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SLOPEWASH PROCESSES IN AN ARCTIC TUNDRA
ENVIRONMENT, BANKS ISLAND, N.W.T.

by

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Submitted to the School of Graduate Studies in Partial
Fulfilment of the Master of Arts Degree in Geography.

University of Ottawa
Ottawa, December 1977



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"Writing is good, thinking is better.
Cleverness is good, patience is better."

Hermann Hesse (1922). Siddhartha.

CHAPTER I

AIMS AND LOCATION OF THE RESEARCH

Introduction

The Canadian Arctic is an area currently undergoing considerable governmental, industrial and public scrutiny, regarding resource planning and use. Policy decisions for the area are far-reaching and require a data-base that is as broad as possible. Although a large number of environmental studies have been undertaken in northern Canada and Alaska, recent oil and gas exploration and pipeline construction, have revealed that "there are critical gaps in information available about the northern environment" (Berger, 1977, p. xvii).

This study investigated a geomorphic process, slopewash, which has been largely neglected in permafrost areas. Because of its single-season duration, the research sought only to provide a minimum base from which more substantial and accurate data collection could begin. The study itself was site-specific, but a systematic methodology was utilised in order to allow the possibility of prediction.

Aims

Slopewash ('ruissellement') embodies two sets of processes: (1) surface wash, the downslope transport of weathered material over the ground surface by running water, and

(2) subsurface wash, the group of processes associated with water moving within the regolith (Young, 1972, pp. 62-70). The existence of surface wash (consisting of raindrop impact and surface flow) has been reported from most climatic zones (e.g. Carson and Kirkby, 1972, pp. 188-230; Young, 1972, pp. 62-74), but its effectiveness as a geomorphological agent has been found to vary considerably. It is generally agreed that due to a sparse vegetation cover and high convectional rainfall intensities, surface wash is potentially most important in semi-arid regions. Subsurface wash has been recognised as being significant in the movement of solution loads, and under certain conditions, its transport of particles by piping can be a major process (Kirkby, 1969). The widespread importance of subsurface wash in the removal of sediment particles, however, remains unproven (Young, 1972, p. 71).

Studies of azonal processes such as slopewash have been neglected in permafrost areas (French, 1976b, p. 141), and thus there is a paucity of quantitative material available for comparative purposes. The limited amount of research that has been undertaken, however, (e.g. Czeppe, 1965; De Ploey, 1977; Jahn, 1961; St-Onge, 1969; Tarr, 1897; Wilkinson and Bunting, 1975) indicates that slopewash processes can be important erosional and transportational agents in areas underlain by permafrost. The present research was undertaken with a view to providing more information on the magnitude of these processes, with the hope that this would lead to an increased understanding of their geomorphic role in permafrost environments.

The study of slopewash traditionally has been undertaken from three differing standpoints: that of the hydrologist, who is basically interested in the quantity of water, its spatial and temporal distribution (e.g. Arnett, 1974; Hewlett and Hibbert, 1965); that of the agricultural engineer, who is concerned with the prediction and amelioration of long-term soil losses (e.g. Wischmeier, 1974; Wischmeier and Smith, 1965); and that of the geomorphologist, who is also interested in sediment movement, with the emphasis on long-term slope evolution. Few studies have adopted an holistic approach (notable exceptions: Horton, 1945; Yair and Lavee, 1974), but this was an objective of the present research, since a study of erosion rates made without reference to the distribution of moisture on the slopes, prohibits the analysis of the morphogenetic system and thereby eliminates the possibility of prediction.

The volumes and sediment loads of all the slopewash subsystems, except rainsplash transport, were measured, their spatial and temporal continuity investigated, and their relative importance in geomorphic and hydrologic terms assessed. In order to explain these variations it was also necessary to study the inputs to the slopewash system. Rainsplash was not measured because it was thought to be of minor significance in the study area due to small amounts of summer rainfall and their generally low intensities. The hydrologic data that were obtained, were compared with predictions based on the Horton (1945) model, which is thought (Freeze, 1972) to have some validity in areas possessing frozen ground, and on the theory of partial area contribution which is generally accepted as

being realistic in humid environments (Betson and Marius, 1969; Dunne and Black, 1970; Kirkby, 1969; Ragan, 1968). The geomorphological data were examined in the light of Soviet and Eastern European theories of 'thermo-planation' (Dylik, 1972) and other hypotheses relating to slope segment evolution.

Location

The study was carried out at a locality in north-central Banks Island, N.W.T., Canada between May 27 - July 27, 1977. This island, lying between 71°N and 75°N , is the fourth largest in the Canadian Arctic Archipelago (Figure 1A) and within the zone of continuous permafrost. A camp, known as Thomsen Fly Camp, established by the Geological Survey of Canada (G.S.C.) and situated on the west bank of the Thomsen River (lat. $73^{\circ}14'\text{N}$, long. $119^{\circ}32'\text{W}$) (Figure 1B), was used as a base for the duration of the field season. The three sites examined in the study were located within easy travelling distance of the base camp, 1-1.5 km south of it (Figure 2).

The ubiquitous nature of slopewash processes would have made it possible to conduct the research in many areas of the Canadian Arctic. Ultimately, the choice of study area was determined by the availability of logistical support. Moreover, a study of air photographs of the Thomsen Fly area (taken in 1961 at a scale of approximately 1:60,000) revealed the existence of undulating terrain, thought suitable for the instrumentation (Figure 3).

Thomsen Fly Camp is in the vicinity of the junction between the Cretaceous Christopher Shale and Eureka Sound Formations (Thorsteinsson and Tozer, 1962), but it is probable

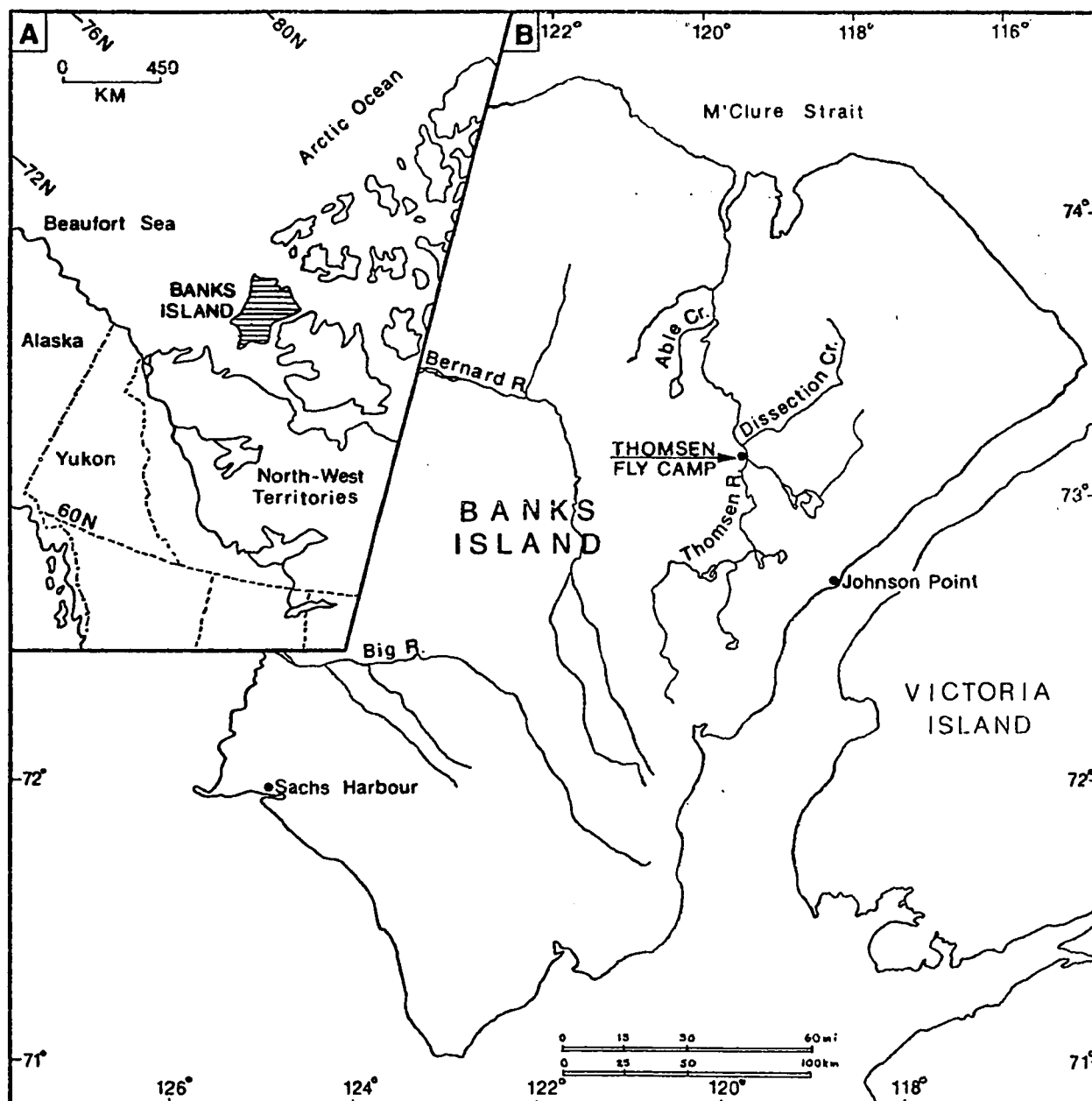


FIGURE 1: - (a) Location map of Banks Island.
(b) Location of study area.

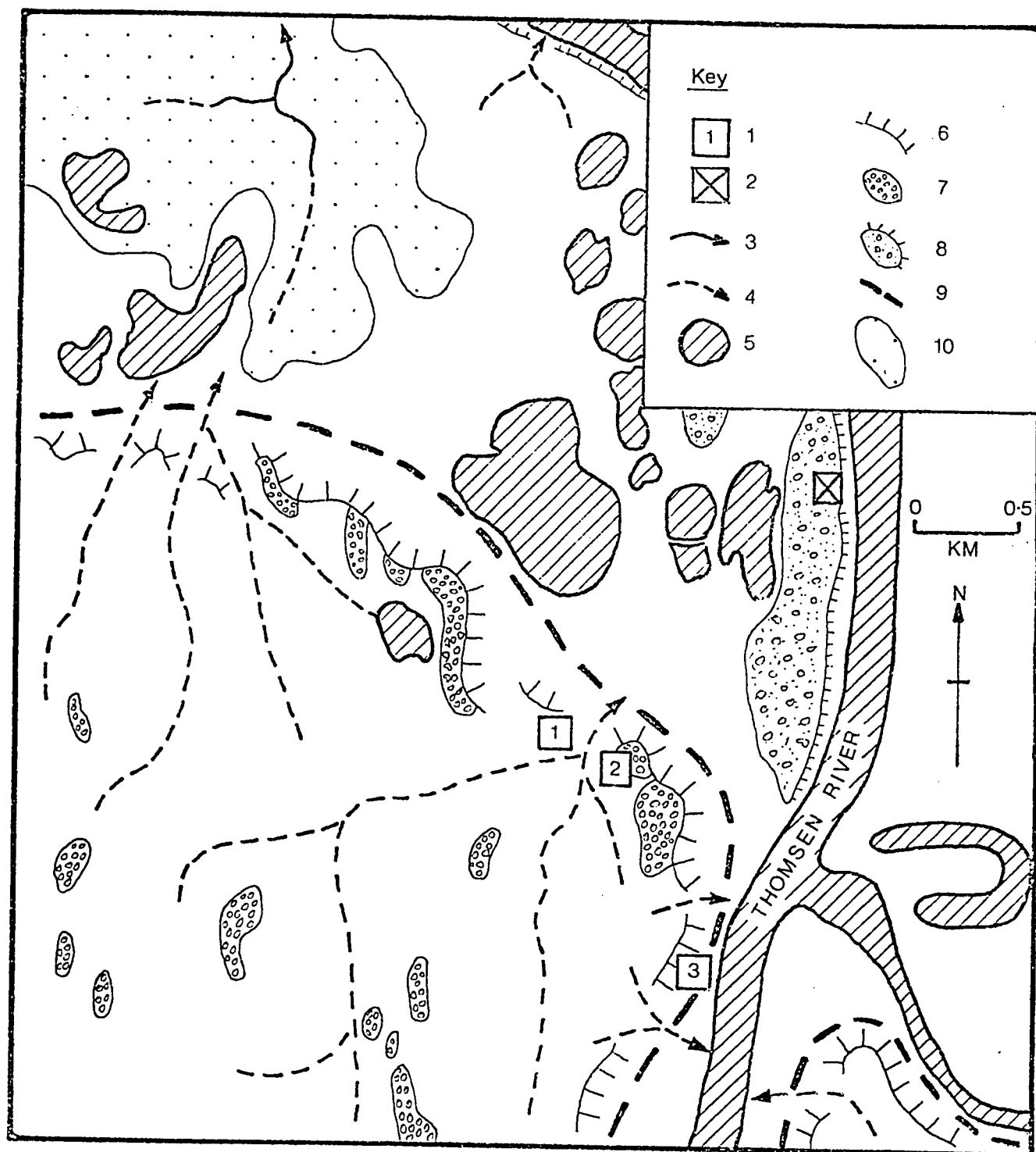


FIGURE 2: - Map of study area.

1- slopewash sites; 2- base camp; 3- streams; 4- ephemeral streams;
 5- water; 6- major breaks of slope; 7- gravel and sand deposits;
 8- Thomsen River terrace remnants; 9- boundary between Thomsen River
 alluvial lowlands and dissected upland terrain; 10- sand plain.



