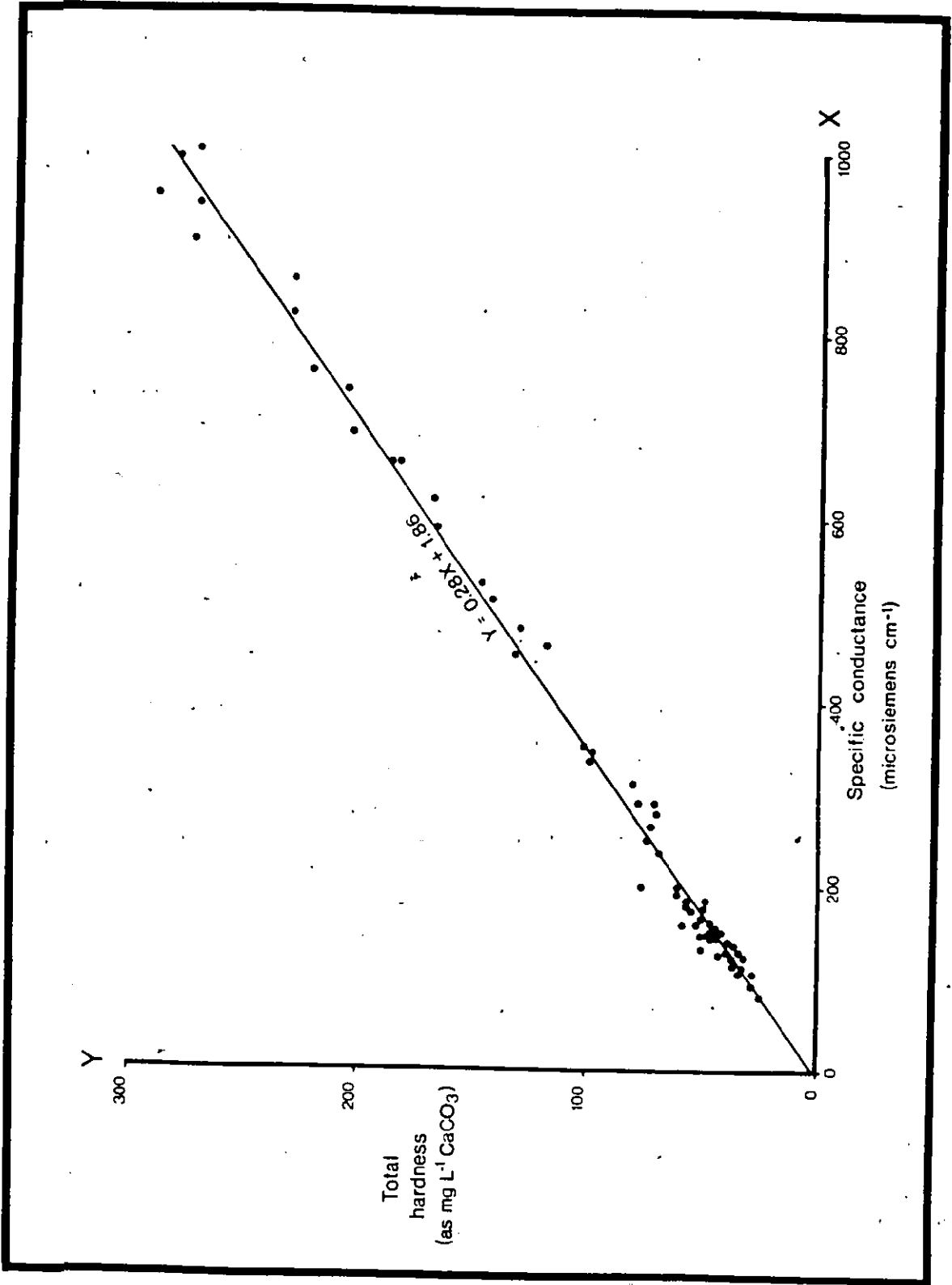


FIGURE 77 : CORRELATION OF SPECIFIC CONDUCTANCE AND TOTAL HARDNESS IN SURFACE AND SUBSURFACE WASH SAMPLES

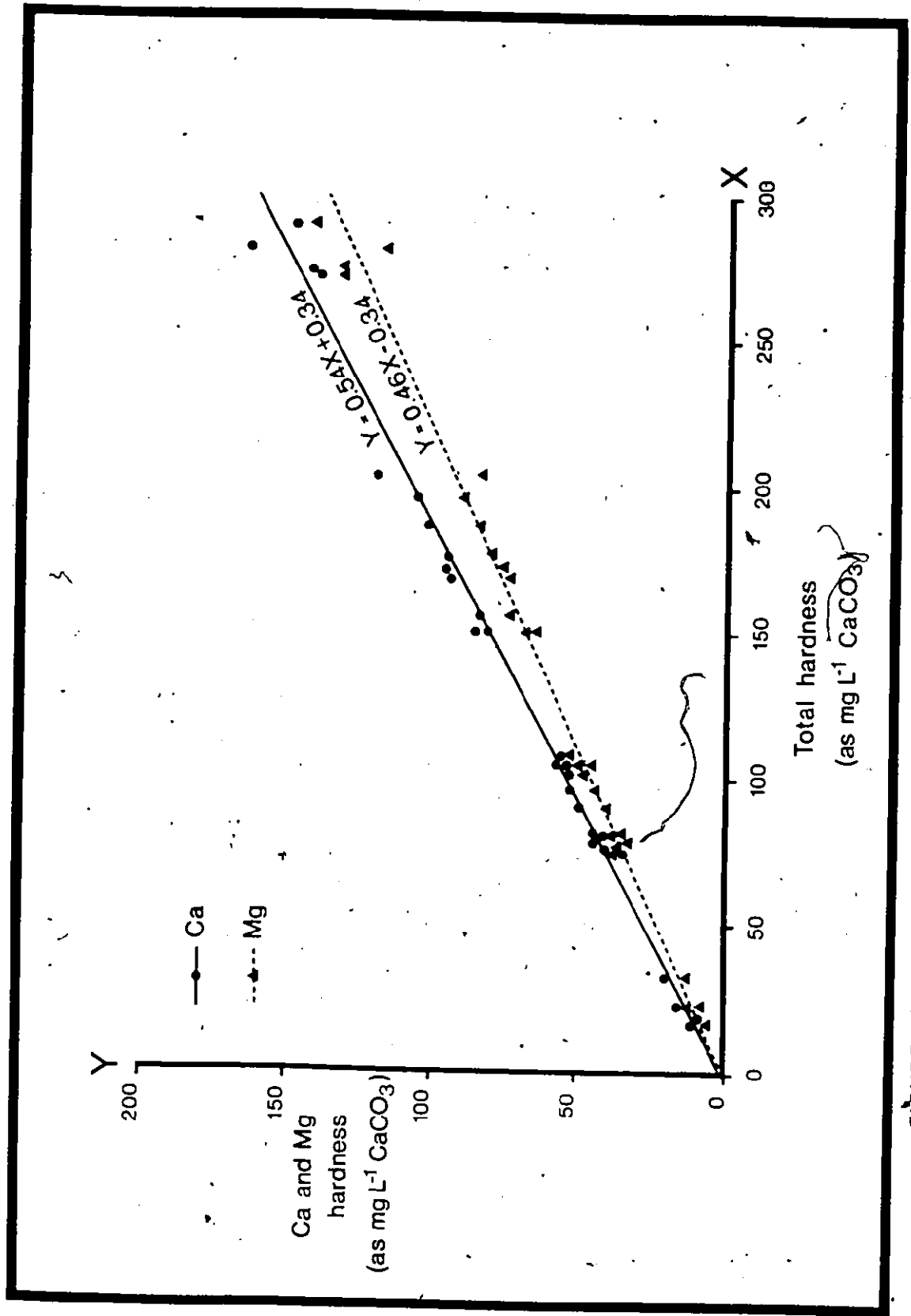


FIGURE 78 : CORRELATION OF TOTAL, CALCIUM AND MAGNESIUM HARDNESS IN SURFACE AND SUBSURFACE WASH SAMPLES

SAMPLE	LABORATORY ANALYSIS					FIELD TEST	Predicted <sup>a</sup>	
	Ca <sup>2+</sup> (mg L <sup>-1</sup> )	Mg <sup>2+</sup> (mg L <sup>-1</sup> )	Na <sup>+</sup> (mg L <sup>-1</sup> )	K <sup>+</sup> (mg L <sup>-1</sup> )	SiO <sub>2</sub> (mg L <sup>-1</sup> )		Ca <sup>2+</sup> (mg L <sup>-1</sup> )	Mg <sup>2+</sup> (mg L <sup>-1</sup> )
1	14.69	7.63	0.22	0.27	1.58	220	14.2	8.9
2	15.97	8.81	0.26	0.39	1.68	260	16.7	10.4
3	2.01	1.13	0.07	0.11	0.37	33	2.5	1.7
4	12.63	5.86	0.21	0.24	1.26	190	12.4	7.8
5	15.30	8.36	0.25	0.31	1.58	240	15.7	9.6
6	12.48	6.95	0.20	0.27	1.47	200	13.2	8.1
7	17.18	9.47	0.26	0.30	1.74	280	18.2	11.1
8	0.12	0.08	0.05	0.04	0.26	2.8	N.D.	N.D.
95% Confidence Limits	± 0.03	± 0.02	± 0.01	± 0.01	± 0.19			

<sup>a</sup> From Figures 77, 78 and atomic weights

TABLE 29: Solute concentrations, laboratory analysis and field tests  
(laboratory analysis undertaken by T. Gallie, University of British Columbia)

the study area, the major ones present are  $\text{Ca}^{2+}$  and  $\text{Mg}^{2+}$ .

Calculation of the T.D.S. concentrations from specific conductance using the conversion factor of 0.65 (the middle of the range given by Rainwater and Thatcher (1960)) reveals a residual in the solutes which indicates the presence of unidentified anions in solution.

T.D.S. values in samples of surface wash range between 2 and 488  $\text{mg L}^{-1}$ , with a mean of 103  $\text{mg L}^{-1}$  (Table 30). Data from individual runoff plots show considerable deviation from this overall distribution, however, and plot 1 exhibits both the highest mean and standard deviation, while plot 4 shows the lowest. The explanation for these variations is to be found in an examination of diurnal and seasonal changes in concentrations.

If the snowbank is assumed to be chemically pure, solute concentrations measured at the surface flow collector depend only on conditions existing between the collector and the snow. Variables include (see Figure 37, p.61): (1) the travel time of the water between the snow edge and the collector, (2) the degree of flow mixing, (3) the solubility of the regolith, and (4) the temperature of the flow. Since (3) and (4) are likely to remain reasonably constant during the ablation period at a particular site, variation in concentrations is largely attributable to (1) and (2). The time taken by the flow to move to the sampling point on a particular slope depends on the distance between the snow edge and the collector, and flow velocity. The flow mixing depends on the flow type (laminar/transitional/turbulent) and the surface roughness.