FIGURE 3: - Air photograph of study area (enlargement of part of A-17379-40). Base camp marked by asterisk.

that the three slopewash sites are underlain by Christopher Shale (J.-S. Vincent, personal communication, July 1977).

Bedrock exposures in the immediate area are uncommon, and due to the presence of permafrost, which acts as an impermeable barrier to water movement, slopewash processes are mainly affected by the material above the permafrost table. Surficial deposits, therefore, are of greater significance than bedrock in the study of these processes, assuming that the deposits are thick enough to constitute the whole of the active layer.

The Thomsen River is the largest on Banks Island and rises in the upland terrain to the east of Jesse Harbour. It flows north for over 170 km before entering M'Clure Strait, and unlike the majority of streams on Banks Island which exhibit typical nival regimes, it continues to flow throughout the summer. In the vicinity of Thomsen Fly Camp, the river occupies a broad valley (Figure 4), in which exist a number of terraces, often associated with relic meltwater channels. The region itself was not glaciated in the last (Wisconsin) glaciation, but an earlier ice-sheet probably overode the area and the rest of Banks Island (Craig and Fyles, 1960; Fyles, 1962).

The terrain in the Thomsen Fly Camp area, is regarded as reasonably typical of the middle reaches of the Thomsen River Lowlands. It can be divided into (1) an alluvial lowland (Figure 5), consisting of terrace remnants, tundra ponds and a sand plain (Figure 6), and (2) an upland area (Figure 7) dissected by numerous tributary valleys, possessing gravel caps on many interfluvies and a relative relief of



FIGURE 4: - Thomsen River Valley (upstream, June 15).



FIGURE 5: - Alluvial lowland and base camp (June 9).



FIGURE 6: - The sand plain.
Note: rod is 1 m long.



FIGURE 7: - The dissected upland.

20-50 m. Within the lowland area, the sand plain is a distinctive feature. It is situated 2.5-3.0 km north west of Thomsen Fly Camp as is approximately 4 km² in extent. Similar deposits are found at a number of other localities in the valley of the Thomsen River, but their presence adjacent to the study area probably provides a greater supply of wind-borne particles than would be normal for the region as a whole.

Typically, slopes in the upland area possess short summital convexities grading into long near-rectilinear or concave segments. Except for actively-eroding terrace bluffs and gullies, however, slope angles are low and rarely exceed 15°. In the lowland region, the terrain is generally flat and drainage is poor, giving rise to numerous tundra ponds.

The action of processes other than slopewash is readily observable in this part of Banks Island. For example, ice-wedge polygons, non-sorted stripes and other forms of patterned ground are indicative of mass-wasting and thermal contraction processes. Equally, the Thomsen River and associated gullies, together with blow-outs in the sand plain, indicate the operation of fluvial and aeolian processes. An assessment of the importance of slopewash in relation to these other processes is made in Chapter 4.

Vegetation in the study area is typical of the transition zone between true tundra and polar desert. The interfluves possess a low percentage vegetation cover (often less than 10%). Mountain avens and various saxifrages are the most common plants. In contrast, the wet tundra meadows and snowbank locations exhibit a near-continuous vegetation cover. Grasses,

sedges, mosses and willow are widespread. It would appear that the variety of vegetation and its distribution is related to exposure and moisture (i.e. snow) conditions in the area.

Long-term climatological records for Banks Island are available from Sachs Harbour (located approximately 225 km south-west of Thomsen Fly Camp) where readings have been taken since 1955; additional meteorological data were collected at Johnson Point (65 km south-east of the study area) between 1972-1975. Both data sets show that the island experiences a relatively severe periglacial climate (Table 1): mean daily temperatures are greater than 0°C for only three months of the year when the maximum daily air temperature rises to $+5^{\circ}\text{C}$ or more. In winter, temperatures often fall below -25°C and this extreme cold is combined with aridity: Sachs Harbour receives less than 100 mm precipitation annually (water equivalent). As a result of the size of the island a regional difference occurs in the duration of thaw between the north and the south (Table 1).

The summer of 1977 was abnormally warm on Banks Island. Maximum daily air temperatures of $+21^{\circ}\text{C}$ were recorded twice at Thomsen Fly Camp during the month of July, and the temperature rose to $+20.5^{\circ}\text{C}$ at Sachs Harbour in June, exceeding the 5-yearly maximum of $+18^{\circ}\text{C}$ recorded between 1955 and 1960 (Thompson, 1967). Commencement of thaw, defined as the first date when a sequence of days with mean temperatures above 0°C , is not followed by a sequence of the same or greater length with a mean temperature less than 0°C (Thompson, 1963), was earlier than usual. It occurred on May 30 at both Thomsen Fly

Air temperatures ($^{\circ}\text{C}$)

	J	F	M	A	M	J	J	A	S	O	N	D
Sachs Harbour ^a 1955-1960	-29.4	-31.5	-28.1	-20.0	- 8.8	1.8	5.5	4.5	-1.6	-12.6	-24.3	-28.0
1977 ^b					- 5.6	5.3	8.0	6.0				
Johnson Point ^c 1972-1974	-30.9	-33.5	-33.8	-24.0	-10.0	-0.4	4.9	2.4	-3.1	-14.5	-22.7	-25.7
Thomsen Fly Camp 1977							4.9	12.3 ^d				

Duration of Thaw^e

	<u>Commencement</u>	<u>Finish</u>
N. Banks Island	June 15	September 1
S. Banks Island	June 15	September 12

- a From Thompson (1967)
b Station records
c From Miller (1975)
d Up till July 26
e Thompson (1963)

TABLE 1. - Selected climatological data: Banks Island.

Camp and Sachs Harbour. The greatest differences were observable in the monthly mean temperatures: those recorded from May to August at Sachs Harbour considerably exceeded the averages, and the July mean temperature at Thomsen Fly Camp was even higher.

Previous geomorphological and hydrologic work in the study area was restricted to a reconnaissance in 1975 by a field party of the Geological Survey of Canada (Day and Anderson, 1976; Day and Gale, 1976); other research which had been undertaken within a radius of 60 km had focussed upon surficial geology (Fyles, 1962; Vincent et al., 1975), and permafrost related processes and landforms (e.g. French, 1974a, 1975; French and Egginton, 1973; Pissart and French, 1976). During the 1977 field season, various aspects of the hydrology of the Thomsen River itself were investigated by the G.S.C..

CHAPTER II

METHODOLOGY AND TECHNIQUES

Terminology

The following distinctions in terminology are made:

1) surface runoff refers to above-ground water movement only;

2) overland flow and surface flow are equivalents and include a consideration of sediment as well as water movement, but exclude reference to rainsplash processes;

3) surface wash includes surface flow and rainsplash processes;

4) subsurface runoff refers to downslope water movement in the regolith;

5) throughflow and subsurface wash are equivalents and include reference to subsurface sediment transport as well as water movement;

6) plot is used when referring to a delimited surface runoff-producing area;

7) site is a general term and refers to the plot and surrounding hillslope.

Introduction and accuracy of methods

There are several fundamental problems related to the validity of any short-term process study. First, a major

theoretical difficulty arises from the nature of geomorphic processes whose responses are often non-linear and dependent on threshold values. Thresholds are particularly evident in hydrologic studies, but are problematical due to the large number of linkages within the climatic input system: the isolated consideration of one factor is not usually possible in the natural situation due to the uncontrollable nature of the climatic inputs. Second, the complexity of geomorphic systems normally eliminates the viability of a large-area study unless very considerable amounts of equipment and man-years are available. The results of a small-area study, however, are of little significance unless they can be utilised to make predictions over an area larger than that of the study itself. Third, any prediction of amounts of water and sediment 'per thousand years' derived from the results of a single-season study, involves unwarranted suppositions: even if the climatic inputs are near-average, the possibility remains that an atypical event with a large recurrence interval is more important than all the intervening events, as a result of geomorphic thresholds being exceeded.

The slopes studied can be conceptualized as open systems (Figure 8); this facilitates the examination of the various inputs and outputs of mass and energy which occur, and provides a partial solution to the first two problems outlined above.

A further complication associated with this study was that much of the equipment and many of the methods employed, were experimental. Data obtained towards the end of the study

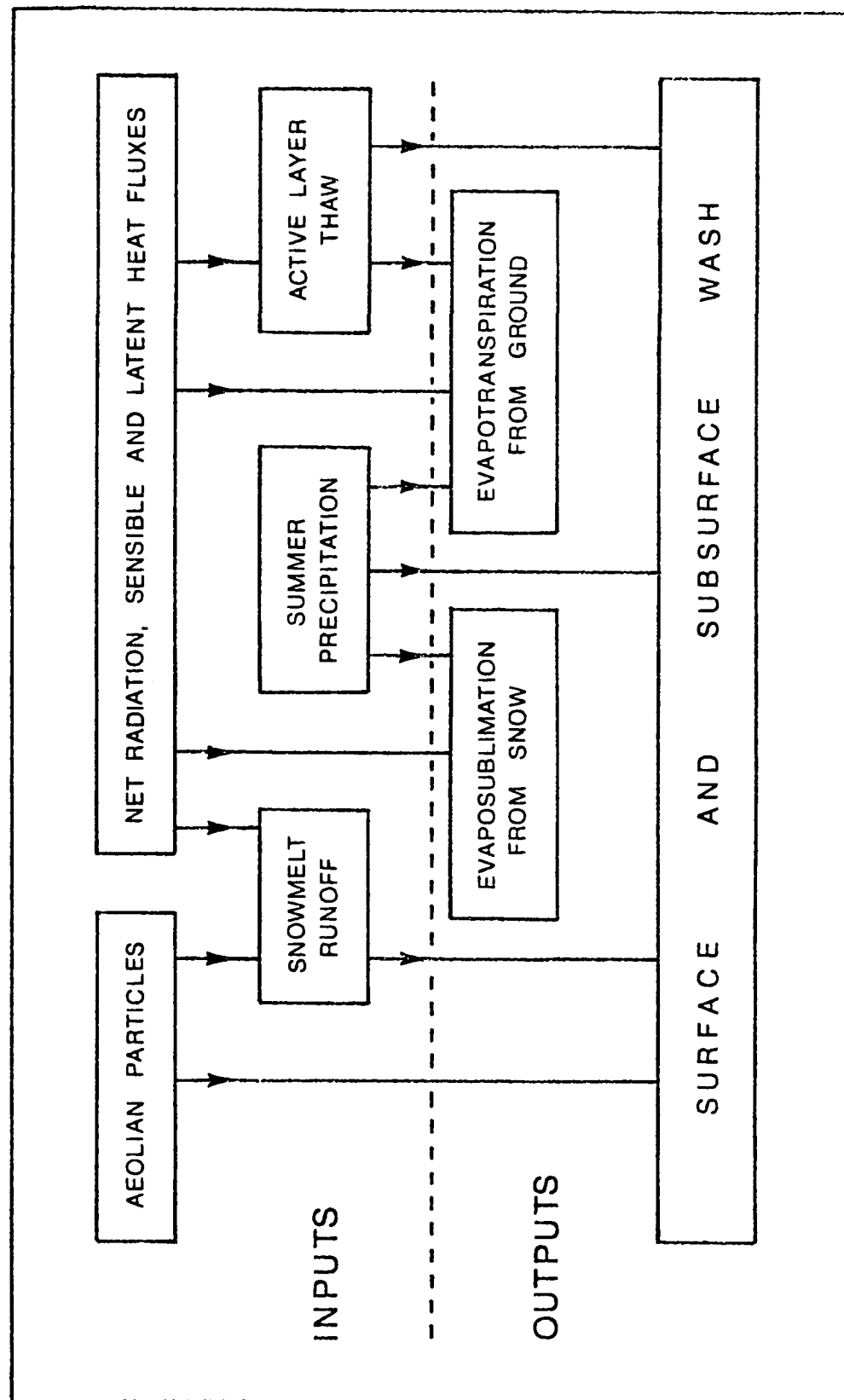


FIGURE 8: - Major components of the slopewash system.

Note: arrows indicate direction of influence.

were probably more precise than those collected at its commencement, but it is certain that all measurements made during the research contained errors, mostly of the precision rather than the accuracy type (Schenck, 1961). Efforts were made to maintain accuracy within $\pm 10\%$ and results are considered to be within these confidence limits unless otherwise stated.

Inaccuracies were attributable to a number of groups of factors: (1) observer error (more probable in cold or wet weather); (2) equipment design (for example, some water leaked from underneath the overland flow collectors); (3) limited time (for example, in the number and frequency of snow depth samplings); (4) expense (for example, only one net radiometer was available); (5) inadequate theory (for example, the assumption that the snowpack contributory area was the same as that of the underlying terrain); (6) disturbance of the system by the presence of the measuring apparatus (for example, the thawing associated with the insertion of the throughflow collectors); (7) transformation of data (for example, in the calculation of the water equivalents of the snowpacks).

It must also be admitted that the estimate of $\pm 10\%$ is uncertain: on some occasions, errors may have been cumulative, whereas at other times, they may have tended to cancel each other out. Nevertheless, the recognition that errors exist and knowledge of their probable magnitudes are essential information in the assessment of the results of the study.

Inputs

(a) Net radiation, sensible and latent heat fluxes

These factors influence runoff and sediment yields directly by providing the energy input for snowmelt, but also indirectly by affecting evapotranspiration and the thawing of the active layer. Other energy inputs can occur from precipitation, or from the ground beneath the snowpack, but the former is normally very slight during the snowmelt period and the latter can be assumed to be zero if the ground remains frozen underneath the snow (Price and Dunne, 1976).