On a diurnal scale, T.D.S. concentrations exhibit an inverse relationship with surface runoff, with peaks of concentrations occurring at "night", and flow peaks during the day (Figure 79A). Since the distance

Location	Years	Number of samples	Mean concentration (mg L <sup>-1</sup> )	Standard deviation (mg L <sup>-1</sup> )	Minimum concentration (mg L <sup>-1</sup> )	Maximum concentration (mg L <sup>-1</sup> )
Plot 1	1977-1979	46	216.7	95.3	65	450
Plot 2	1977-1979	58	58.3	34.2	13	156
Plot 3	1977-1979	121	93.0	85.5	2	488
Plot 4	1978-1979	28	52.2	32.3	20	124
All plots	1977-1979	253	103.0	91.6	2	488

TABLE 30: Total dissolved solids concentrations in samples of surface wash

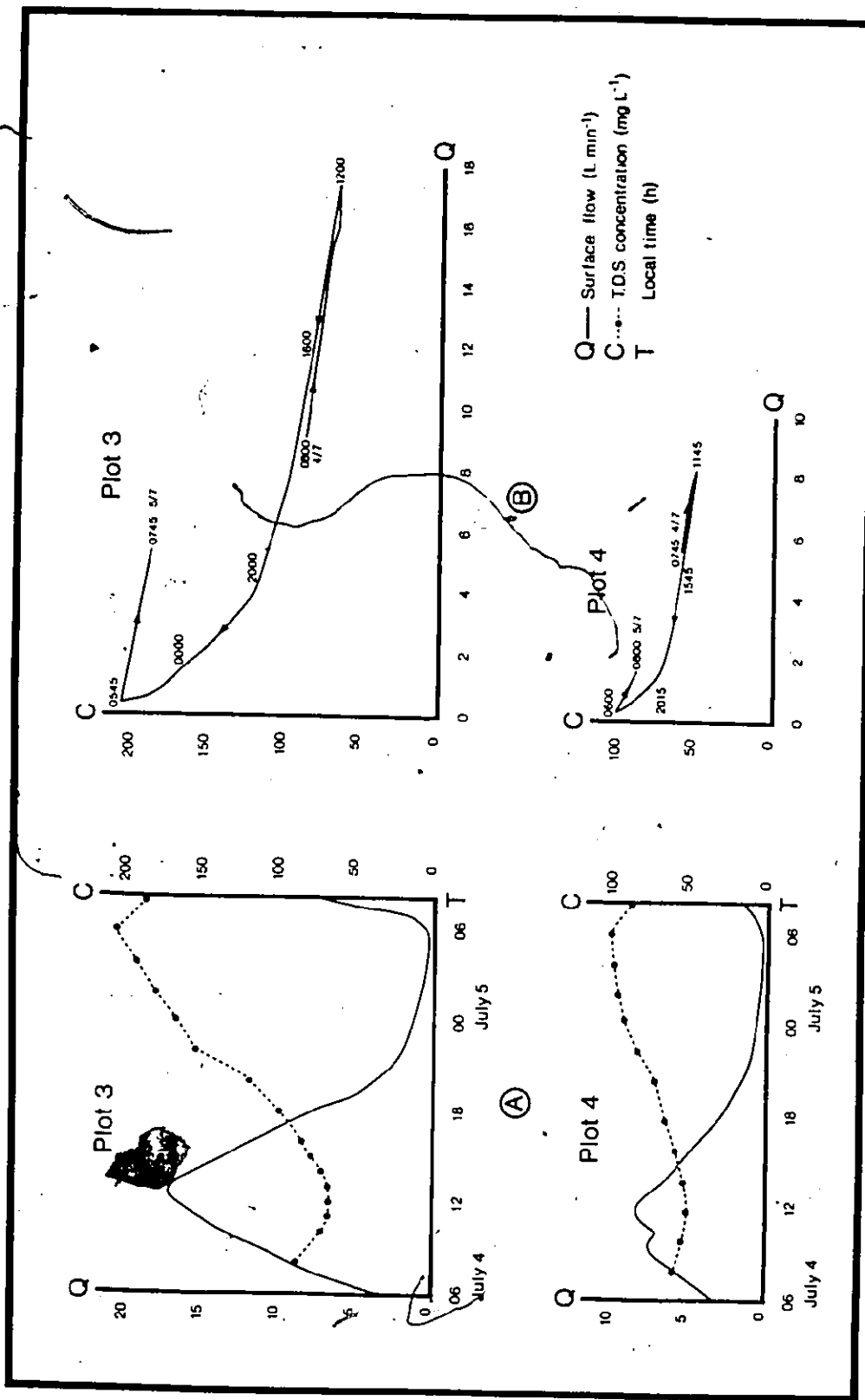


FIGURE 79 : SOLUTE CONCENTRATIONS IN SURFACE WASH, PLOTS 3 AND 4, JULY 4 - 5, 1978

A - Diurnal variation;

B - Correlation of solute concentrations with surface flow discharge.

between the snow edge and the sampling point usually varies little during a single day, changes in T.D.S. concentrations largely are attributable to velocity changes associated with the diurnal hydrographs. Mean velocity decreases at lower discharges and as a consequence, a greater concentration of solutes develops by the time the water leaves the plot. The relationship between discharge and T.D.S. values is likely to be nonlinear (Figure 79B). This is expected since bed roughness becomes increasingly important at low discharges making the relationship between surface flow velocity (i.e. time of contact) and discharge nonlinear. Moreover, increasing roughness at low discharges may cause greater flow mixing. For plots 3 and 4 (Figure 79B), the relationship between discharge and T.D.S. is best described by a semi-logarithmic curve. The hysteresis is thought to result from the retreat of each snow edge (0.5 m at plot 3 and 0.9 m at plot 4 on July 4-5, 1978).

On a longer time-scale, T.D.S. concentrations increase as the season progresses. While in part this change may be related to higher flow temperatures (increasing the speed of chemical reactions), the most obvious explanation is the greater distance travelled by the water over the ground surface to the collector, as the snowbank ablates.

The degree to which the major factors controlling the solute concentrations have been identified can be tested by multiple regression of T.D.S. concentrations against surface wash discharge and distance travelled by the flow. Both linear and logarithmic descriptions of discharge were attempted. The former provides the highest explanation of variance for plots 1 and 2; the latter is best for plots 3 and 4 (Table 31). Multiple correlation coefficients for the equations vary between +0.872 and +0.981, and all are significant at better than the 0.1% rejection level. Although the confidence limits at the 95% level

Plot	Year	N	Regression equation <sup>a</sup>	R	R <sup>2</sup>	95% confidence limits
1	1979	21	$Z = 122.05 - 14.0 X + 12.28Y$	+0.902	0.813	± 79.90
2	1977-1979	58	$Z = 0.01 - 3.78 X + 4.37Y$	+0.899	0.809	± 29.88
3	1978-1979	110	$Z = 64.39 - 37.41 \ln X + 8.22Y$	+0.872	0.760	± 83.13
4	1978-1979	28	$Z = 13.15 - 10.61 \ln X + 9.00Y$	+0.981	0.962	± 12.60
<sup>a</sup> X = Surface flow ( L min <sup>-1</sup> ) Y = Distance between snow edge and surface flow collector (m) Z = Total dissolved solids (mg L <sup>-1</sup> )						

TABLE 31: Best-fit multiple regression equations - T.D.S. concentrations in surface wash vs. surface flow discharge and distance between collector and snowbank edge

are large, the equations support the explanation of the T.D.S. variation. These formulae can be used to predict both the downslope increase in solutes away from the snowbank edge (Figure 80A), and the decrease in solute concentrations measured at a fixed point as discharge increases (Figure 80B).

The gradients in the regression equations show that solute concentrations increase away from the snow edge fastest at plot 1, the least vegetated site, and slowest at plot 2, the most vegetated. Since the regolith composition is believed to be similar at both sites, the action of vegetation in this case must be to reduce contact between part of the flow and the ground surface. It is hypothesised that a relatively slow-moving water layer stays in contact with the ground, and that a solute concentration gradient exists between this and faster-moving laminae above.

In summary, the solute concentrations in surface wash at the runoff plots are high but variable. Between 76% and 96% of variation in concentrations can be explained by the distance between sample point and snow edge, and by linear or logarithmic descriptions of discharge. The implications of these relationships are first, that a diurnal variation is present, with low concentrations at high flows and vice versa, and second, that concentrations increase during the season as the snow melts and the snowbank edge retreats upslope.

(d) Dissolved sediment amounts

The multiple regression equations relating T.D.S. concentrations to discharge and distance from the snow (Table 31), can be used as sediment rating curves for their respective plots. The total weight of dissolved solids can be calculated by multiplying concentrations derived for each 30-

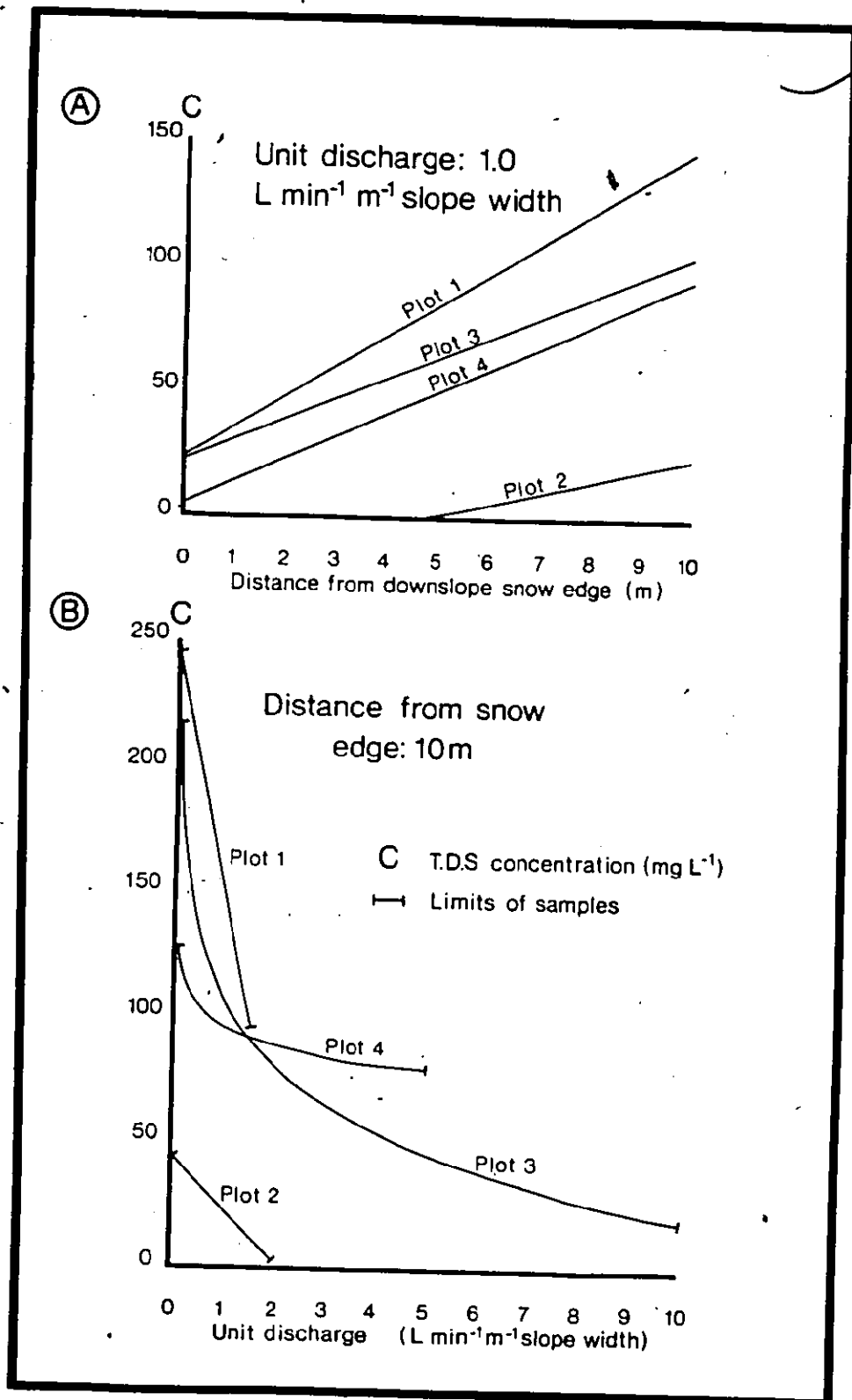


FIGURE 80 : SOLUTE CONCENTRATIONS IN SURFACE WASH, PLOTS 1 - 4  
 A - Variation downslope;  
 B - Variation with discharge.

minute period of the discharge record, by the discharge over that time. There is some uncertainty inherent in the values produced, however, because of the large values of the 95% confidence limits. The only case where it is not possible to derive totals in this way, is plot 1, for 1977. The break-up of the snow cover into a number of irregular masses during that year, precludes assessment of the distance to the snow edge. A sufficient number of conductivity measurements exist, however, for direct use to be made of the T.D.S. data.