Work in the Canadian Prairies has shown that when the snow cover is complete, 93% of the energy for melt is derived from net radiation and only 7% comes from turbulent heat exchange (Gray and O'Neill, 1974). When the snow becomes patchy, however, the energy contribution for melt by net radiation diminishes to 54-64%, the turbulent transfer of sensible heat increases to 41-49%, and 3-5% of the energy is used for evaposublimation (De Walle and Meiman, 1971; Gray and O'Neill, 1974). Therefore, the timing and magnitude of surface melt at the beginning of summer, can be estimated (assuming the snow is isothermal) from a record of the net radiation balance. A time-lag occurs, however, because nocturnal radiative losses are accumulated into a heat deficit, which has to be satisfied by the first positive flows of the next day before any free water can be produced (Price and Dunne, 1976). After large snow-free areas appear, however,

the prediction of peak diurnal melt is much less accurate, unless variations in the turbulent heat transfer are monitored.

Net radiation was measured using a Swissteco net radiometer and the signals were recorded on a Rustrack recorder. Difficulties were experienced because the Rustrack under-read the voltage produced, when compared with an accurate laboratory millivoltmeter. Fortunately, a correlation between the two produced an r^2 value of +0.998 (significant at the 0.001 level) so that a conversion from the values as recorded on the chart to the true net radiation was simple. The instruments were positioned for approximately three weeks at each of the runoff sites in order to obtain comparable figures. In addition, the air temperature at a height of one metre was measured continuously by a Weathermeasure hygrothermograph inside a Stevenson screen at site 1. Variations in the turbulent heat transfer were not monitored quantitatively.

(b) Water

(i) Snowmelt

The importance of the snowmelt input to the hydrological cycle in permafrost areas is shown by the strongly-peaked hydrographs produced by many streams in early summer. The nival flow regime is well documented in the literature (e.g. Church, 1972; Cook, 1967; McCann and Cogley, 1973; St-Onge, 1965), but in spite of its importance to streams, the geomorphic significance of snowmelt to slope development may be less, since the ground is frozen at the time when the majority of runoff takes place.

The data that were required concerning each of the snowpacks consisted of their initial water equivalents and the changes that took place in these as ablation occurred. The determination of the water equivalent necessitated measurement of the snowpack density, depth and areal extent. Care was taken at all times to avoid walking over the snow on the plot as this would have caused compression and alteration of flow lines.

The density of snow was obtained using a 500 ml snow tube of the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) type, in conjunction with a portable triple-beam balance accurate to 0.1 g. The real accuracy of these measurements at site 1, however, was considerably less than is suggested by this figure. For the first four days, weighing was undertaken in the open and winds caused the balance to move continually up and down. Precision was nearer ± 10 g which gave an approximate final error of $\pm 5\%$. This problem was solved after June 1 when a tent was set up at the runoff site and the remainder of the weighings were made under cover. The measurement of density required the creation of a snow pit and a vertical sampling face (Figure 9). Four pits were dug at each site, and depending on the length of the snowbank, were 4-10 m apart. If the snow appeared homogeneous in a pit, one density sample was considered adequate, but if obvious differences in crystal size and moisture content existed, two or more samples were taken. For the calculation of the water equivalents, the readings were averaged.



FIGURE 9: - Snow sampling pit and equipment.

Note: snow thermometer, snow sampling tube and ice bands in snow.

The depth of snow was determined with a steel rod graduated in centimetres; accuracy was within half a division. The direction of the snow course posed considerable problems. At site 1, the runoff plot was initially delimited by ice 'walls' formed by compression following walking up and down the plot edge. This method proved inadequate in the prevention of overland flow both entering and leaving the plot: it was impossible to avoid spreading a certain amount of mud on the snow surface, and this caused the 'walls' to ablate before the rest of the pack. Had the runoff plot been laid out perfectly, its edges would have cut contour lines orthogonally, and the flow lines would have run parallel to the 'walls'. In this case, it would not have mattered that they disappeared first. However, this method of delimiting the plot was rejected for the other two sites, since it was impossible to know the underlying terrain contour pattern and thus be sure that the plot was correctly defined, until after the snow had ablated. Instead, the plot edges were defined with plastic edging, only as the snow retreated. This method also had its disadvantages: whereas at site 1, the course of snow depth points was just outside and parallel to the plot boundary, at site 2 it was only after the plot boundaries were defined (at the end of the melt) that it was obvious how eccentric the course was in relation to the plot itself. The assumption also was made that the runoff contributing area of the snowpack was the same as that of the underlying terrain. This was not necessarily the case as the presence of ice layers may have diverted flow

within the snow. Nevertheless, the second method was considered the better since the very eccentricity of the snow course at site 2 illustrated how difficult it was to judge the line of maximum slope while a snow cover still existed.

To calculate changes in the snow-pack water equivalent, the reductions in thickness at each of the snow sampling points were averaged and it was assumed that similar melt rates had occurred on the plot. This appeared to be the only feasible way that data from the eccentric snow course could be transformed into 'on-the-plot' data.

The areal extent of the snow cover was plotted by positioning colour-coded dowelling at the edge of the snow. At the end of the season the pieces of dowelling were surveyed and maps of the pattern of snowmelt produced. A planimeter was utilised to calculate the area for each snow survey day.

Other observations of snow included temperature and ice layering. A CRREL snow thermometer showed that the thin snow-pack at plot 1 became isothermal on May 29, and at plot 2, only the deepest part of the pack was less than 0°C on June 6, and by June 7 the whole pack was isothermal. Site 3 was monitored late in the season and by then, the snowbank had long been isothermal. Almost without exception, ice layers were exposed in the snow pits: in one pit at site 2, no less than 8 ice layers were observed, most of them 1 cm or less thick. At another pit, a 5 cm thick ice layer was encountered: this was obviously causing flow diversion because the snow above it was wet, coarse and granular (density 0.46 g cm^{-3}), whilst below the

layer it was fine-grained and powdery (density 0.37 g cm^{-3}). For these reasons, no attempt was made to predict snowmelt runoff by using formulae such as those of Dunne et al. (1976) since they require a large number of input variables and assume that the snowpack is homogeneous.

(ii) Summer precipitation

The amount of summer precipitation is not only very small in a High Arctic environment, but it is also highly variable on an annual basis. Records from Sachs Harbour, Banks Island, include summers with less than 25 mm (1 in.) of rain, and others with more than 75 mm (3 in.) (Miller, 1975). As indicated above, this is one of the major methodological problems associated with the study: a summer storm with a ten-year recurrence interval could have a disproportionate effect on slopewash volume and sediment content, but if it fails to occur within the study's time-period, its effects go unrecorded. Equally, if it does occur and is treated as typical, inflated values are produced. This may not always be the case however, since Pearce (1976b) concluded in a study of overland flow at Sudbury, Ontario, that the greatest amount of erosion is carried out by moderate-intensity-high-frequency events, rather than by the rare high-intensity events. However, the study dealt with Hortonian overland flow and Pearce noted that the same results would not be true in regions where partial area contribution to runoff is normal: in these zones, high intensity events allow a greater area to be subjected to overland flow.

Precipitation was recorded throughout the season at site 1 with a weathermeasure tipping-bucket recording gauge, able to record 0.25 mm (0.01 in.) of rain. June was unusually wet and July unusually dry compared with previous records (Miller, 1975; Thompson, 1967; see Table 2). Another important point was that nearly 98% of the June precipitation fell in 3 storms, the one of June 23 achieving a maximum intensity of 6.5 mm hr^{-1} . This contrasted with the usual type of rain which is light and showery (Miller, 1975). The large number of 'traces' also may have been significant: on Cornwallis Island, a precipitation input measured by a standard gauge as a 'trace', may actually represent 5mm (0.2 in.) of rainfall (Cook, 1960). The same sort of inaccuracies occur in temperate latitudes but they are relatively unimportant there because of large rainfall amounts: in the Arctic where there is very little precipitation and a large number of 'traces', they can be important and may lead to a considerable underestimation of summer precipitation.

During the month of July, five non-recording gauges were set up: one at site 1 for comparison with the tipping-bucket gauge, and two at each of the other two sites. This was an attempt to investigate topographically-induced variations in rainfall amounts (unrelated to gauge efficiency). In a semi-arid region in Israel, an average increase of precipitation of 40-50% was recorded in the valley bottom as compared with the interfluves, in an area where the relative relief was 50 m (Sharon, 1970). The relative relief was similar in the Thomsen Fly Camp region, but the differences in catch were not as

TABLE 2. - Precipitation information: Banks Island.

Month	Day	Time	Amount (mm)	Total Thomsen Fly Camp 1977 (May 27- July 28)	Thomsen River 1975 ^a (June 5- July 31)	Sachs Harbour ^b (av. 1955- 1960)	Sachs Harbour ^c 1977
May	28	20.00-22.00	Tr				
	30	20.00-20.30	Tr				
June	6	16.20-16.45	Tr				
	8	22.00-22.15	Tr				
	13	11.05-11.15	Tr				
		11.45-12.00	Tr				
	14	21.00-22.00	Tr				
	15	09.00-10.00	0.25				
	15	16.00-17.00	2.50				
		17.00-19.00	2.50				
		19.00-21.00	1.25				
		21.00-23.00	Tr (snow)				
	22	18.25-19.45	0.50				
	23	00.00-02.00	0.25				
		02.00-03.00	6.50				
		03.00-04.00	1.50				
	26	14.30-14.45	Tr				
	26	16.45-17.00	Tr				
	27	17.35-18.00	Tr				
	27	20.15-23.00	1.00				
		23.00-01.30	5.25				
	28	01.30-03.30	0.25	21.75 mm (+9 Tr)	3.00 mm (+1 Tr)	4.50 mm	11.20 mm (+5 Tr)
July	2	20.45-21.00	Tr				
		21.25-21.50	Tr				
	6	12.30-12.45	Tr				
	8	01.30-02.30	0.50				
	17	12.30-12.45	Tr				
	19	09.40-10.00	Tr				
	23	15.45-23.00	1.00				
		23.00-03.30	0.50				
	25	01.30-02.30	0.50	2.50 mm (+5 Tr)	15.00 mm (+1 Tr)	25.00 mm	8.80 mm (+3 Tr)

Note: individual events separated by horizontal lines

a From Miller (1975) - b From Thompson (1967) -
c Station records - Tr trace

great (Table 3). The data are too limited to reach a definite conclusion other than to note that differences in catch do exist and that a single rain gauge should not be regarded as representative of a very large area.

(iii) Active layer thaw

Water is present in the regolith in the form of pore or segregated ice; actual amounts are related to the soil moisture content just prior to freeze-back at the end of the previous summer. As the frost-table descends during the season due to positive heat fluxes at the surface, the ice melts and the released water either may move downslope if a saturated zone exists in the thawed layer, or may be held in the pore spaces if there is a moisture deficit.

The varying concentrations of segregated ice make the assessment of the hydrological importance of active layer thaw difficult. Samples of frozen material taken at sites 1 and 3 during the excavation of the throughflow pits did not possess excess ice, although ice veins were observed.

At site 1, reticulate ice veins up to 2.5 cm thick and apparently similar to those described by Mackay (1974), were noted in the top 20 cm of materials, but below this depth they were not readily discernible. At site 3, ice veins were observed in the colluvial materials below the frost-table, but thicknesses were not as great.

Precise evaluation of all aspects of a slope's water balance except for active layer thaw would allow the latter's estimation. It was felt, however, that errors in the

Gauge	Position	Amounts recorded			
		July 8	July 23	July 24	Total
Recording gauge	Site 1	0.5 mm	1.0 mm	0.5 mm	2.0 mm
Non-recording 1	Site 1	0.5 mm	0.88 mm	0.5 mm	1.88 mm
Non-recording 2	Base of site 2	0.5 mm	0.93 mm	0.63 mm	2.06 mm
Non-recording 3	Upper zone of site 2	0.38 mm	1.13 mm	0.63 mm	2.14 mm
Non-recording 4	Upper zone of site 3	0.4 mm	1.0 mm	0.5 mm	1.9 mm
Non-recording 5	Base of site 3	0.5 mm	1.13 mm	0.6 mm	2.23 mm
Conditions		Moderate N.W. Wind	Calm	Calm	

TABLE 3. - Precipitation record variation:
Thomsen Fly Camp (July 8 - July 25).

measurement of other factors were too great to sanction this method of assessment. The small volumes of subsurface runoff collected after the cessation of snowmelt indicated that on the slopes monitored, active layer thaw was a relatively unimportant hydrological input. This was in agreement with the finding of Dingman (1971) and Woo (1976b), both of whom considered the amount of ground-ice derived water in Arctic streams to be minimal.

(c) Aeolian inputs

The inputs of the slopewash system consist of energy and matter: the latter includes particles transported by the wind and deposited on the slopes. The aeolian input must have one or more source areas, and in the study region two obvious possibilities present themselves: (1) the interfluves or high terraces, and (2) the sand plain, 2.5-3.0 km north-west of the base camp. Both of these areas are poorly vegetated and well-drained: they became dry in early summer. Clouds of dust, several tens of metres high, were seen on a number of occasions above the sand plain, and during periods of strong winds, particles in the air were sufficiently concentrated to cause eye irritation at all locations in the study area. Further evidence of active wind action was observed in the dunes in the form of small blow-outs.

Summer aeolian activity was obviously taking place in the Thomsen River region, but it may have occurred also during winter. French (1976b, pp. 204-205) states that the majority of deflation probably occurs during winter months on exposed

snow-free surfaces, and cites the evidence of snow banks containing dirt bands. Of the three runoff sites, no included dirt was visible in the snow pits dug at sites 1 and 2, but substantial amounts of debris were present on the surface of site 2 on June 8. 17 gm of sediment were collected by cleaning 1 m² of the snow surface, melting, and filtering the derived water. It was possible that this had become visible due to concentration following ablation of part of the pack, but the apparent cleanliness of the rest of the snow suggested that winter aeolian deposition was not important at this site.

At site 3, the majority of the snowbank had ablated by the time observations commenced, but apart from a line of sand on the surface of the upper part of the bank (see Chapter 3), the surface of the snow was clean. Considering the thickness of snow that must have ablated (certainly more than one metre), this suggested that it was almost particle free, supporting the hypothesis that the effectiveness of aeolian activity was minimal during the winter. Reasons for this are unclear, but it is likely that the best summer source areas, the blowouts in between vegetated sand ridges 2-3 m high, are covered by snowbanks during the winter and therefore are not erodable at that time.

Outputs

(a) Evaporative losses

(i) Evaposublimation from snow

Evaporation from snow is related to the continuity of the snow cover: losses from a continuous cover are very small

(Male and Gray, 1975), but total consumption of isolated snowbanks by evaporsublimation may reach 20-24% of their initial volume (Gray and O'Neill, 1974; Rechard and Raffelson, 1974). It is possible also, for condensation to take place from the air to the snow surface if there is a positive vapour pressure balance, but amounts are usually small (Martinec, 1976).

A method adapted from that of Gray and O'Neill (1974) to measure evaporation was used in this study. Two two-level, white, plastic containers, 20 cm diameter by 15 cm deep, were filled with snow, weighed and then embedded in the snowpack. They were reweighed the following day in the belief that differences in weights would correspond to net gains or losses as a result of precipitation, condensation and evaporation. The two levels in the containers were connected by three small holes, the aim being to simulate natural conditions and allow melt-water to drain out of the snow, but not to escape.

Field use revealed a number of errors in the method: the white plastic absorbed a considerable amount of radiation and the basal part of the container often was not able to hold all the water produced. The result was a solid mass of ice floating freely in water - clearly very different from the snow surface around it. Another problem was that on days of rapid melt, the outside of the container became exposed and the snow level on the inside also descended: this affected evaporsublimation by reducing wind-speed over the snow surface. These errors alone should have produced consistent results even if the results themselves differed by an unknown factor from the

'true' value. Considerable inconsistencies in values obtained, however, revealed that the major source of error was of the precision type and the prime suspect was the weighing process. Clearly the method was not accurate enough for the large weights of snow and the small differences involved. Only at site 2 were moderately consistent results obtained, and these were used in a water balance study (Chapter 3).

(ii) Evapotranspiration from the ground surface

The measurement of evapotranspiration generally requires extensive instrumentation or a willingness to accept unrealistic assumptions (for example: that the actual rate of evaporation is equal to the potential rate). Stewart and Rouse (1976), however, have adapted a form of the potential combination model (Slatyer and McIlroy, 1961) for use on wet and dry tundra surfaces. When this model was employed by Woo (1976a) on Ellesmere Island, an estimate of 150 mm of evaporation in one summer was obtained, indicating that evaporation can be important in the water balance of a small Arctic drainage basin.

The formula (Woo, 1976a) used was:

$$LE = \alpha \frac{S}{S + \gamma} \cdot R_n \quad (1)$$

where LE is the latent heat flux;
 S is the slope of the saturated vapour pressure-temperature curve;
 γ is the psychrometer constant;
 R_n is net radiation;
 α is a coefficient.

$S/(S + \gamma)$ is a slowly changing function of temperature, and for the range +6.6 — 27.7°C, Wilson and Rouse (1972) used

the formula:

$$\frac{S}{S + \gamma} = 0.434 + 0.012 T_a \quad (2)$$

where T_a is the screen height air temperature.

For a lower temperature range ($0 - 15^{\circ}\text{C}$), a better estimate is obtained with:

$$\frac{S}{S + \gamma} = 0.400 + 0.015 T_a \quad (3)$$

Combining (1) and (3):

$$LE = \alpha(0.400 + 0.015 T_a) R_n \quad (4)$$

Thus evapotranspiration from the tundra should be estimable after measuring net radiation and air temperature, and providing a value of α . Rouse and Stewart (1972) found that for saturated surfaces, $\alpha=1.26$, and that for fairly dry upland tundra sites, $\alpha=1$; these figures were used by Woo(1976a). Recent work, however, has shown that although the α value for saturated sites is accurate, that for unsaturated areas is too high (P. Marsh, McMaster University, personal communication, October 1977). According to Marsh, an average seasonal value of $\alpha=0.6$ for these sites is more reasonable, but in fact α decreases from 1.26 when the ground is saturated, to an unknown but very low value when the soil is dry. In view of these problems, seasonal evapotranspiration totals were not evaluated: however, amounts were calculated for plot 2 but for the snowmelt period only. The area downslope from the snowbank was considered saturated while the area upslope was thought to be unsaturated. Because the maximum amount of time

that the area upslope was without a water input was only 12 days and since no alternative existed, the value of $\alpha=1$ was used for this zone on the plot.

(b) Slopewash

(i) Overland flow

An understanding of the distribution, timing and amount of overland flow was one of the chief aims of the research. In relation to overland flow, the period of snowmelt can be divided into three stages (Zavodchikov, 1965). The stages usually are not equal in length even in the same region, but vary from place to place and from year to year.

During the first stage the ground is snow-covered and melt from the snow surface percolates vertically unless stopped by ice layers. A saturated zone develops at the base of the snow and a downslope movement of water then occurs, but at a velocity up to one hundred times greater than that of the vertical movement (Colbeck, 1974). Active layer thaw at a particular point on the slope does not begin until after the disappearance of the immediate cover (Cogley and McCann, 1975; Dunne et al., 1976; Norum et al., 1973; Pissart, 1967; Washburn, 1973, p. 204). The frozen soil normally has a very low infiltration capacity, in the order of 10^{-5} to 10^{-9} cm sec⁻¹ (Burt and Williams, 1976), so that the majority of the melt-water becomes surface runoff (Dunne et al., 1976; Dylik, 1972; French, 1976b, p. 142; Williams and van Everdingen, 1973).

The effectiveness of overland flow at this stage is a matter of debate: Rudberg (1963) argues that on Axel Heiberg

because the substratum is frozen during the snowmelt period, melt is much less important in terms of erosion than heavy summer precipitation. Church (1974) suggests that the cover of vegetation beneath snowbanks renders meltwater runoff ineffective, and that this is accentuated by frozen ground. Dunne and Black (1971) and Pearce (1976a) provide further support for this view: they note that even thin, discontinuous snow covers are capable of shielding the ground surface, sufficiently that the soil remains frozen. Pissart (1967) on the other hand, postulates that under these conditions, snowmelt still is able to agitate very fine particles at the surface and transport them downslope.

The second stage commences when snow-free areas appear on the slope and active layer thaw begins. The average depth of thaw can be estimated from the formula (Terzaghi, 1952):

$$Z = Bt^{\frac{1}{2}} \quad (5)$$

where Z is the depth of thaw in metres;
 t is the time in days since commencement of thaw;
 B is an empirical coefficient which varies from 0.07 to 0.15 depending on the type of soil and moisture content.

Terzaghi's formula describes the thawing process as an exponential one, giving rise to a straight-line graph of the logarithm of thaw depth against time. The depth of thaw at particular points, however, can be very variable: for example, Dingman (1971) states that the depth to the frost table can vary as much as 9 cm (3.6 in.) within a horizontal distance of 30 cm (12 in.) and Kerfoot (1972) describes differences of the same order - in one instance a 22 cm variation within a

horizontal distance of one metre. Another complicating factor is weather conditions: thaw is accelerated on days with large positive surface heat fluxes and this leads to a scatter of points in the plot of thaw depth against time.

When the thawed layer is shallow (0-20 cm), there is often sufficient moisture available to saturate it and overland flow continues (Tigerman and Rosa, 1949; Woo, 1976b). Concentration of water into rills can occur on steeper slopes and this is usually the stage of peak runoff and erosion. The ability of snowmelt-derived overland flow to transport fines is reported from Ellef Ringnes where it has been observed on slopes developed in gabbro, sandstone and shale (St-Onge, 1965; 1969). Values of sediment transport of up to $18 \text{ gm m}^{-2} \text{ yr}^{-1}$ have been reported from Spitsbergen in flows derived from melting snow (Jahn, 1961), although it is possible that some of this sediment was aeolian in origin. A figure of 0.03 metric tons yr^{-1} for the sediment transport of 68 rills over a 450 m wide slope on Devon Island (Wilkinson and Bunting, 1975) illustrates the importance of surface flow in a permafrost environment.

The final stage of snowmelt begins when all snow, except that forming the snowbank, has ablated, and ends when the snowbank itself disappears. Extensive overland flow no longer occurs but a precipitation input can cause surface runoff, particularly in the area immediately below the ablating snowbank. This is a special case in the theory of partial area contribution to runoff, since the areas of saturation and hence

the contributory areas, are related to the persistence of snowbanks. Other slopes which are above a snowbank or have become devoid of snow, are usually too dry to generate runoff due to continual evaporation and throughflow: the small precipitation amounts normally are not sufficient to restore soil moisture saturation unless they occur very soon after the disappearance of the snow.

The surface runoff plots in this research were defined with plastic edging (Figure 10): this was inserted into the ground to a depth of approximately 5 cm. At site 1 the whole plot was defined at one time, but at sites 2 and 3, edging was inserted upslope as the snow front retreated and the ground thawed (Figure 11). Overland flow was collected at the outlet of the plots in copper boxes (Figure 12) attached by pipes to calibrated storage containers (Figure 13) capable of holding 60 litres of water (Figure 14). The containers had been calibrated in the laboratory prior to departure for the field: volume was related to depth so that field measurement consisted only of using a metre rule to take the depth of water.

Several difficulties were encountered in the method of collection. The first problem of ensuring that no flow was lost beneath the collectors, was overcome by using a silicon compound to make a seal between the metal and the ground surface. This worked successfully at all three sites, but at site 2 not all the water which flowed over the plot passed into the collector. At this site, it was found that the collector inadvertently had been placed upon a band of moss,

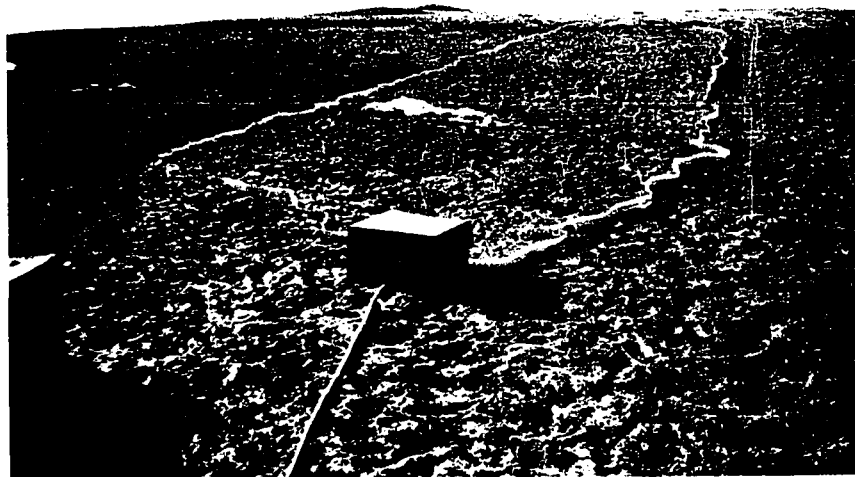


FIGURE 10: - Plastic edging, plot 1 (June 3).



FIGURE 11: - Plot 2.

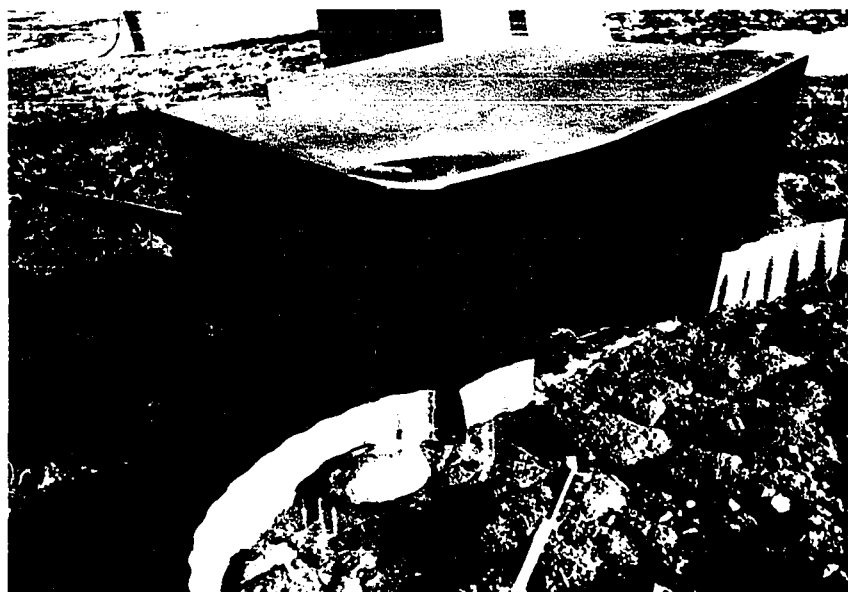


FIGURE 12: - Overland flow collector.
Note: sediment trapped in collector.