The calculated totals of sediment lost by solution at the plots, vary from  $0.17 \text{ kg a}^{-1}$  to  $13.02 \pm 11.51 \text{ kg a}^{-1}$ , and mean solute concentrations range between  $44 \pm 30 \text{ mg L}^{-1}$  and  $252 \pm 80 \text{ mg L}^{-1}$  (Table 32). Concentrations are greatest at plot 1 and least at plot 2. In addition, there is some variation apparent between years at the same location. At plot 2, for example, yearly averages vary by up to 16% about the three-year mean, presumably in response to changes in snow ablation patterns and discharge rates.

Best-estimate values of weight loss  $\text{m}^{-2}$  for the full plot areas encompass two orders of magnitude, from  $0.38 \text{ g m}^{-2} \text{ a}^{-1}$  to  $49.64 \text{ g m}^{-2} \text{ a}^{-1}$ , with lowest values for plot 2 and highest for plot 3. The two-year average for the interfluvial site is higher than the three-year average for plot 2, primarily because average T.D.S. concentrations are four times as great at the former.

As in the case of suspended sediment, only certain areas of the plots are affected by solution loss. If total weight loss is distributed over these alone, rates  $\text{m}^{-2} \text{ a}^{-1}$  increase by two to three times, and proportionally, the rates for plots 3 and 4 become still higher (Table 32). Even greater spatial detail is possible, however, because information is

Plot	Year	Mean solute concentration <sup>a</sup> (mg L <sup>-1</sup> )	Total weight loss <sup>a</sup> (kg)	Weight loss over all the plot (g m <sup>-2</sup> )	Weight loss over affected <sup>a</sup> areas (g m <sup>-2</sup> )	Maximum loss <sup>a</sup> (g m <sup>-2</sup> )
1	1977 <sup>b</sup>	182 ± N.D.	0.17 ± N.D.	1.17 ± N.D.	2.92 ± N.D.	N.D.
	1979	252 ± 80	2.10 ± 0.67	4.80 ± 0.18	8.92 ± 2.82	21.02 ± 6.66
2	1977	44 ± 30	1.16 ± 0.79	2.20 ± 1.50	7.41 ± 5.04	17.78 ± 12.08
	1978	63 ± 30	0.74 ± 0.35	1.41 ± 0.67	3.83 ± 1.82	7.31 ± 3.47
	1979	51 ± 30	0.20 ± 0.12	0.38 ± 0.22	1.24 ± 0.73	2.94 ± 1.72
3	1978	94 ± 83	13.02 ± 11.51	49.64 ± 43.88	164.81 ± 145.70	612.53 ± 541.50
4	1978	70 ± 13	1.41 ± 0.25	21.07 ± 3.79	60.58 ± 10.90	102.86 ± 18.51
<p>a Including 95% confidence limits assessed from Table 31</p> <p>b Confidence limits not calculated due to different method of estimation</p>						

TABLE 32: Concentrations and weights of solutes removed from the plots in surface wash

available concerning the weight of sediment lost on each day, and the area downslope of the snow over the same period. Since there is no evidence of saturation concentrations being reached in the surface wash on the plots, the assumption can be made that each day's losses are spread relatively evenly over the exposed areas of the plot. Seasonal weight loss then, is greatest in the area immediately upslope of the surface flow collector, since this portion of the slope is exposed to all surface wash from the snowbank. Losses decrease upslope and are zero at the final runoff-producing position of the snowbank. For example, best-estimate losses immediately upslope of the collectors at plots 3 and 4 in 1978 are  $612.5 \text{ m}^{-2} \text{ a}^{-1}$  and  $102.9 \text{ g m}^{-2} \text{ a}^{-1}$  (Table 32), and become zero respectively 27 m and 13 m further upslope (Figure 81). The rate of change in loss rates downslope corresponds to periods of slow or rapid retreat of the snow edge. The former is shown by steeply sloping sections of the graph (e.g. June 14 - 24 at plot 3, Figure 81), and the latter by gently sloping sections (e.g. June 27 - 28 at plot 3). The maximum rates for these small areas of the plots are 2.7 and 3.7 times the rates for the areas of the plots affected by wash, and 4.4 and 12.3 times the rates averaged over the whole plot areas.

The rates of weight loss by solution can be converted to rates of surface lowering using the bulk density factor of  $0.67 \text{ Mg m}^{-3}$  (Table 33). Best-estimate values range from  $0.6 \times 10^{-3}$  to  $74.1 \times 10^{-3} \text{ mm a}^{-1}$  for the plots,  $1.9 \times 10^{-3}$  to  $246.0 \times 10^{-3} \text{ mm a}^{-1}$  for the affected areas and  $4.4 \times 10^{-3}$  to  $914.2 \times 10^{-3} \text{ mm a}^{-1}$  for the maximum rates over small areas. World data available for comparison with these figures are extremely limited since most measurements of net solution loss have been made by examination of the stream solute load. The latter includes solution accomplished by the stream itself, and

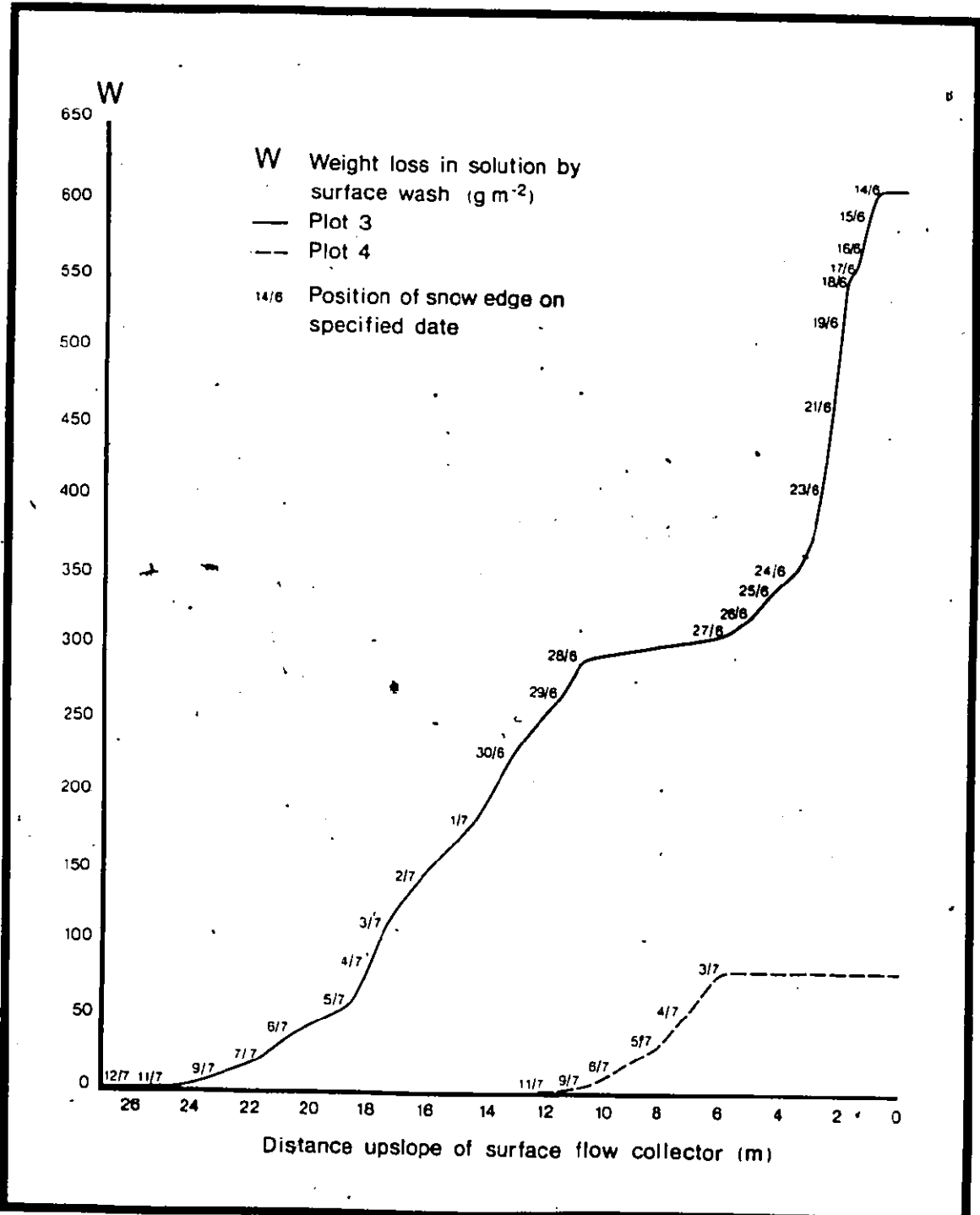


FIGURE 81 : VARIATION IN RATES OF WEIGHT LOSS BY SOLUTION IN SURFACE WASH, PLOTS 3 AND 4, 1978

Plot	Year	Surface lowering over all the plot <sup>a</sup> ( $10^{-3}$ mm a <sup>-1</sup> )	Surface lowering over affected area <sup>a</sup> ( $10^{-3}$ mm a <sup>-1</sup> )	Maximum rate of lowering ( $10^{-3}$ mm a <sup>-1</sup> )
1	1977	1.75 ± N.D.	4.36 ± N.D.	N.D.
	1979	7.16 ± 0.27	13.31 ± 4.21	31.37 ± 9.94
2	1977	3.28 ± 2.24	11.06 ± 7.52	26.54 ± 18.03
	1978	2.10 ± 1.00	5.72 ± 2.72	10.91 ± 5.18
	1979	0.57 ± 0.33	1.85 ± 1.09	4.39 ± 2.57
3	1978	74.09 ± 65.49	245.99 ± 217.46	914.22 ± 808.21
4	1978	31.45 ± 5.66	90.42 ± 16.27	153.52 ± 27.63

a Including 95% confidence limits assessed from Table 29.  
Bulk density assumed to be 0.67 Mg m<sup>-3</sup>

**TABLE 33:** Rates of surface lowering by solution through the activity of surface wash

the effects of subsurface wash in addition to surface wash solution. Comparison can be made, however, with a set of data from the alpine environment of the Colorado Front Range (Thorn, 1974). Thorn obtained a value for lowering due to solution by surface wash downslope of a large snowbank of  $6.7 \times 10^{-3} \text{ mm a}^{-1}$ . This figure falls within the range measured at the Thomsen River runoff plots, but is an order of magnitude less than the maximum recorded.

In summary, while some uncertainty exists due to large confidence limits, rates of surface lowering due to surface wash solution over plots 1 and 2 are similar to values recorded in a nonpermafrost alpine environment. Rates for plots 3 and 4, on the other hand, are up to ten times greater. Calculations for only those areas actually affected by wash increase the rates of surface lowering several times and raise the best estimate value at plot 3 to  $246 \times 10^{-3} \text{ mm a}^{-1}$ . At an even smaller scale, rates of surface lowering increase downslope of the final position of the snowbank and reach a best-estimate maximum of  $0.9 \text{ mm a}^{-1}$  at plot 3.

(e) Suspended, dissolved and aeolian sediment  
in surface wash

Comparison of the best-estimate concentrations of suspended and dissolved sediment in surface wash (Tables 26 and 32), reveals the dominance of the solute load. This is further illustrated by the rates of lowering by both processes over the areas of the plots affected by surface wash (Table 34). An important conclusion of this study is that lowering by solution in surface wash is 8 - 30 times more effective than the removal of sediment in suspension, and the latter constitutes only 3.4 - 10.5% of the total sediment load. Higher proportions of sediment are moved in suspension at the interfluvial site than at the snowbank -

Plot	Year	Best-estimate rate of lowering ( $10^{-3} \text{ mm a}^{-1}$ )		Lowering due to suspended sediment (%)		Lowering due to solution (%)	
1	1977	4.87	Average	10.5	Average	89.5	Average
	1979	14.43	9.65	7.8	9.2	92.2	90.8
2	1977	11.72	Average	5.6	Average	94.4	Average
	1978	5.96	6.54	5.96	4.7	96.0	95.3
	1979	1.94		4.6		95.4	
3	1978	254.59		3.4		96.6	
4	1978	95.61		5.4		94.6	

TABLE 34: Best-estimate rates of surface lowering by surface wash over affected areas of the plots

locations, probably as a result of the lack of protection by vegetation. It should be noted, however, that although the total rates of lowering are an order of magnitude greater at plots 3 and 4 than at plots 1 and 2, the latter are on slopes more typical of the Thomsen River area and their rates are probably the more usual.

Measurements of the concentration of aeolian detritus (mainly mineral material) on the snow surface at the runoff sites (Table 35) suggest that in some areas of the plots a build-up of material may be occurring. Maximum inputs of aeolian debris on the snow, although lower than others recorded in high Arctic areas (see p. 100), exceed some of the loss rates induced by surface wash. The data must be treated cautiously, however, for the reasons.

First, aeolian erosion may have occurred following ablation of the snow, removing material deposited during the spring. Second, aeolian sediment amounts varied across the plots, and the areas sampled were those possessing maximum concentrations. Average values probably were much lower. Overall, the data further serve to emphasize the importance of solution rather than suspended sediment removal in surface wash, since some of the suspended sediment movement recorded may not have resulted from ground erosion, but from the removal of recently deposited sediment.

#### Subsurface sediment transport

##### (a) Dissolved sediment concentrations

Samples of subsurface wash analysed for solute concentrations were collected at the exit of plot 4 (4 depths: 0.06 m, 0.16 m, 0.31 m and 0.46 m) and 130 m downslope of plot 3 (3 depths: 0.06 m, 0.16 m and 0.31 m). In addition, a number of samples were obtained from a collection gutter at a depth of 0.06 m level within the boundaries of plot 3.

Location	Aeolian detritus on snow surface ( $\text{g m}^{-2}$ )		
	June 8	June 11	July 9
Site 1	0.037	1.105	No snow left
Site 2	0.008	6.229	0.715
Site 3	0.010	0.106	1.293
Site 4	6.337	7.338	0.380

TABLE 35: Concentration of aeolian detritus on snow surface at the runoff sites, 1979.

Concentrations were determined by the same methods used for surface wash samples (see p. 223). No differences are extant in the trends of total hardness or its components in relation to specific conductance for surface and subsurface wash. The best record of concentrations in subsurface wash is from 1978 when 250 samples were collected over a period of 45 days.

Three major trends exist in the solute concentration data. First, a seasonal trend of (1) relatively low values at any one collection depth during the majority of the snowmelt period, (2) an increase in concentrations during the last 2 - 3 days of melt and the 10 days that follow, and (3) a plateau concentration level with some minor fluctuations (Figures 82 and 83). For example, typical values at the 0.16 m depth at site 3 during period (1) were 200 - 240 mg L<sup>-1</sup>. The plateau level (3) for the same gutter developed at very low discharges, was 400 - 500 mg L<sup>-1</sup>, and rising concentrations existed between the two states as discharges decreased.

Second, a trend exists for solute concentrations at any time to be greater at the lower collection levels. For example, analysis of the relationships between 31 pairs of samples collected from the 0.06 m and 0.16 m depths at site 3, and 36 sample pairs from the 0.16 m and 0.31 m depths (Figure 84), gives differences between the first two sample means of 28.2 mg L<sup>-1</sup>, and between the second two of 181.5 mg L<sup>-1</sup>. Correlation coefficients are +0.969 and +0.912 respectively. These high values illustrate the consistency of the trend, with only one point falling above the 1:1 line in the comparison of the 0.06 m and 0.16 m depths, and all falling above in the comparison of the 0.16 m and 0.31 m depths (Figure 84).

Third, the changes in concentrations at a particular depth are inversely, but nonlinearly, related to discharges. If T.D.S. concentrations

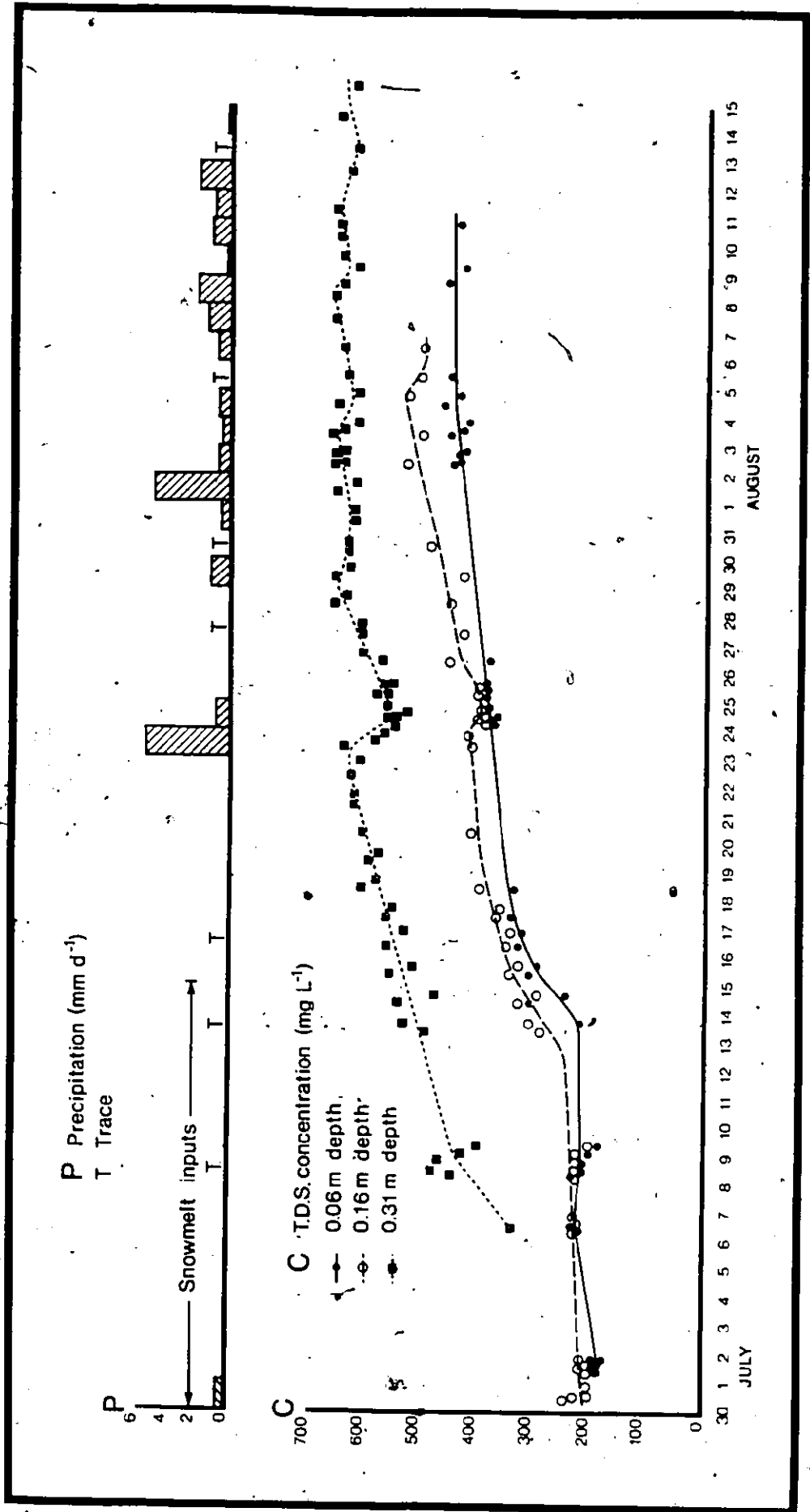
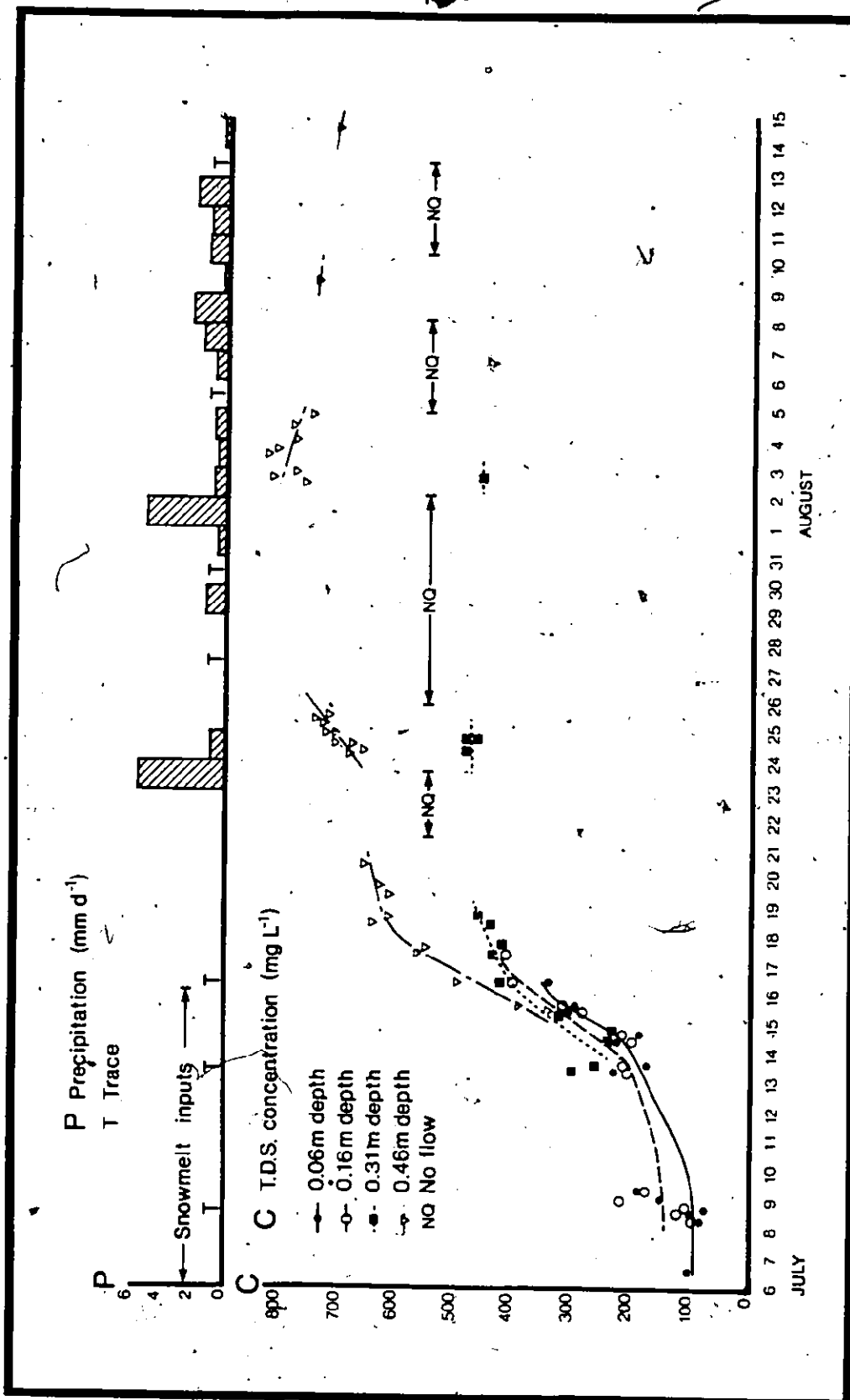


FIGURE 82 : SEASONAL VARIATION OF T.D.S. CONCENTRATIONS  
IN SUBSURFACE WASH, SITE 3, 1978  
Note: lines are eye-determined best-fit.