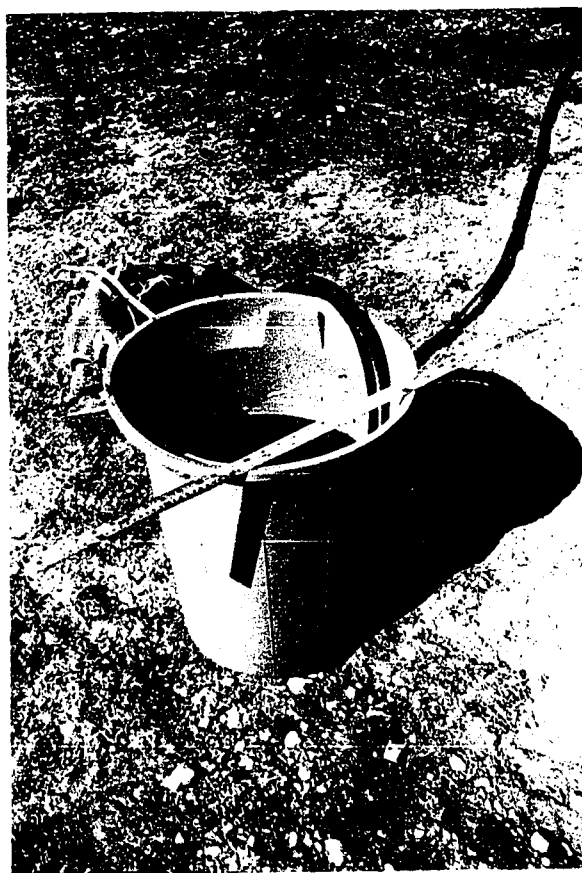


FIGURE 13: - Overland flow storage container.

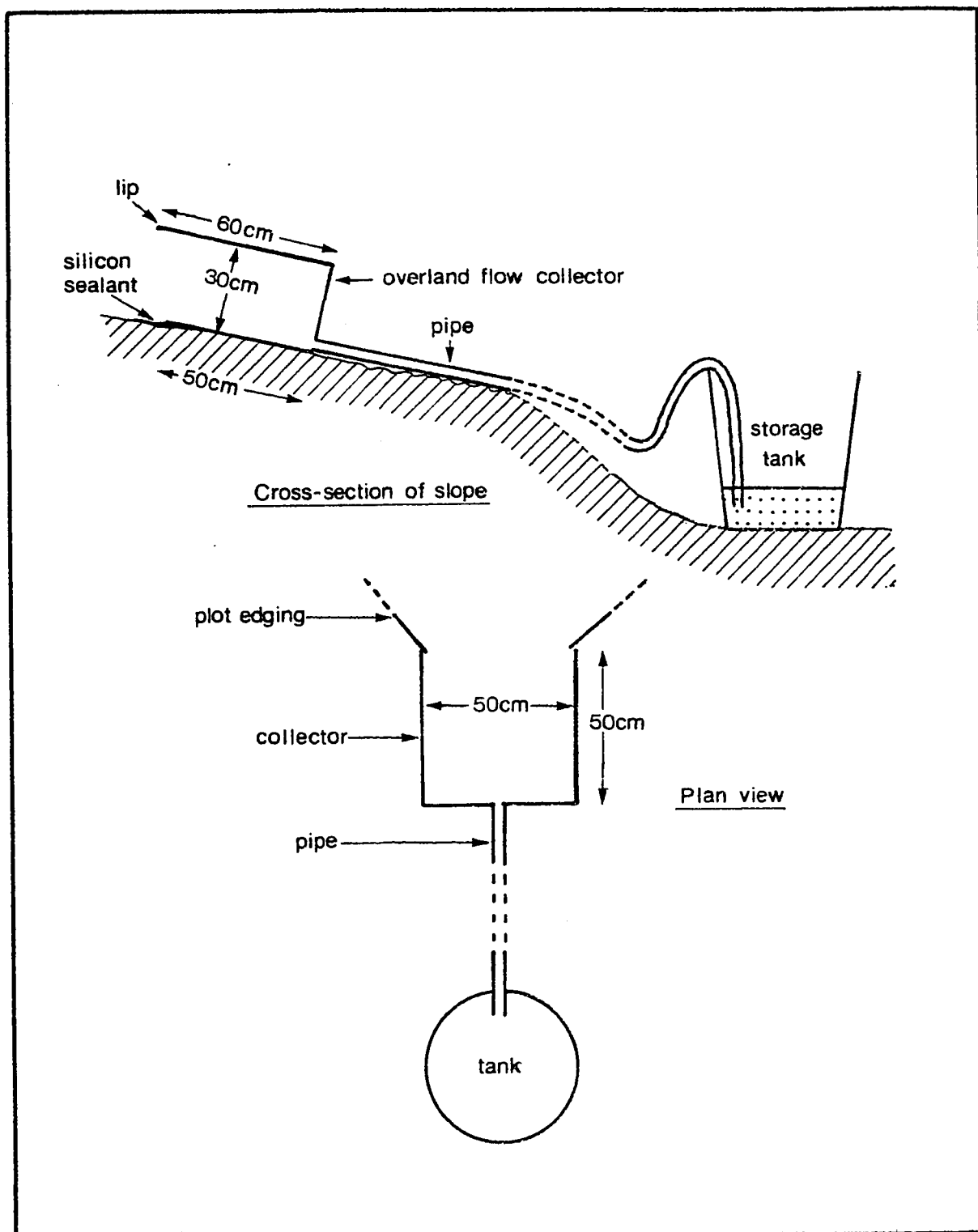


FIGURE 14: Overland flow collector and storage container.

and a considerable quantity of water must have passed under its base plate and through the moss itself. This may not have been an abnormal phenomenon as vegetation lines of this type are quite common in the study area. The flow posed a problem of definition, however, since it could not be regarded as surface runoff, nor was it taking place within the regolith.

It is thought that these lines of vegetation are the High Arctic equivalent of the seepage lines described by Dingman (1971) from Alaska, and Zoltai and Pettapiece (1973) from the Mackenzie River Valley and northern Yukon. Such lines occur widely on Banks Island (e.g. French, 1976b, Figure 9.5; personal communication, 1977) and have been observed on Melville, Cameron and other islands in the Canadian Arctic. Dingman considered that this type of water movement may be an important contributor to stream recession flow. Seepage lines may also be significant in terms of slope development since vegetation probably acts as a trap for sediment moving down-slope.

The process of water movement through seepage lines does not fit clearly into either of the categories of slope-wash as defined by Young (1972); it is suggested, therefore, that a third type of slopewash be recognised, known as 'intra-vegetational wash'. Observations indicate that in terms of velocity, it is closer to surface than subsurface runoff, but volumes and sediment loads remain uncertain.

A second problem exists in relation to the design of the surface wash collectors: they are only suitable for areas where the

amplitude of micro-topographic relief is small. Therefore, they could not be sited in hummocky areas. In part, this determined the final choice of sites for the runoff plots.

At site 2 an unforeseen difficulty was encountered. The very low slope angle, between 3° — 4° , necessitated a long pipe leading from the collector to the storage bin in order to attain the hydrostatic head required to allow flow. Despite an internal pipe diameter of 2.5 cm, frictional resistance in the pipe hindered flow such that water began to pond up in front of the collector. The solution was a shorter pipe and the positioning of storage bins on their sides. This reduced their capacity by greater than 50% and highlighted the labour-intensive nature of the instrumentation, since the storage bins needed to be emptied more frequently. For example, at peak melt at plot 2, a bin had to be measured and emptied on average every 1.5 minutes. This obviously prevented the concurrent measurement of runoff at a number of plots. Furthermore, it resulted in gaps in the data record since continuous 24 hour monitoring was impossible.

Suspended sediment in the overland flow was sampled by filling litre polythene bottles with water from the storage bins. The samples were filtered in the field using an Østrem filter pump (Figure 15) and the filter-papers were brought back in petri dishes for accurate weighing in the laboratory. The filter papers had been weighed individually without drying before departure for the field. Each filter paper was dried for 12 hours at a temperature of 105°C before reweighing. In

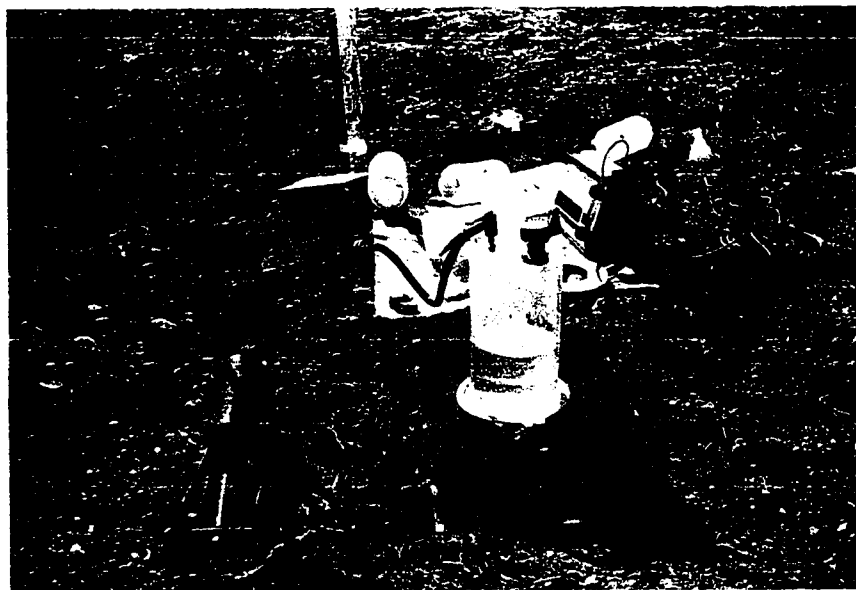


FIGURE 15:- Østrem filter pump.



FIGURE 16:- Site 1: throughflow pit.

Note: terrain disturbance around
pit.

addition, the average difference in weight between a dried filter paper and one in normal atmospheric conditions was established and used as a correction factor. Inaccuracies in the method were related to the estimation of this factor and to the filtering process: ideally, the filter pump and bottles would have been flushed out with distilled water after each filtration but this was not possible in the field.

Solute concentrations in the overland flow were measured with a Chemtrix portable conductivity meter, samples being taken at the same time as those of suspended sediment. The measurements were made in micromhos cm^{-1} , which can be transformed into equivalent values of milligrams litre $^{-1}$ of sodium chloride or calcium carbonate.

(ii) Subsurface wash

Throughflow measurements at a subarctic location (Schefferville, Québec), have shown that during snowmelt, no subsurface runoff occurs at a depth of one metre (Dunne et al., 1976). The fluctuations of active layer water levels have been recorded on Ellesmere Island (Woo, 1976b), and velocities, but not amounts of interflow, have been measured on Devon Island (Wilkinson and Bunting, 1975). The examination of the recession curves of stream hydrographs in Alaska, led Dingman (1971) to conclude that subsurface routing of water is very important in permafrost areas. Its absolute importance at a particular locality, however, may depend on a wide range of interlinked factors including surficial materials, rate of thaw, vegetation type and slope configuration.

Throughflow was collected by means of copper guttering welded to brass frames and inserted into the ground across the slope. Downslope flow was diverted laterally into plastic pipes leading to storage containers, by means of a polythene sheet in contact with the back of the guttering (Figure 16).

At site 1, a frame with four gutters attached at average levels of 6 cm, 16 cm, 31 cm and 46 cm (Figure 17) was positioned before thaw occurred. A site was chosen outside the boundaries of the plot in order to avoid disturbance of the plot itself, and a pit approximately 60 cm deep, together with a trench 3 m long, were dug in the seasonally frozen ground with a pick-axe. The frame with pipes and polythene attached was lowered into the pit and each gutter was separately backfilled. A second pit was dug 3 m downslope from the throughflow collectors: its purpose was to hold the storage containers at levels whereby water could flow through the pipes connecting the guttering and the containers.