**FIGURE 83 : SEASONAL VARIATION OF T.D.S. CONCENTRATIONS IN SUBSURFACE WASH, SITE 4, 1978**  
 Note: lines are eye-determined best-fit.

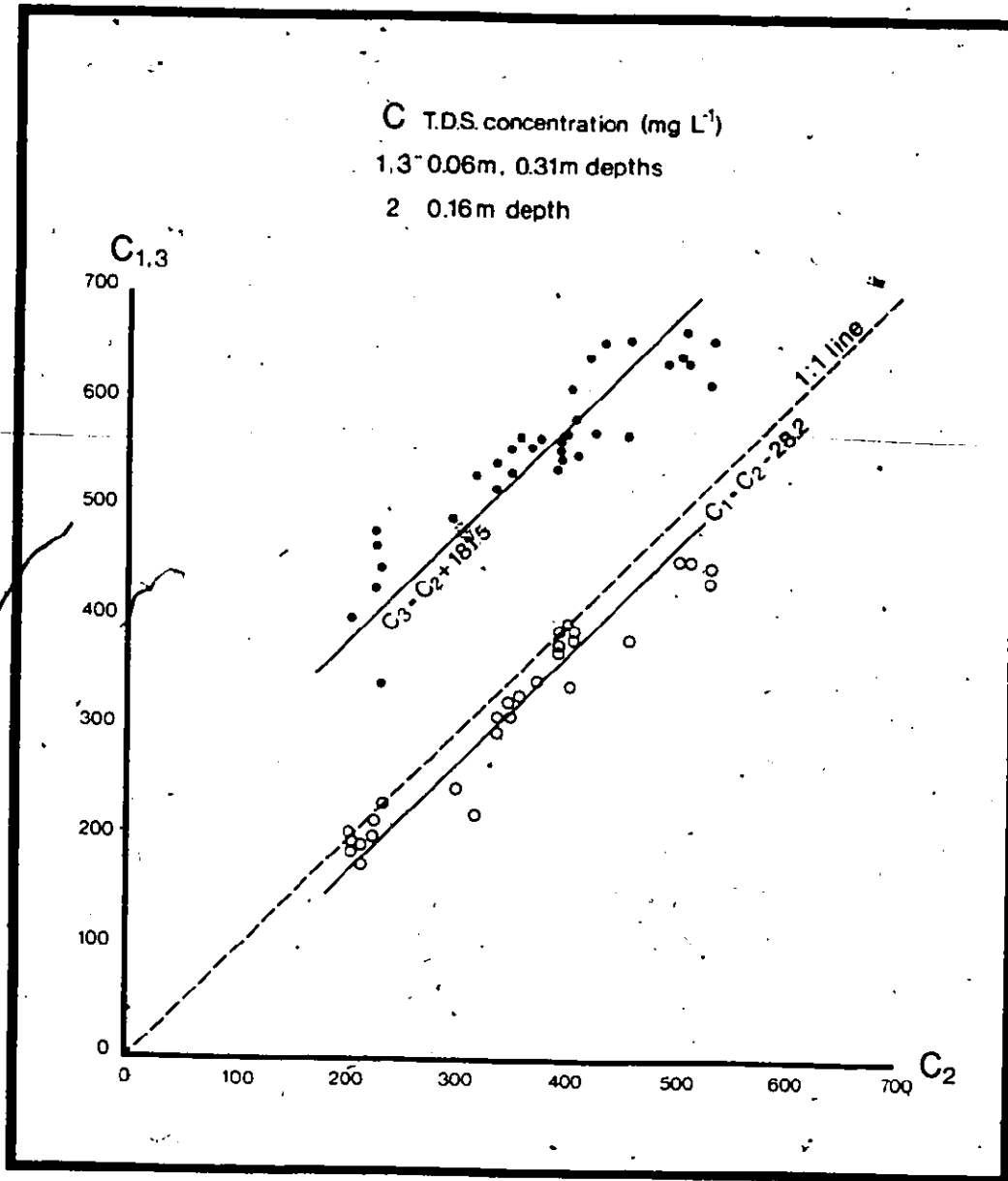


FIGURE 84 : CORRELATIONS BETWEEN T.D.S. CONCENTRATIONS IN SUBSURFACE WASH AT THREE DEPTHS, SITE 3

are plotted against the logarithms of discharges for each collection depth a series of lines is produced. All lines apparently possess a section of zero gradient at very low discharges and a section of rapidly declining concentrations with increasing discharges (Figures 85 and 86). In addition, some collection depths appear to have a second section of near-zero gradient at high discharges.

Explanation of the three identified trends is forthcoming from existing theory. The key factors in the variation of concentrations are (Figure 37, p. 61) (1) the time of contact between regolith and water, (2) the composition of the regolith, and (3) the rate of chemical reactions. The relatively low concentrations present during the bulk of the snowmelt period, correspond to discharge values that vary little (see p. 188) and by inference, to flow speeds and times of contact that are almost constant. The moisture supply is continuous, so that water is moved through the regolith and is prevented from achieving high solute concentrations by short residence times. Some small increases during this stage probably result from lengthening travel times caused by the upslope retreat of the snow edge. Towards the end of melt, the duration of recession flow from the snowbank becomes shorter and the moisture supply becomes intermittent. Discharges at each depth decline, and as increasing proportions of the flow occur through the soil pores rather than through fissures and larger voids, residence times increase. Opportunity now exists for minerals with lower dissolution rates to pass into solution, and for other minerals to reach higher solute concentrations. When the last of the snowbank ablates, the moisture supply is cut off. If there is no precipitation, it can be inferred that all water samples subsequently collected have residence times equal to, or greater than, the number of days since the cessation of melt. Eventually, 10 - 15 days after the end of melt, the

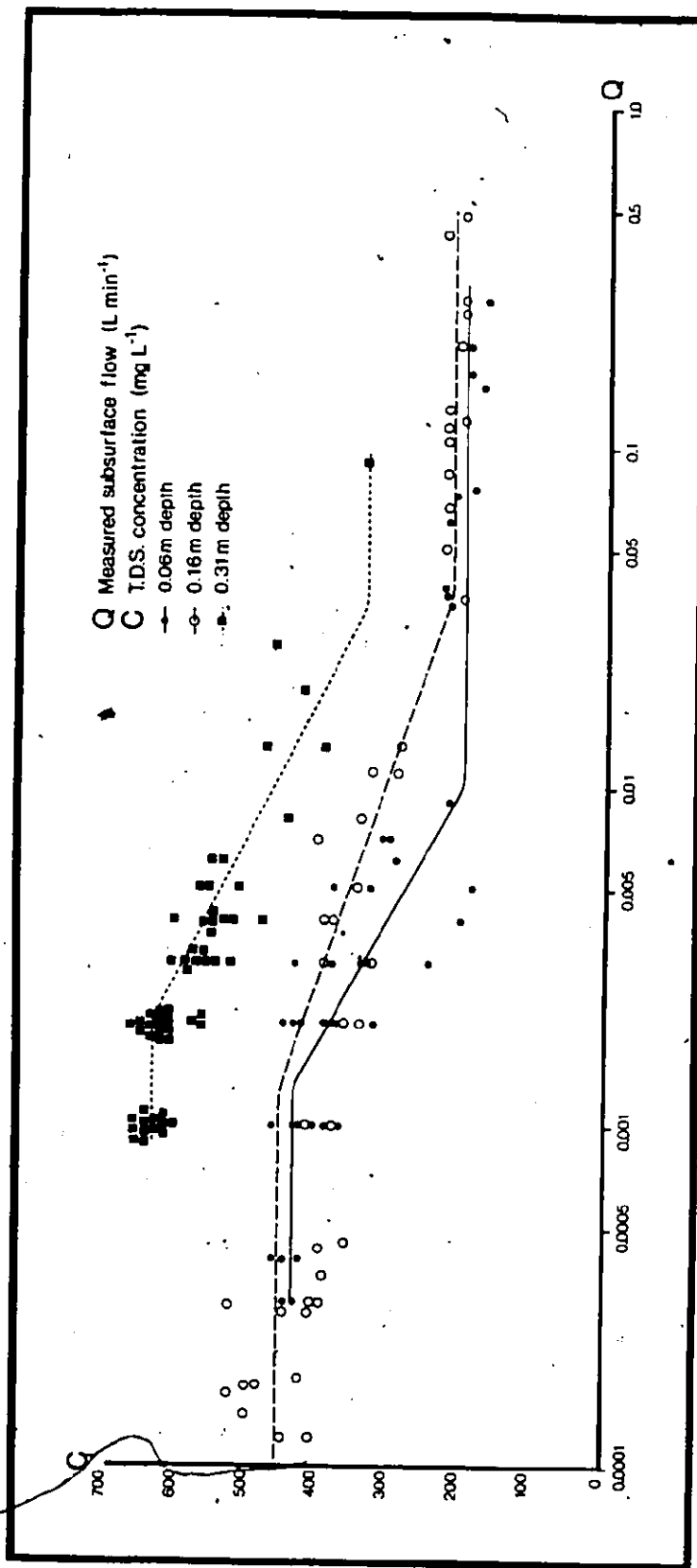


FIGURE 85 : VARIATION OF T.D.S. CONCENTRATIONS IN SUBSURFACE

WASH WITH DISCHARGE, SITE 3, 1978

Note: lines are eye-determined best-fit.