A number of technical problems were encountered in this installation. First, the aim was to leave the area in a condition closely resembling its initial state so that the positioning of the apparatus would have only a minor effect on the processes taking place. Unfortunately the delving of the throughflow pit took several hours and the movement around it necessary for the digging left the surface considerably disturbed. The upslope wall of the pit became wet as a result of water issuing from the upper few centimetres of regolith, and the pit itself had to be bailed out several times. This

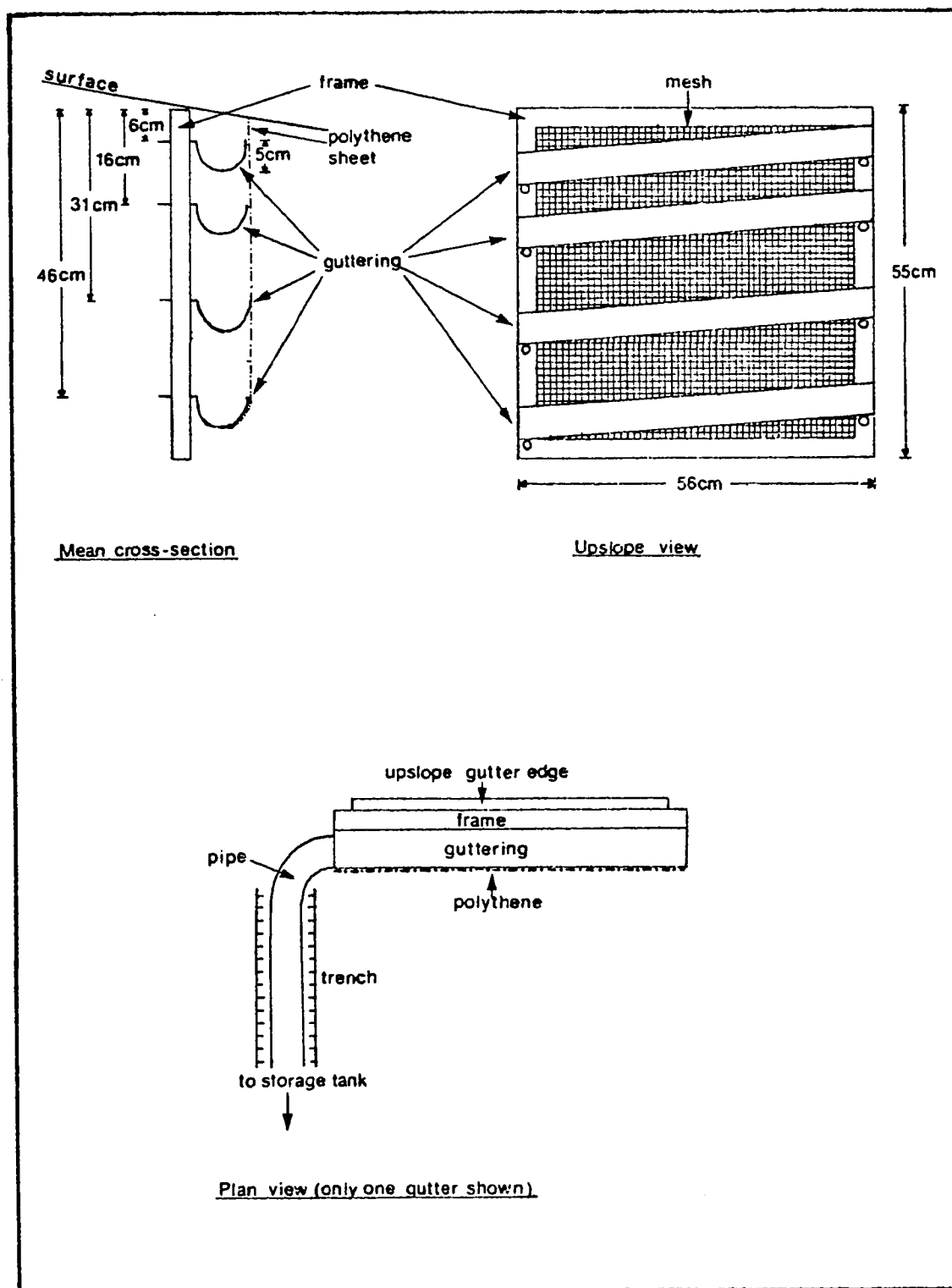


FIGURE 17: - Throughflow collector.

called into question the premise that natural conditions were recorded.

Second, in order to reduce disturbance, the trench between the two pits was filled in as quickly as possible. This turned out to be an error since excavation at the end of the season revealed that the bottom pipe was blocked by a twist in its length and that from the second gutter was probably laid at too low an angle which may have reduced flow. Thus at site 1, only flow collected from the top and third gutters can be regarded as reasonably representative.

A third problem at site 1 was that the pipes were disconnected for two hours on June 2, in order to allow shoring up of the open storage pit, whose sides were collapsing: unfortunately, this proved to be the time of peak flow. It was assumed then, that throughflow would continue for a considerable number of days, so that this break in the record would have been of little importance. In fact, throughflow decreased very rapidly that day and ceased altogether two days later.

A further technical problem of the throughflow collection apparatus related to the existence of frozen ground. Although the gutters collected water at average depths of 6 cm, 16 cm, 31 cm and 46 cm, the depths of their exit pipes were 14 cm, 24 cm, 39 cm and 54 cm respectively. Thus it was possible for subsurface runoff to occur but not be recorded when the gutters were thawed, but the pipes were not.

Throughflow was also recorded at site 3. Here, a single gutter was positioned at an average a depth of 6 cm, close to the overland flow collector, whilst 3 others were inserted at mean depths of 6 cm, 16 cm and 31 cm, approximately 150 m downslope. These collectors were enplaced in a similar fashion to the apparatus at site 1, but the previous experience enabled an increase in data accuracy. Terrain disturbance was greatly reduced by standing on boards while digging and by shortening the time spent on this task. The removal of the surface vegetation and soil in one mass, and its placement in the top gutter meant that permeabilities at this level remained as close as possible to those originally existing. The pipes were laid down with greater care and end-season excavation confirmed that they were in positions to function correctly.

The main methodological problem of subsurface wash measurement relates to the estimation of the contributory area. The assumption cannot be made that this area is merely an extension of the width of the collector upslope, orthogonal to the surface contours. Distortions of flow in the region of the pit due to the alteration of regolith structure during excavation should be expected (Carson and Kirkby, 1972, p. 196; Knapp, 1973; 1974) and these depend also on the degree of saturation of the soil, and the amount of thawing that results from digging the pit.

One advantage of permafrost regions is that the contributory area may be estimated from an examination of the subsurface contours of the frost-table. Accordingly, the

depths to the frost-table were measured at grid points laid out upslope of the throughflow collectors. Each point was marked by dowelling, and the thickness of the thawed layer taken at various intervals during the summer. A T-handled probe was used to measure depth: it was pushed into the unfrozen layer until halted by the frost-table, and the depth was read from centimetre graduations along the probe's length. It was possible to distinguish between the frost-table and stones, since the 'feel' and sounds produced when hitting the two were quite different. Frost-table depths were considered accurate to ± 5 mm.

When all the snow had ablated, the grid points were surveyed using a Zeiss level and staff, and a series of maps showing the contours of the subsurface frost-table were produced. This enabled the calculation of the average width of slope drained by the throughflow collectors and thus an estimate of throughflow per unit width of slope. In this context the usual problems relating to the interpolation of isolines from point data were encountered, notably when one point on the frost-table was at a higher elevation than those surrounding it: it was not possible to know whether this was an outlier, or whether it was connected to an area of the same elevation further upslope. This was not particularly important when the problem occurred at a distance from the collector, but when it was within one metre, it caused considerable uncertainty in the projected contributory area. The analysis also assumed that saturation elevations within the thawed layer correlated

directly with frost-table contours, although this has yet to be proved.

An examination of the lateral eluviation of fines in throughflow was attempted, although Young (1972, p. 71) and French (1976b, p. 142) have questioned the magnitude of this process in general, and in permafrost environments respectively. Eastern European and Soviet researchers, however, regard the removal of particles in throughflow as particularly important to the evolution of periglacial slope forms. They argue that 'thermo-erosional wash' (Dylik, 1972) resulting from the melting of pore ice in the active layer is able to carry fine sediment particles downslope. On a large scale, thermo-erosional wash is thought capable of relief reduction or 'thermo-planation'. The throughflow collected in this study, was filtered to examine particulate sediment, and its dissolved solids content was measured with the conductivity meter.

CHAPTER III

RESULTS AND DISCUSSION

A variety of field data was collected during the two months spent at Thomsen Fly Camp. The following discussion restricts itself to the examination of the three principal processes studied - surface runoff, subsurface runoff and sediment transport. Comparisons are made amongst and between the processes at the three different sites and at various times through the season. Intensive statistical analysis is avoided as the type and limited accuracy of some of the measurements does not warrant such an approach. All times are given in local time, a close approximation of solar time on Banks Island.

The runoff plots

Three sites were chosen in order to assess spatial variability in slopewash processes. A runoff plot was set up at each site, but the labour intensive nature of the research meant that overland flow was studied consecutively, rather than concurrently, at the three plots (Table 4). Inter-site variation of factors affecting overland flow and throughflow was great (Tables 5 and 6), but of particular importance was the presence or absence of breaks of slope within the plots. Rapid changes in slope angle provided the ideal topographic

	Phenomena measured	Timing of study
Site 1	(i) Overland flow	May 28 - June 5
	(ii) Throughflow: beside surface wash site at depths of 6 cm, 16 cm, 31 cm, 46 cm.	June 1 - July 17
Site 2	(i) Overland flow	June 5 - June 19
Site 3	(i) Overland flow	June 20 - July 27
	(ii) Throughflow: beside surface wash site at a depth of 6 cm.	June 22 - July 20
	(iii) Throughflow: 150 m down-slope from surface wash site at depths of 6 cm, 16 cm, 31 cm.	June 21 - July 27

TABLE 4. - Types and timing of slopewash studies.

	Length (m)	Average width (Area/length) (m)	Area (m ²)	Average Slope	Aspect	Position of site on slope
Site 1	28.7	5.2	149.6	5°12'	SE 135°	Interfluve
Site 2	54.1	7.2	387.5	5°19'	NE 315°	Valleyside slope
Site 3	18.8	1.4	26.6	14°18'	SE 125°	Valleyside slope

TABLE 5. - Dimensions and positions of surface runoff plots.

	Micro-relief	Surficial Materials	Soil series ^a	Vegetation cover	Vegetation type
Site 1	<u>Dryas</u> hummocks, 5-10 cm in height, 10-20 cm in diameter. Frost fissures. Lag gravels.	Sandy gravels	Polar Desert	10 - 20%	<u>Dryas integrifolia</u> <u>Saxifrage</u> sp. <u>Popaver</u> sp.
Site 2	Intra-vegetational washlines lower half of plot. 8-20 m from top of plot: large <u>Cassiope</u> earth hummocks. Top 8 m of plot: small <u>Dryas</u> hummocks, less than 10 cm in height.	Silty colluvium	Bernard Series	60 - 80%	<u>Dryas integrifolia</u> <u>Cassiope</u> sp. Mosses and lichens
Site 3	Lower 9 m of plot: almost level. Upper 9 m: small hummocks, 10-15 cm in height, less than 25 cm in diameter.	Sandy colluvium	Upper 9 m: Hummocky. Lower 9 m: Polar Desert Tundra Interjacence.	Upper 9 m: 20% Lower 9 m: 40%	<u>Dryas integrifolia</u> <u>Saxifrage</u> sp. <u>Graminae</u> sp.
Lower Throughflow Site	Irregular surface: relief amplitude maximum 20 cm.	Silty colluvium	Meadow Tundra	100%	<u>Salix</u> sp. <u>Graminae</u> sp. <u>Juncus</u> sp.

^a After Tedrow and Douglas (1964); Tedrow (1974).

TABLE 6. - Surface characteristics of slopewash sites.

conditions for snowbank formation, and while these conditions existed at sites 2 and 3, they did not at site 1 (Figure 18).

A topographic map of each surface runoff plot was produced following detailed surveying at the end of the summer (Figures 19, 20 and 21). Although the contour interval on the maps is 10 cm, the contours themselves are regarded as generalised, since micro-topography on all plots gives rise to a relief amplitude greater than this figure. Only on the lower part of plot 2 are topographic differences due to intra-vegetational wash lines identifiable; individual hummocks are not distinguished, however, since it was not feasible to survey them singly.

The different methods used to delimit the runoff plots (Chapter 2) resulted both in dissimilar sizes (Table 5) and outlines: the shapes of plots 2 and 3 represented the areas draining naturally to their respective overland flow collectors, whereas the relatively regular outline of plot 1 was the result of the imposition of unnatural boundaries on the surface runoff.

Surface runoff

The primary source of water for surface runoff in the study area was snowmelt. The relationship was not a linear one, however, as can be illustrated by a description of the stages leading up to and including the observations of surface runoff at plot 1.

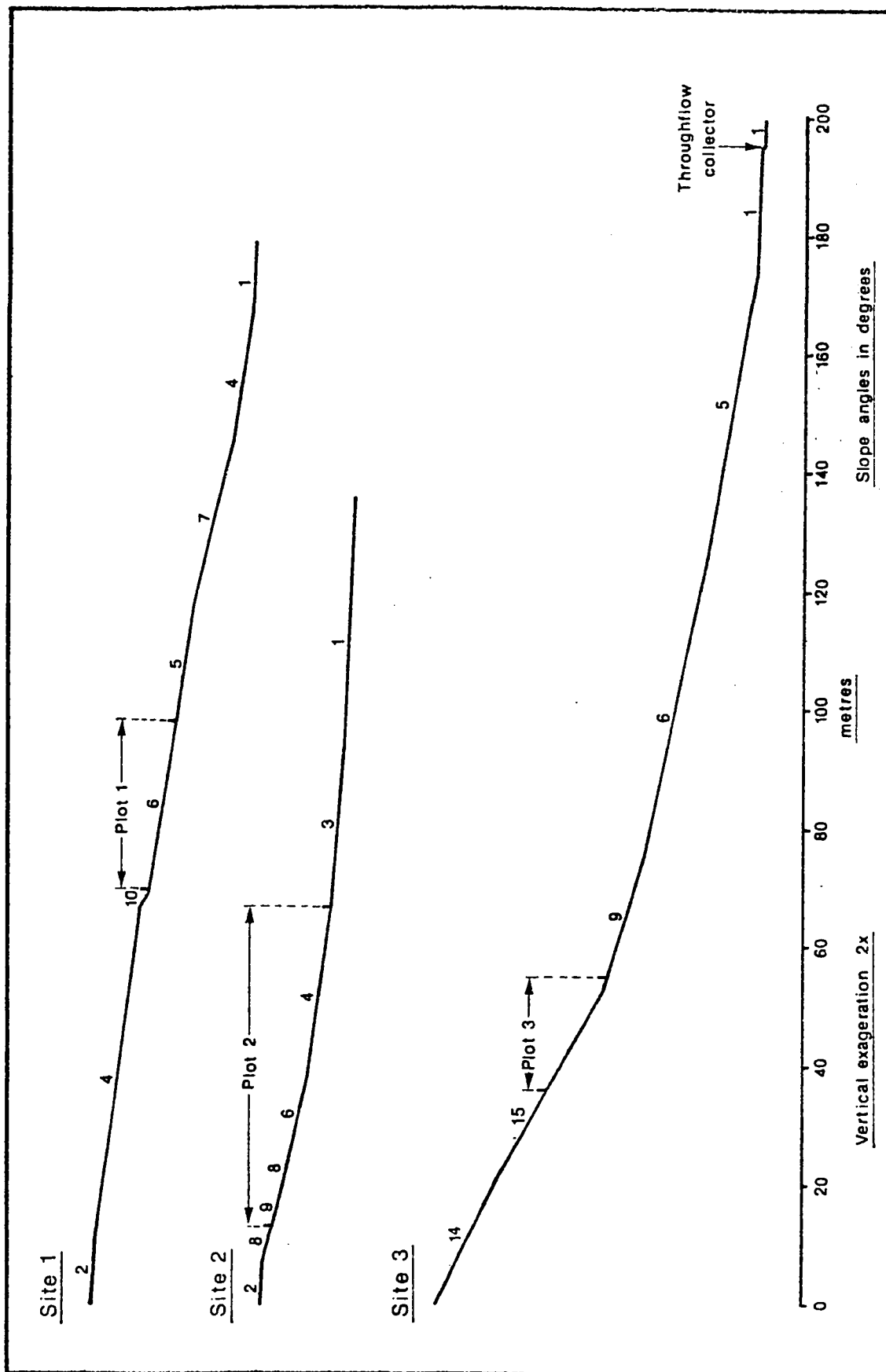


FIGURE 18: Plots 1, 2 and 3: slope profiles.

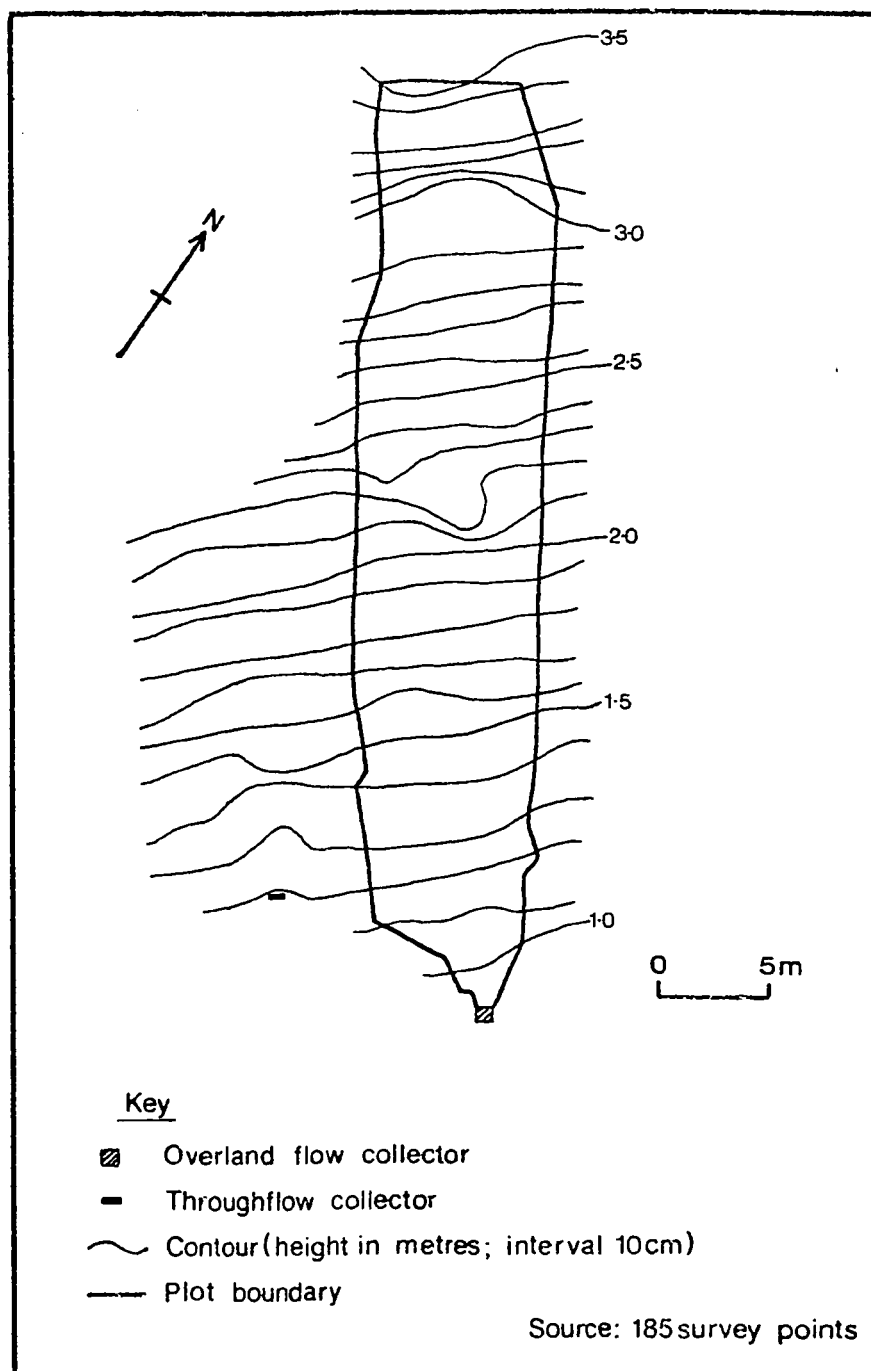


FIGURE 19:- Plot 1: topographic map.

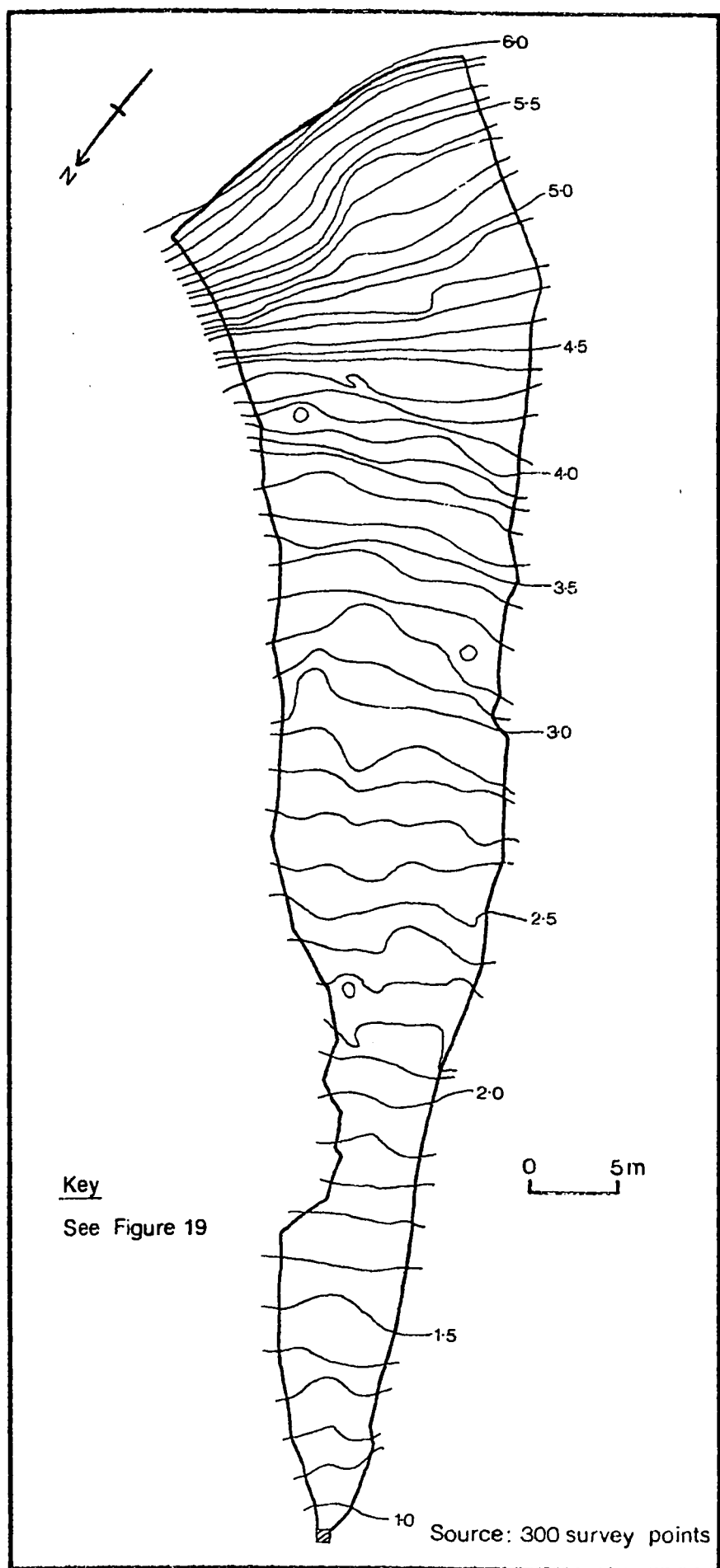


FIGURE 20: - Plot 2: topographic map.

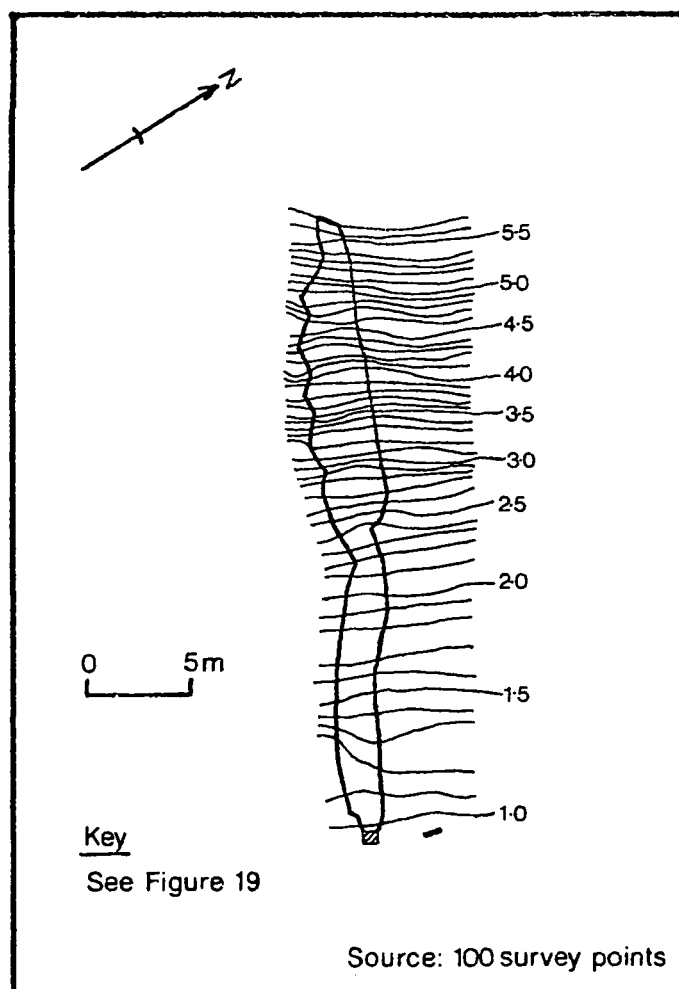


FIGURE 21: - Plot 3: topographic map.

(a) Plot 1

Melt had not begun on arrival at Thomsen Fly Camp on May 25. The next day, however, the air temperatures rose above 0°C and combined with strong winds ($40\text{--}50\text{ km hr}^{-1}$), resulted in considerable quantities of snow ablating in situ. The thin covers on terraces in the area were particularly affected. No runoff was observed and it was concluded that the majority of the snow sublimated, while any liquid water infiltrated into the very thin thawed layer.

The installation of the overland flow collector at plot 1, was completed on May 28, by which time snow cover was absent along the ridges in the area. The snow adjacent to the plot was not yet isothermal since temperatures within it at a depth of 33 cm (the deepest point) were -3°C . Snow densities averaged 0.35 gm cm^{-3} , a typical figure for tundra snow (Slaughter and Crook, 1974). By May 29 the density had increased to 0.41 gm cm^{-3} and the snowpack had become isothermal. On this date standing water began to develop on the plot, but the surface subsequently became drier in spite of a considerable decrease in the water equivalent of the snowpack. Ablation continued on May 30, but it was not until the afternoon of May 31 that surface runoff was observed. Up to this time, 70% (water equivalent) of the snow on the plot had ablated without runoff over the surface occurring (Table 7). The surface runoff of May 31 was not recorded due to leaks in the plot edging, but under the influence of clear skies, further flows occurred on June 1 and 2 and these were collected.

Time	Loss of water equivalent of snowpack m ³	Amount of surface runoff m ³	Surface runoff as a percentage of change in water equivalent
15.00 May 28 - 21.00 May 29	2.01	-	0
21.00 May 29 - 21.00 May 31	5.23	0.025 ^a	0.5
21.00 May 31 - 21.00 June 1	1.94	0.670	34.5
21.00 June 1 - 10.00 June 3	0.93	0.279	29.9
10.00 June 3 - 10.00 June 4	0.22	-	0
Original water equivalent of pack: 10.33 m ³			
Total surface runoff: 0.974 m ³			
Percentage of water loss as surface runoff: 9.4%			
a Estimate			

Table 7. - Plot 1: surface runoff production.

By the end of June 2, only 10% (water equivalent) of the original pack was left and this was not enough to produce standing water or flow on June 3, despite continuing sunshine.