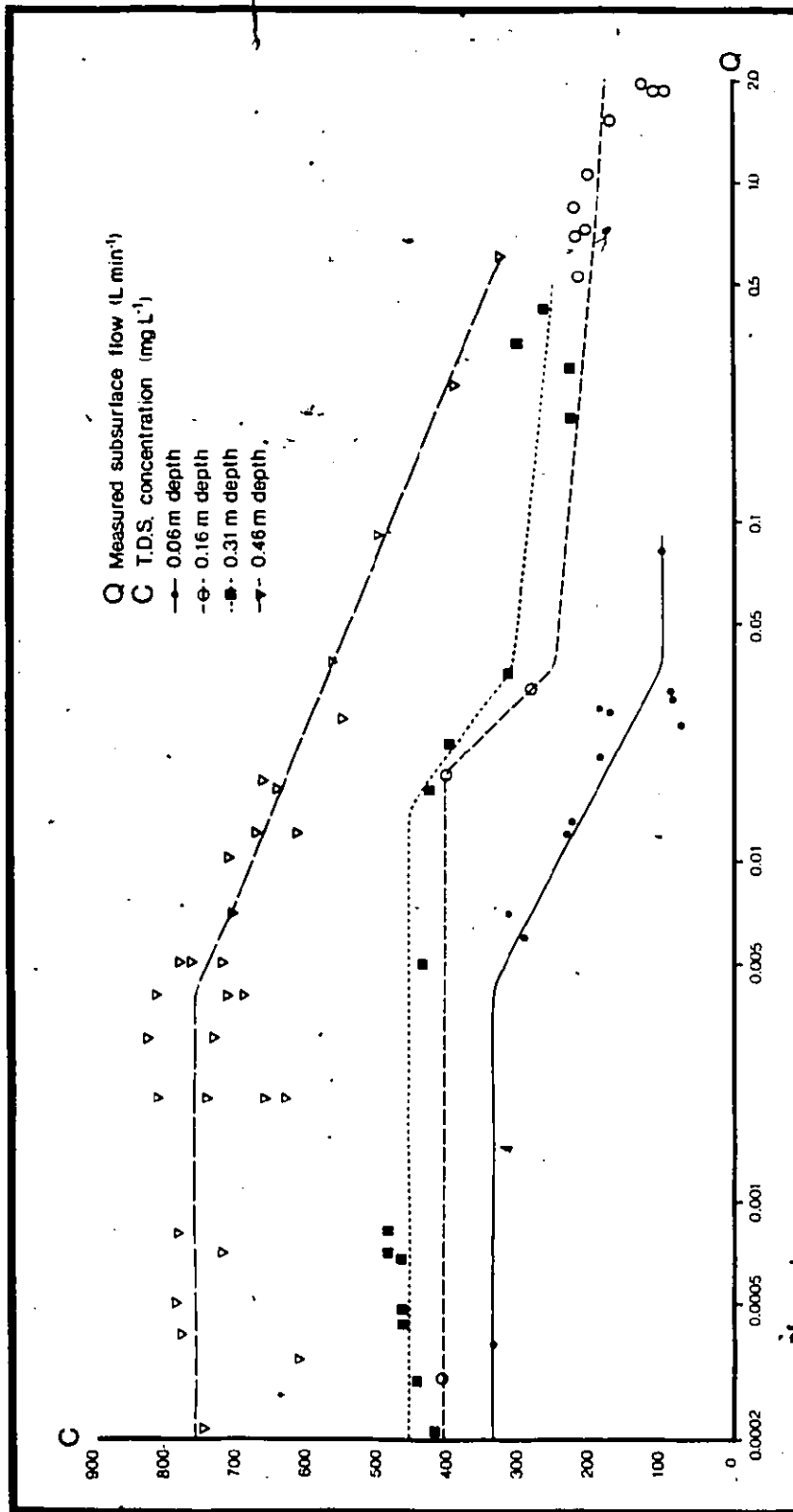


FIGURE 86 : VARIATION OF T.D.S. CONCENTRATIONS IN SUBSURFACE WASH WITH DISCHARGE, SITE 4, 1978  
Note: lines are eye-determined best-fit.

solutes in the water reach equilibrium with the soil minerals and further increases in T.D.S. values are prevented. If precipitation does occur, some decline in concentrations may be induced (e.g. July 23, 1978, Figure 82) as discharges increase (see Figure 68), p. 191). Since concentrations do not decline to the values existing during snowmelt, however, much of this subsurface discharge is probably soil water displaced by the precipitation which has been in contact with the regolith for several days. As a result, the relationships between subsurface discharge and T.D.S. concentrations (Figures 85 and 86), are nonlinear and exhibit considerable scatter, because discharge is acting as an analogue of residence time and after precipitation events is not a very accurate indicator of this variable.

The differences in T.D.S. concentrations at different levels but in the same location at the same discharge, can be hypothesised to result from (1) longer flow paths at depth, and hence greater residence times, (2) different minerals within the regolith, and (3) a combination of (1) and (2). The last appears to be the most probable explanation and indeed it is possible to dismiss (1) as the sole explanation because of the continued presence of the vertical concentration gradients even after very long residence times at all depths.

Downslope changes in concentrations at particular depths can be inferred from the principles and observations outlined above. During the period of continuous moisture addition, residence times and solute concentrations increase downslope. As flow becomes intermittent, the slope drains from the top first, and with lower discharges in the upper areas, it appears that the pattern may reverse with higher concentrations upslope. Eventually, when all residence times are large and equilibrium is reached, little or no change in concentrations downslope is likely. Limited data from samples

collected at the 0.06 m depth within plot 3, and 130 m further downslope, support these inferences. They illustrate the first two stages, with the crossover between downslope and upslope gradients apparently occurring on July 9, in 1978 (Figure 87). Samples for the expected zero gradient were not collected as subsurface volumes became too small to test.

The concentrations measured in subsurface flow can be compared with others from permafrost areas. The maximum T.D.S. value of 827 mg L<sup>-1</sup> recorded at plot 4 exceeds others in the literature. The total hardness value of 365 mg L<sup>-1</sup> as CaCO<sub>3</sub> for the same sample is greater than others from areas of carbonate rocks (Table 6, p. 107). These high values may account for the presence of the plateau concentrations, rather than a general trend for increases throughout the summer in response to higher temperatures and biogenic CO<sub>2</sub> production (e.g. Woo and Marsh, 1977).

In summary, concentrations of solutes in subsurface wash are at their lowest during periods of continuous snowmelt. Towards the end of melt and following its completion, concentrations increase in response to longer residence times, consequent on the decline of discharges and flow velocities. After 10 - 15 days, equilibrium concentrations are reached and further increases in residence times or decreases in flow velocities have no effect. Precipitation events during this period result in increased discharges and reduced solute concentrations, but the latter still remain greater than others reported in the literature. Concentrations vary spatially on a slope, tending to increase with depth at a particular location, and change from downslope increases during the melt season, to potential increases or uniformity after the end of melt.

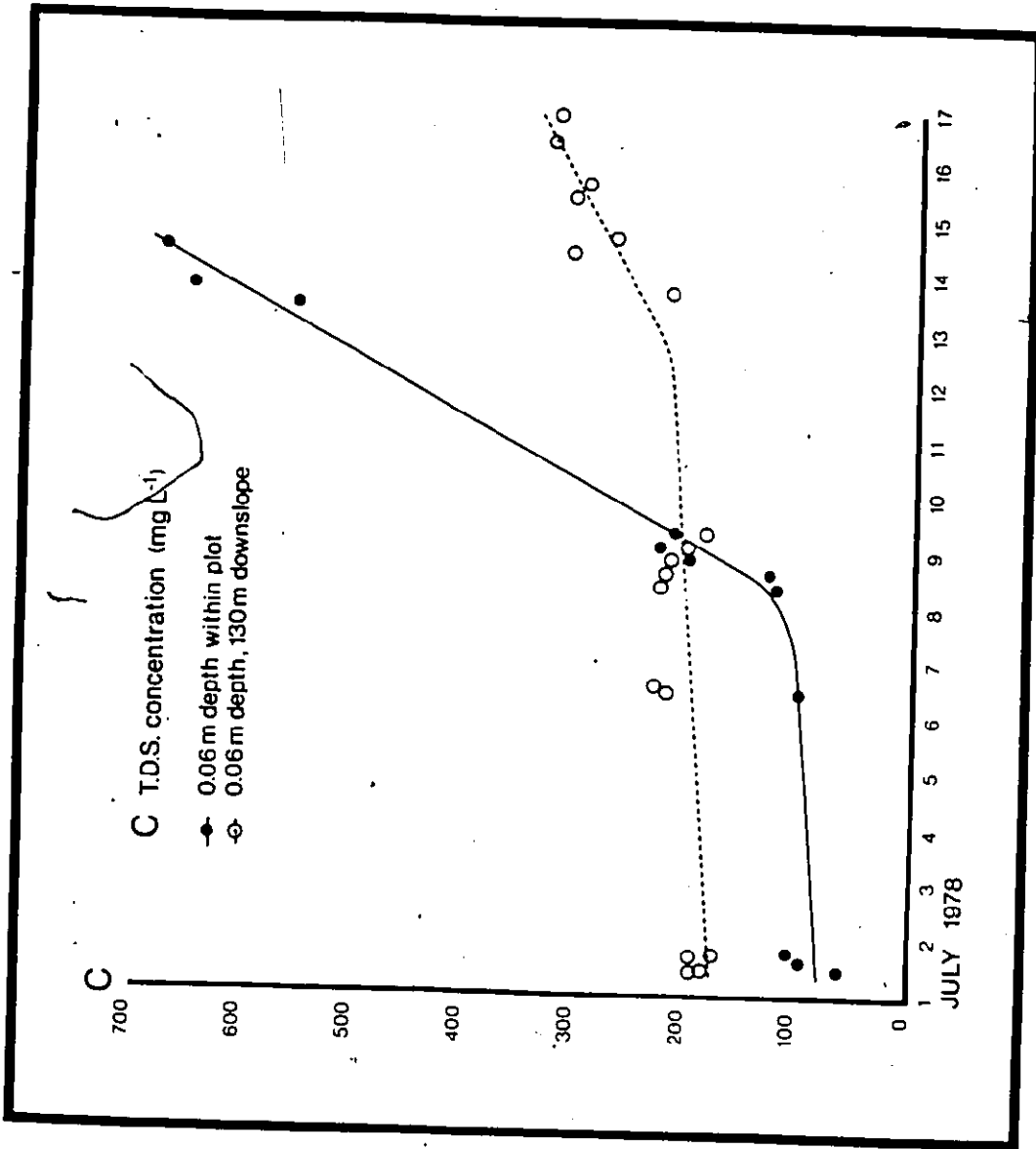


FIGURE 87 : DOWNSLOPE VARIATION IN T.D.S. CONCENTRATIONS IN  
SUBSURFACE WASH, SITE 3, 1978

Note: lines are eye-determined best-fit.

(b) Dissolved sediment amounts

Subsurface flow was not measured continuously at the runoff sites except after the end of snowmelt when discharges were low. For this reason, calculation of the weight of sediment removed in solution by this process can only be approximate. It requires estimates of the total volume of water leaving the plot as subsurface flow, and the average concentration of solutes in the discharge (assuming that concentrations in the snow are zero).

The total volume of subsurface flow during the snowmelt period can be calculated for plots 1, 2 and 4 from residuals in the water balance equations (Tables 11, 12 and 13, Chapter 5). At plot 3, however, it is known that the 1978 residual is too small, since some daily surface runoff coefficients exceed unity. A better estimate of the subsurface flow volume can be obtained from the flow rates shown in Figure 66 (p. 186).

The bulk of snowmelt-induced subsurface flow occurs at high discharges and relatively low T.D.S. concentrations. Samples used for estimating an average concentration level, therefore, do not include those collected during desaturation of the active layer during the last 2 - 3 days of melt, nor any collected thereafter. Data used originate from plot 1 in 1977, plot 3 in 1978 and plot 4 in 1978 and 1979. The average T.D.S. figure obtained is  $156 \text{ mg L}^{-1}$  (standard deviation of  $51 \text{ mg L}^{-1}$ ,  $n = 29$ ) and it is assumed that this can be applied to all the plots. An additional amount of sediment is removed from the plots in the post-melt period, but measurements indicate a very rapid decline in discharges following the end of melt (see p. 191), and even at high concentrations, the sediment transport rate must be small. Moreover, discharge data do not exist for all plots and it is impossible to ascertain the contributory widths for those

available. Consequently, solute totals include only those removed during the melt period.

The weight of sediment removed, calculated by multiplying flow volume by average concentration, ranges from  $4.12 \text{ g m}^{-2} \text{ a}^{-1}$  to  $43.26 \text{ g m}^{-2} \text{ a}^{-1}$  at the plots (Table 36), with the lowest value for the interfluvial site and the highest for the location of the largest snow-bank. Weight loss over the three-year period at plot 2 is similar for 1977 and 1978, but much less in 1979 when considerable volumes of water left the plot by flowing through the snow. The same methods employed for suspended and dissolved sediment in surface wash, can be used to convert the weight losses over the plot into losses over the areas affected by wash (those downslope of the final position of the snow edge). This conversion increases the range of weight losses to between  $7.64 \text{ g m}^{-2} \text{ a}^{-1}$  and  $143.72 \text{ g m}^{-2} \text{ a}^{-1}$ , but the relative order of the plots remains the same (Table 36).

Conversion of the weight loss figures to surface lowering requires an estimate of the dry bulk density of the material. The density of the regolith is greater at depth than at the surface, but less than that of solid rock. A value of  $1.5 \text{ Mg m}^{-3}$ , an average of 50 frozen samples with less than 10% visible ice, taken from the top 1 m of boreholes on Melville Island (EBA Engineering Consultants Ltd., 1978) was used for the calculations.

Results for lowering over the whole plots vary by a factor of ten and over the affected areas by a factor of twenty. Plot 1 in 1979 possesses the minimum rate over the affected area of  $5.09 \times 10^{-3} \text{ mm a}^{-1}$ , while plot 3 in 1978 shows the highest rate with 95.81

Plot	Year	Weight loss over all the plot ( $\text{g m}^{-2} \text{ a}^{-1}$ )	Weight loss over affected area ( $\text{g m}^{-2} \text{ a}^{-1}$ )	Surface lowering over all the plot ( $10^{-3} \text{ mm a}^{-1}$ )	Surface lowering over affected area ( $10^{-3} \text{ mm a}^{-1}$ )
1	1977	6.28	15.70	4.19	10.50
	1979	4.12	7.64	2.75	5.09
2	1977	32.95	110.94	21.97	73.96
	1978	25.73	69.92	17.15	46.61
	1979	10.31	33.80	6.87	22.53
3	1978	43.26	143.72	28.84	95.81
	1978	29.43	84.57	18.99	56.38
a	Bulk density assumed to be $1.5 \text{ Mg m}^{-3}$				

TABLE 36: Rate of weight loss and surface lowering  
due to solution by subsurface wash

$\times 10^{-3}$  mm a<sup>-1</sup>. Clearly, the interfluvial location is subject to the lowest rates of loss by this process, and large snowbank sites to the highest rates.

### Slopewash sediment transport

#### (a) Transport in solution

Total slopewash solution can be determined by adding together figures for its surface and subsurface components (Tables 33 and 36). Rates averaged over the plots range from  $7.44 \times 10^{-3}$  to  $102.93 \times 10^{-3}$  mm a<sup>-1</sup>, and for affected areas are  $24.38 \times 10^{-3}$  to  $341.80 \times 10^{-3}$  mm a<sup>-1</sup> (Table 37). A major conclusion of the study is that denudation accomplished by solution in slopewash is 16 to 250 times more important than the removal of suspended sediment particles by surface wash (Table 37). The difference in rates is most pronounced at plot 2 and least at plot 1, with ratios at the largest snowbank plots of 28 - 40 times.

The solution rates obtained in this study (Table 37) can be compared with others obtained by calculating loads in rivers (i.e. rates averaged over whole catchments and including the solution activity of the river itself). Global values in Young (1974, p. 70) cover two orders of magnitude, from  $1 \times 10^{-3}$  to  $100 \times 10^{-3}$  mm a<sup>-1</sup>, a range similar to that exhibited at the runoff plots. Since some parts of the study area produce no slopewash it is likely that the average rate of solution in the Thomsen River area lies in the middle of the world range.

A comparison with rates measured by Rapp (1960) in the polar environment of the Karkevagge, northern Lapland indicates that solution is of equal or even greater importance in central Banks Island. Rapp (1960) recorded rates of  $5 \times 10^{-3}$  to  $10 \times 10^{-3}$  mm a<sup>-1</sup> which are of the same order as those for plots 1 and 2, but less than one tenth of the rate

Plot	Year	Surface lowering by solution over all the plot ( $10^{-3}$ mm a $^{-1}$ )	Surface lowering by solution over affected area ( $10^{-3}$ mm a $^{-1}$ )	Ratio of lowering by solution to suspended sediment removal <sup>a</sup>
1	1977	5.94	14.86	28:1
	1979	9.91	18.40	16:1
2	1977	25.25	85.02	133:1
	1978	19.25	52.33	214:1
	1979	7.44	24.38	250:1
3	1978	102.93	341.80	40:1
4	1978	50.44	146.80	28:1
<p>a ratios equal for all the plot and affected areas  Source: Tables 28, 33, 36</p>				

TABLE 37: Best-estimate rates of lowering by solution in  
slopewash over all and affected areas of the runoff plots

at plot 3. Rapp (1960) concluded that solution transport was the most important slope process operating in the Karkevagge and it is possible that this is also the situation in the study area. The rates recorded at Thomsen River lend credence to the view that solution activity is very important in periglacial environments and run contrary to results obtained on Bathurst Island where dissolved load in streamflow constituted only 14% of the total annual sediment load (Wedel et. al., 1978).

(b) The relative importance of surface and subsurface wash transport

The total erosive action of slopewash can be assessed by adding together all the best-estimate figures for weight loss and surface lowering. A number of trends are visible in the data (Tables 38 and 39).

Total weight loss is greatest at the location of the largest snowbank,  $94.7 \text{ g m}^{-2} \text{ a}^{-1}$  over plot 3, and  $314.3 \text{ g m}^{-2} \text{ a}^{-1}$  over its affected area. Losses become smaller as snowbank size decreases and are least at the interfluvial site - a two-year average of  $8.5 \text{ g m}^{-2} \text{ a}^{-1}$  over the whole of plot 1 and  $18.2 \text{ g m}^{-2} \text{ a}^{-1}$  over its affected area.

Among the three snowbank locations, the proportion of total weight loss achieved through the action of surface wash increases with initial snowbank size. Only plot 3 has a greater amount of slopewash erosion accomplished by surface wash (54.3%). At the interfluvial location, the pattern reverses itself between the two years. This may be the result of measurements being made at different plots on the same slope, or may indicate a location sensitive to winter snow distribution and melt meteorological conditions. At plot 2, in contrast, the trend is consistent over the three-year period.

PLOT	YEAR	WEIGHT LOSS OVER ALL THE PLOT ( $\text{g m}^{-2} \text{ a}^{-1}$ )			WEIGHT LOSS OVER AREAS AFFECTED BY WASH ( $\text{g m}^{-2} \text{ a}^{-1}$ )		
		Surface wash	Subsurface wash	Total	Surface wash	Subsurface wash	Total
1	1977	1.3	6.3	7.6	3.3	15.7	19.0
	1979	5.2	4.1	9.3	9.7	7.6	17.3
2	1977	2.3	33.0	35.3	7.9	110.9	118.8
	1978	1.5	25.7	27.2	4.0	69.9	73.9
	1979	0.4	10.3	10.7	1.3	33.8	35.1
3	1978	51.4	43.3	94.7	170.6	143.7	314.3
4	1978	22.3	29.4	51.7	64.1	84.6	148.7

TABLE 3B: Rates of weight loss due to slopewash

PLOT	YEAR	LOWERING OVER ALL THE PLOT ( $10^{-3}$ mm a $^{-1}$ )			LOWERING OVER AREAS AFFECTED ( $10^{-3}$ mm a $^{-1}$ )		
		Surface wash	Subsurface wash	Total	Surface wash	Subsurface wash	Total
1	1977	2.0	4.2	6.2	4.9	10.5	15.4
	1979	7.8	2.8	10.6	14.4	5.1	19.5
2	1977	3.5	22.0	25.5	11.7	74.0	85.7
	1978	2.2	17.2	19.4	6.0	46.6	52.6
	1979	0.6	6.9	7.5	1.9	22.5	24.4
3	1978	76.6	28.8	105.4	254.6	95.8	350.4
4	1978	33.3	19.0	52.3	95.6	56.4	152.0

TABLE 39: Rates of surface lowering due to slopewash

The trends in the surface lowering data (Table 39) are the same as for weight loss, but the relative importance of subsurface action is reduced because of the larger density conversion factor. As a result, in addition to plot 3, plot 4 experiences more lowering by surface wash than by subsurface wash. Rates of lowering range from a two-year average of  $8.4 \times 10^{-3} \text{ mm a}^{-1}$  for plot 1, to  $105.4 \times 10^{-3} \text{ mm a}^{-1}$  for plot 3. Within the plots, however, rates on the affected areas are 2 - 3 times greater, and within these, selected areas are lowered still faster (e.g.  $> 0.9 \text{ mm a}^{-1}$  at the lowest part of plot 3 (see Table 33, p. 238)). Inter-year variation about the three-year mean at plot 2 is  $\pm 44.7\%$  for weight loss over the whole plot, and  $\pm 46.0\%$  for surface lowering. It appears likely that similar variation would be observed at the other plots and long-term rates could be more or less than those observed.

In view of the greater proportions of the study area represented by plots 1 and 2 than by plots 3 and 4, two major conclusions can be reached concerning the efficacy of sediment transport by surface and subsurface wash in the Thomsen River region. First, at most locations, subsurface wash is more important as an agent of denudation than surface wash. Second, this importance arises from the relative insignificance of particle sediment movement in surface wash and the overall importance of solution in both surface and subsurface wash. (The results of the study support the findings of Rapp (1960) and emphasise the role of solution processes in periglacial areas.

CHAPTER 7

CONCLUSIONS


The conclusions reached in this study may be summarized from both geomorphic and hydrologic viewpoints. They also suggest lines of approach for future research.

#### Geomorphic aspects of slopewash

Comprehensive studies of both dissolved and suspended sediment in surface and subsurface wash have not been undertaken previously in permafrost areas. Certain elements of the slopewash system have been investigated, however, and the results of the present study generally support conclusions reached elsewhere regarding the magnitude of slopewash erosion in periglacial areas.

Measurements at the four runoff plots indicate the existence of very low concentrations of suspended sediment in snowmelt-induced surface wash. The maximum value measured was at the interfluvial site where an incomplete vegetation cover enabled a concentration of  $70 \text{ mg L}^{-1}$  to develop. The absence (with one exception) of diurnal concentration cycles corresponding to daily snowmelt hydrographs, suggests that sediment removal is limited by sediment availability, not flow transport capacity.

Rates of sediment removal at the four plots ranged from  $0.02$  to  $1.74 \text{ g m}^{-2} \text{ a}^{-1}$ . Rates of removal and lowering were greatest at plot 3, the site of the largest snowbank, and least at plot 2, the location of a smaller snowbank. The values are comparable to, or lower than those measured by Jahn (1961) on Spitsbergen, and both sets are among the lowest in the world. Jahn's results may be typical for rocky or talus slopes which are unable to support a closed vegetation cover. The Thomsen River figures likely are more representative of permafrost areas covered by fine-grained surficial deposits where the chief factors determining plant growth are



moisture supply and duration of the growing season. The influence of moisture supply on plant cover affects sediment removal, and probably is the cause of lower denudation rates being measured at plot 2, than at the interfluve plot.

It can be inferred from these relationships that on similar terrain, rates of lowering attributable to suspended sediment removal in surface wash downslope of large snowbanks, are likely to be greater in the polar desert areas as a result of reduced vegetation cover on the snowbed. In the tundra zone south of the study area, increased vegetation growth probably causes a decline in erosion values, in spite of greater winter snowfalls and higher ablation rates. If correct, this inference qualifies the suggestion of French (1978, p. 142) that maximum slopewash denudation rates occur in the transitional zone between tundra and polar desert.

Surface lowering effected by solution was evaluated separately for surface and subsurface wash. Solute concentrations reached a maximum value of  $488 \text{ mg L}^{-1}$  in the former, and  $827 \text{ mg L}^{-1}$  in the latter. Concentrations in surface wash were inversely related to surface flow discharge and to distance downslope of the ablating snow. Solute levels in subsurface wash were lowest during snowmelt and increased nonlinearly during declining subsurface flow following the end of the ablation season. The high concentrations of solutes, up to an order of magnitude greater than the suspended sediment maxima, indicate the importance of solution processes. Moreover, since both surface and subsurface flow are involved, the relative significance of solution is further increased.

Rates of weight loss in solution were highest at the largest snowbank site and least at the interfluve site. In surface wash the rates corresponded to estimates of surface lowering of between  $0.57 \times 10^{-3}$  and  $74.09 \times 10^{-3} \text{ mm a}^{-1}$ . In subsurface wash, weight losses for

the snowmelt period indicated lowering rates of between  $2.75 \times 10^{-3}$  and  $28.84 \times 10^{-3} \text{ mm a}^{-1}$ . If these figures are combined, the true dominance of solution as a denudational agent is revealed. Total surface lowering by solution exceeded suspended sediment removal by 16 to 250 times. These values represent averages for the total plot areas. At the micro-scale, annual solution loss increases downslope from zero at the final runoff position of the snow edge, to a maximum at the first point on the slope affected by all the snowbank meltwater. The rate of lowering by solution in surface wash alone, exceeded 0.9 mm at this point on plot 3 in 1978.

Results obtained in this study support Rapp's (1960) conclusions for the Karkevagge, northern Scandinavia. Rates of lowering by solution were found to be the most important denudational process in that area. While major differences exist between the Karkevagge and the Thomsen River area, in view of higher loss rates measured on the plots, it is possible that similar conclusions regarding the efficacy of solution processes apply to central Banks Island.

#### Slopewash hydrology

The main aim of the hydrological investigations was a detailed examination of slope hydrology in a continuous permafrost area. In the past, the slope hydrological cycle has been subsumed in catchment studies and there is a paucity of comparative data.

It proved possible to estimate the water balance of an upland area of approximately  $0.5 \text{ km}^2$ , within which the four study plots were located. In this area, approximately 92% of the water input was derived from snowmelt, and only 8% came from summer rainfall. Of this available water,

9% evaporated from snow, 22% was lost by evapotranspiration, and the remaining 69% was estimated to have left the area in stream runoff. In contrast, losses to the atmosphere at the runoff plots were generally lower, and consequently, values of runoff contribution were higher than for the upland as a whole. For example, 96% of the water inputs at the largest snowbank are thought to have reached the stream channel in 1978. The higher figures indicate that all four study plots make average to above-average contributions to streamflow, and imply that other terrain units in the upland (e.g. ridges) lose all their water inputs to evaporation, and fail to augment stream discharge.

Snowmelt-induced surface runoff coefficients at the runoff plots varied at the annual and daily time-scales, and from site to site. This variability is a key feature of the slope hydrological regime in a permafrost area and even after the period of study, generalisations must be regarded as tentative. In the three years of records from plot 2, for example, between 2% and 26% of the snow initially present on the plot left as surface flow. In contrast, the surface runoff coefficient at plot 3 in 1978 reached 74%.

The variability at the annual time-scale is caused by changes in the winter snow distribution at the sites, and by inconstant meteorological conditions during the ablation season. The former influence the pattern of snowmelt on the plot and the path by which water leaves it. The latter affect surface runoff coefficients through the timing of periods of high and low energy inputs in relation to the size of the remaining snowbank. Over short time periods a strong positive relationship existed at all four plots between daily meltwater production and corresponding surface flow discharge. In addition, at two of the plots a positive relationship existed between

surface runoff and antecedent moisture. At plot 4, frost-table depth was a negative influence on surface flow discharge. All these parameters vary from one ablation season to the next, and from site to site and account for the lack of consistency in the plot surface runoff coefficients.

The frequency of surface flow at the sites also was variable. While 32 consecutive days of surface flow were recorded at plot 3 in 1978, only 4 days of flow were observed at plot 1 in 1977 and 1979. These figures affect surface runoff coefficients because the more rapid the snowmelt, the higher the runoff coefficients, since threshold subsurface and evaporative losses are exceeded by wider margins. This can be illustrated by a comparison with data from the discontinuous permafrost zone. Landals and Gill (1973) measured runoff coefficients averaging 55% near Yellowknife, N.W.T. (8 plot years, ranging from 1% to 82%), a figure substantially higher than the 25% average recorded at Thomsen River (7 plot years, ranging from 2% to 74%). This difference is thought to be the result of lower ablation rates due to reduced energy inputs in the more northerly location. If this is a general trend, it implies that in higher latitudes than the Thomsen River area, under similar site conditions, surface runoff coefficients are still lower.

Subsurface flow at the runoff plots varied in importance as a route for water loss according to snowbank size. In absolute terms it was most important at the largest snowbank location, but in percentage terms it was least important at that site. Limited measurements suggest that it is of minor significance at such sites during most of the ablation season when surface flow is continuous. As surface flow becomes intermittent during the last few days of melt, however, the relative importance of subsurface flow increases. The decline of subsurface flow discharge occurs very rapidly

after the end of snowmelt and values drop by two or more orders of magnitude in two to three days. Revitalisation occurs in response to rainfall inputs and flow may be sustained at low levels by slow ground ice thaw. In general, however, discharges of subsurface flow after snowmelt are unimportant compared to those developed during the ablation period and this is probably linked to the rapid decline of discharge in nival streams.

One of the aims of the study was to examine the frequency, magnitude, and spatial distribution of rainfall-induced surface flow. Over a three year period, rainfall was able to generate surface flow at the plots on only two occasions. Moreover, both precipitation events were matched in intensity and duration by others which failed to cause surface flow. These observations indicate that, with the exception of high recurrence interval events, rainfall must occur during or immediately following the ablation season in order to cause surface flow. Under these conditions, the areas producing surface runoff are those already at or near saturation as a result of the continuing or recent presence of melting snow up slope. Thus the response of a watershed to a rainstorm of particular intensity and duration may range from a high surface runoff coefficient, to a zero coefficient, depending on the timing of the event relative to the distribution of snowfree areas. The probability of a location being a source area is highest for one downslope of a perennial snowbank, and least on an interfluvium where only a shallow winter snowpack accumulates. Except in rare cases, therefore, the partial area theory for runoff generation is the most relevant conceptual model for a permafrost area. Runoff source areas are variable, but the most likely ones are contiguous with the stream channel, downslope of large melting snowbanks.

### Suggestions for future research

A number of avenues of future research into the action of slopewash processes in permafrost areas appear promising.

First, the hypothesis of increased suspended sediment concentrations in surface wash in polar desert areas, consequent on a decline in vegetation cover downslope of large snowbanks, requires further testing. In addition, the role of the basal ice layer in preventing snowbed erosion should be examined in detail.

Second, this study and others emphasize the role of solution as a major erosional process in periglacial areas. Similar measurements are required from additional locations at a variety of latitudes, and in areas which were glaciated by the "classical" Wisconsin ice sheet where the regolith may be thin.

Third, high recurrence interval snowmelt and rainfall events should be subject to further examination, particularly in regard to their effects on the distribution of source areas of surface flow, and on suspended sediment concentrations developed. The latter could be undertaken through field simulation of snowmelt events and experiments of this type were started in the study area in 1979.

Fourth, there is a need for well-instrumented watersheds in permafrost areas, on a par with those in temperate regions. The link between slopewash and stream runoff requires further examination, as does the movement of sediment on the slopes in relation to concentrations present in streamflow.

All four proposals for future work necessitate detailed and long-term investigations of slope and stream processes in a number of permafrost areas. As demonstrated in the present study, few data currently are available for comparison with any results which might be produced.

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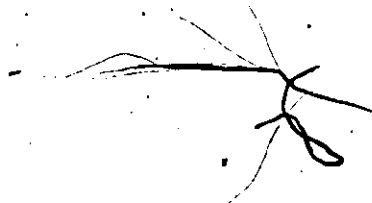
## ABSTRACT

The magnitude and frequency of slopewash processes in the continuous permafrost zone were studied during three summer field seasons (1977-1979). Hydrological and geomorphological aspects of the processes were examined at four small runoff plots located in north-central Banks Island, N.W.T..

Slope hydrological processes exhibited a high degree of variability, both in inter-site and inter-year comparisons. At all sites, however, snowmelt was the most important input and summer precipitation was of very limited significance. During the study, rainfall caused surface flow to develop on two occasions, and then only in areas near or at saturation due to antecedent snowmelt. Consequently, the partial area contribution theory of stream runoff production appears the most valid conceptual model for the study area. Snowmelt-generated surface runoff coefficients at the plots varied on a daily basis in response to energy inputs, and on an annual basis in response to melt season meteorological conditions and winter snow distribution. Only the largest snowbank studied produced a seasonal surface runoff coefficient exceeding 50%.

Suspended sediment concentrations in surface wash were very low, and weight loss due to this process was correspondingly insignificant. The influence of vegetation in protecting the ground surface from the high unit discharges developed at the snowbank sites was of prime importance, and the

maximum suspended sediment concentrations ( $70 \text{ mg L}^{-1}$ ) occurred at the interfluvial site which was largely unvegetated. In contrast, dissolved sediment concentrations were high in both surface and subsurface wash, and surface lowering at the plots by these processes exceeded that by suspended sediment removal by 16 to 250 times. Total rates of denudation by slopewash ranged from  $6.2 \times 10^{-3}$  to  $105.4 \times 10^{-3} \text{ mm a}^{-1}$  averaged over the plot areas, and from  $15.4 \times 10^{-3}$  to  $350.4 \times 10^{-3} \text{ mm a}^{-1}$  averaged over the partial areas of the plots actually affected by wash. In both cases, the maximum lowering rate occurred at the largest snowbank, and the minimum at the interfluvial location. The results suggest that in permafrost areas slopewash can be an important agent of denudation, but in locations where vegetation growth is promoted by high unit discharges, its effects are mainly through solution and not particle sediment removal.



## RESUME

L'importance et la fréquence du ruissellement dans la zone du pergélisol continu ont été étudiées pendant trois étés (1977-1979). Nous avons examiné les aspects hydrologiques et géomorphologiques du processus dans quatre parcelles expérimentales localisées dans le centre-nord de l'île de Banks, T.N.O..

Les processus hydrologiques montrent une grande variabilité aussi bien d'un site à l'autre que d'un été à l'autre. Pour toutes les parcelles toutefois, l'eau de fonte des neiges consistait l'apport le plus important alors que les précipitations d'été étaient peu significatives. Pendant la durée de l'étude, la pluie ne provoqua des déversements qu'à deux occasions, et même alors seulement dans les zones déjà saturées ou proches de la saturation à la suite de la fonte des neiges. Par conséquent, la théorie de l'aire de contribution partielle au ruissellement semble valide dans la région étudiée. Les coefficients d'écoulement résultant de la fonte des neiges dans les parcelles varient sur une base journalière en fonction de l'influx d'énergie, et sur une base annuelle en fonction des conditions météorologiques de la saison de fonte et de la distribution hivernale de la neige. Seul le plus grand banc de neige étudié a un coefficient de ruissellement saisonnier dépassant 50%.

Les concentrations de sédiments en suspension dans l'eau de surface étaient très basses et les pertes de matière résultant de ce processus négligeable. L'influence de la végétation protégeant la surface du sol des débits majeurs apparaissant dans les sites où existent des bancs

de neige était primordiale et les concentrations maximales en sédiments de suspension ( $70 \text{ mg L}^{-1}$ ) ont été mesurées sur l'interfluve qui était largement dénudé. Au contraire, les concentrations en sédiments dissous étaient élevées à la fois dans l'eau de surface et dans l'eau s'écoulant sous la surface, et l'érosion des parcelles par ces processus était 16 à 250 fois plus importante que celle résultant de l'enlèvement de matériel sous forme de suspensions. Pour l'ensemble des parcelles, les vitesses d'érosion par ruissellement variaient entre  $6.2 \times 10^{-3}$  à  $105.4 \times 10^{-3} \text{ mm a}^{-1}$ , mais elles étaient comprises entre  $15.4 \times 10^{-3}$  et  $350.4 \times 10^{-3} \text{ mm a}^{-1}$  si l'on ne considérait que les parties des parcelles effectivement affectées par le ruissellement. Dans les deux cas, les plus grande vitesses étaient observées autour du plus grand banc de neige et les plus basses sur l'interfluve. Ces résultats suggèrent que dans les régions à pergélisol le ruissellement peut être un agent d'érosion important et que là où la croissance de la végétation est favorisée par des débits élevés, son action est surtout le fait de la dissolution plutôt que de la mise en suspension des sédiments.