The non-linear response of surface runoff to snowmelt can be partially explained by a consideration of other factors in the slopewash system. The amount of snow which ablated without causing surface runoff may reflect weather conditions: until May 31, the sky had been obscured by clouds so that amounts of incoming radiation were low, and melt occurred as a result of turbulent heat transfers. This melt was mainly in the form of sublimation, and the free water that was produced was not present in quantities large enough to cause saturation and runoff. The three consecutive days of sunshine and the accompanying values of radiation, together with higher air temperatures (Figure 22), caused melting sufficient to exceed the evaporative and subsurface losses.

If the above hypothesis is correct, and if the weather had remained overcast but with air temperatures above 0°C , the snow might have continued to ablate without giving rise to surface runoff. Equally, had the sky cleared two or three days earlier, the amount of flow produced might have been considerably greater. Spatially this suggests that different sections of a slope may be affected from year to year, depending on the prevailing conditions of cloud cover and insolation. The occurrence of surface runoff at site 1, in fact, may have been a rare event.

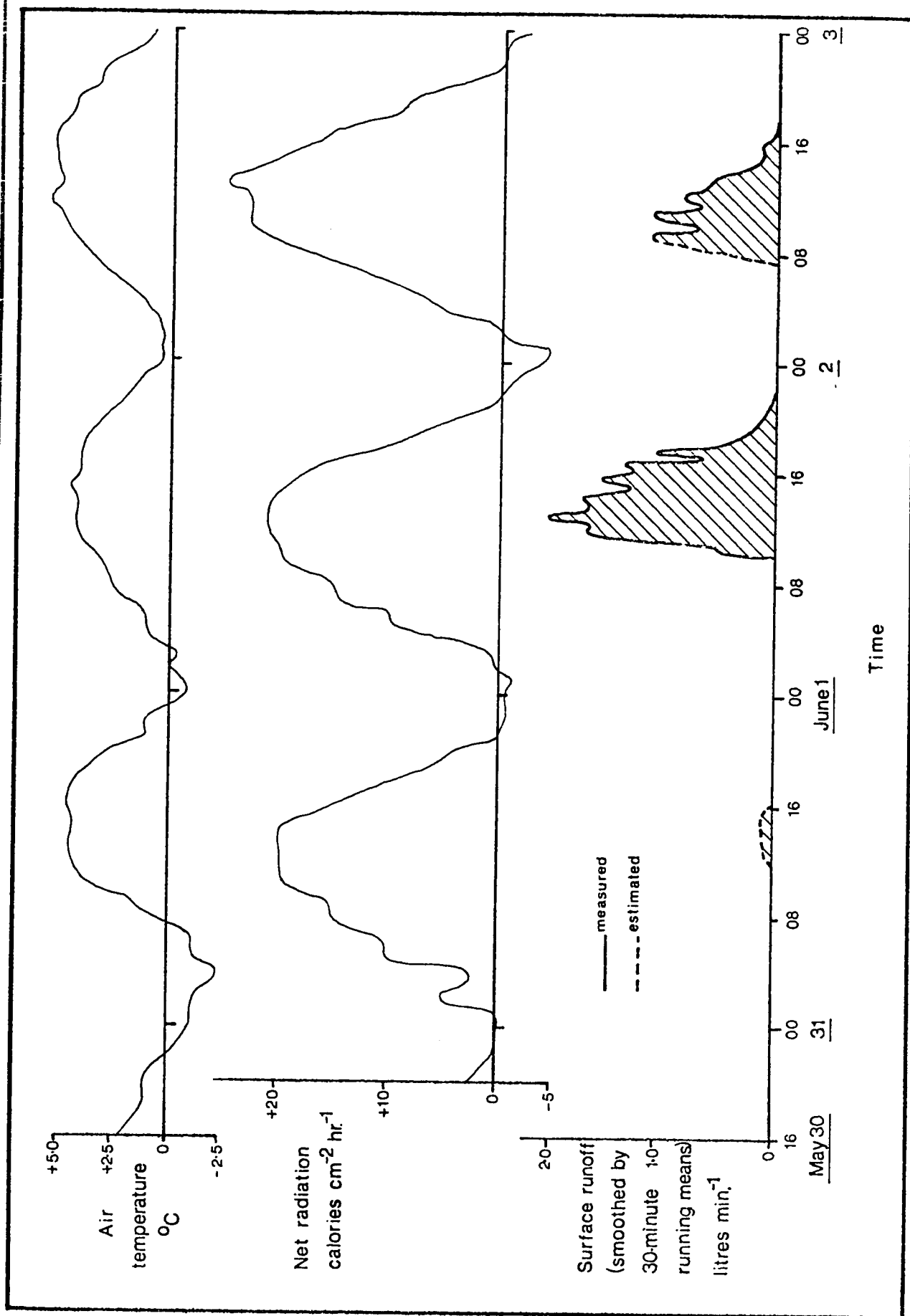


FIGURE 22: - Plot 1: surface runoff, net radiation and air temperature (May 30 - June 2).

There were clear correlations between net radiation, air temperature and surface runoff (Figure 22). Multiple regression was not feasible because of the partial correlation between air temperature and net radiation. However, net radiation alone, may be regarded as independent of surface runoff. Thus an attempt was made to complete the data record on the morning of June 2 by the regression of surface runoff against net radiation. A number of different lag times were used (Table 8).

Within the above context, the lag time represents the time taken for water released at the surface of the snow to percolate to the saturation level in the pack, and then to move horizontally through the snow and over the ground surface to the overland flow collector. The lag time showing the highest coefficient of determination (r^2) was 2.5 hours. Tests for significance of the r^2 and α and β regression coefficients were not undertaken. This was determined by the data: since random sampling was not employed, parametric statistical tests could not be used.

Normally, the vertical movement of water in the snow controls the lag time. However, Colbeck (1974) has mentioned the exception of a shallow snowpack lying on a gentle slope where the flow along the base dominates the timing of runoff. The latter was precisely the situation encountered at site 1. Because horizontal conditions had not greatly changed, it was expected that the optimum regression equation obtained from June 1 would be applicable to June 2. This assumption proved

Lag (hours)	Correlation coefficient (r)	r ²
0	+ 0.650	0.423
0.5	+ 0.790	0.625
1.0	+ 0.800	0.640
1.5	+ 0.790	0.625
1.75	+ 0.800	0.640
2.0	+ 0.814	0.664
2.25	+ 0.800	0.640
2.5	+ 0.815	0.666
2.75	+ 0.770	0.594

TABLE 8. - Plot 1: correlation coefficient values for surface runoff vs net radiation for differing lag times, (June 1, 10.30 h - 21.30 h. N=38).

N.B. As the significance of correlation coefficients cannot be evaluated, the optimum lag time identified as 2.5 hours may not be correct.

incorrect, however, as the equation under-predicted runoff: this was thought to be the result of the influence of higher air temperatures on June 2. Because of the poor applicability of the June 1 equation, the break in the discharge graph was completed by hand.

Air temperatures and values of net radiation were greater on June 2 than June 1, but the amount of surface runoff collected was less. The increased peak discharge of surface runoff per unit area of snow surface (Table 9), however, suggests that either the runoff coefficient was greater on June 2 and/or surface melt occurred more rapidly. The runoff coefficient in fact was probably less on June 2 (Table 7). The active layer was increasing in thickness at an average rate of 5 cm day^{-1} , allowing more subsurface runoff to take place, and the higher air temperatures and greater travel distance to the plot outlet augmented evaporation. It was theorised therefore, that the rate of melt was lower on June 1 than June 2 and that the smaller hydrograph produced by the latter resulted merely from a reduction in water supply as the snow-bank diminished in size.

The variations evident in the falling limbs of the two surface runoff hydrographs (Figure 22) are not attributable to radiation or air temperature variations since the changes are too rapid and too great. They may be due to different 'waves' of melt-water passing through the snow and down the slope. Since the snow was neither spatially (Figure 23) nor vertically homogeneous (ice layers were present), water produced

Date	Peak one-hour flow (litres)	Snow area (m ²)	Peak discharge/snow area (cm ³ cm ⁻² hr ⁻¹)
June 1	111.6	45	0.247
June 2	62.8	24.7	0.255

TABLE 9. - Plot 1: peak surface runoff per unit snow area.

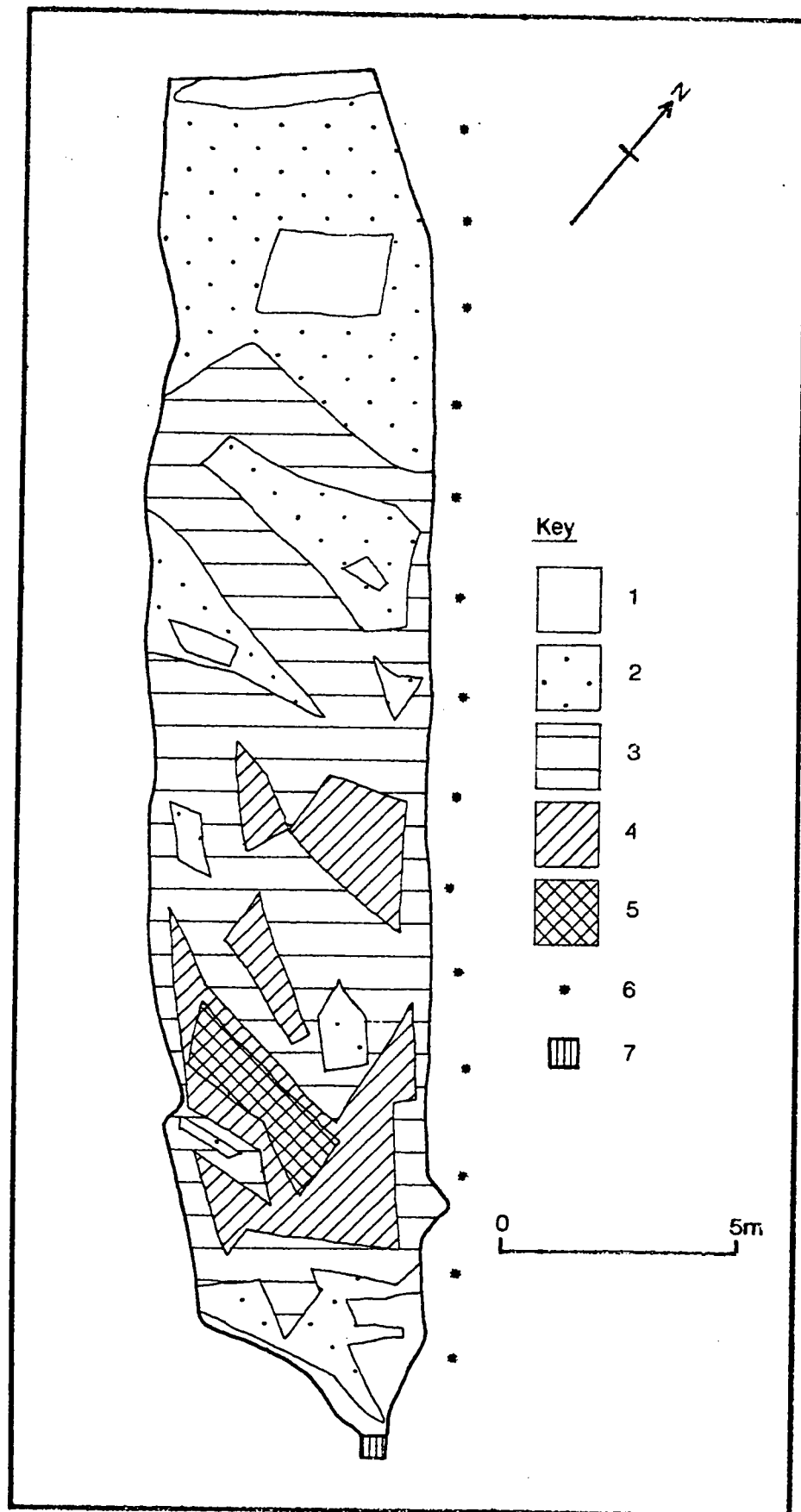


FIGURE 23: Plot 1: snowmelt pattern.

For explanation of symbols see page 70.

EXPLANATION OF SYMBOLS: FIGURE 23.

- 1 - Area snow-free before 21.00 h May 29;
- 2 - New snow-free area, 21.00 h May 29 - 21.00 h May 31;
- 3 - New snow-free area, 21.00 h May 31 - 21.00 h June 1;
- 4 - New snow-free area, 21.00 h June 1 - 10.00 h June 3;
- 5 - New snow-free area, 10.00 h June 3 - 10.00 h June 4;
- 6 - Snow-depth sampling point;
- 7 - Overland flow collector.

contemporaneously on parts of the snow surface did not necessarily reach the collector at the same time. The double peak on June 2 may have been the result of melt waves from the two major remaining areas of snow arriving at the collector.

(b) Plot 2

The second surface runoff site was set up on June 4 and monitoring commenced the following day. Some melt had already occurred at the site, but the bulk of the snowbank (127 m^3 water equivalent on the plot) remained. In all, twelve hydrographs had been obtained by June 17 when snowmelt-derived surface runoff ceased. As at site 1, there was an obvious correlation between surface runoff, prevailing weather conditions and the water supply present in the remaining snow. The daily surface runoff amounts calculated from measurements and simulations are shown in Figure 24. Simulations were based on regression against net radiation as at plot 1, but there were many more gaps in the data record, since flow often occurred for periods of 16 hours or more per day and could not be fully monitored.

Some success was achieved with the regression of surface runoff against lagged net radiation (Table 10). A trend of generally decreasing lag times was expected: as the thickness of the snowbank diminished, so the time taken for the melt-wave to reach the saturation zone should have decreased. In general, the results supported this hypothesis, but the increasing lag times from June 13 - June 17 required explanation. It was hypothesised that these were the result of an increasing ice content in the snowbank which reduced the total permeability

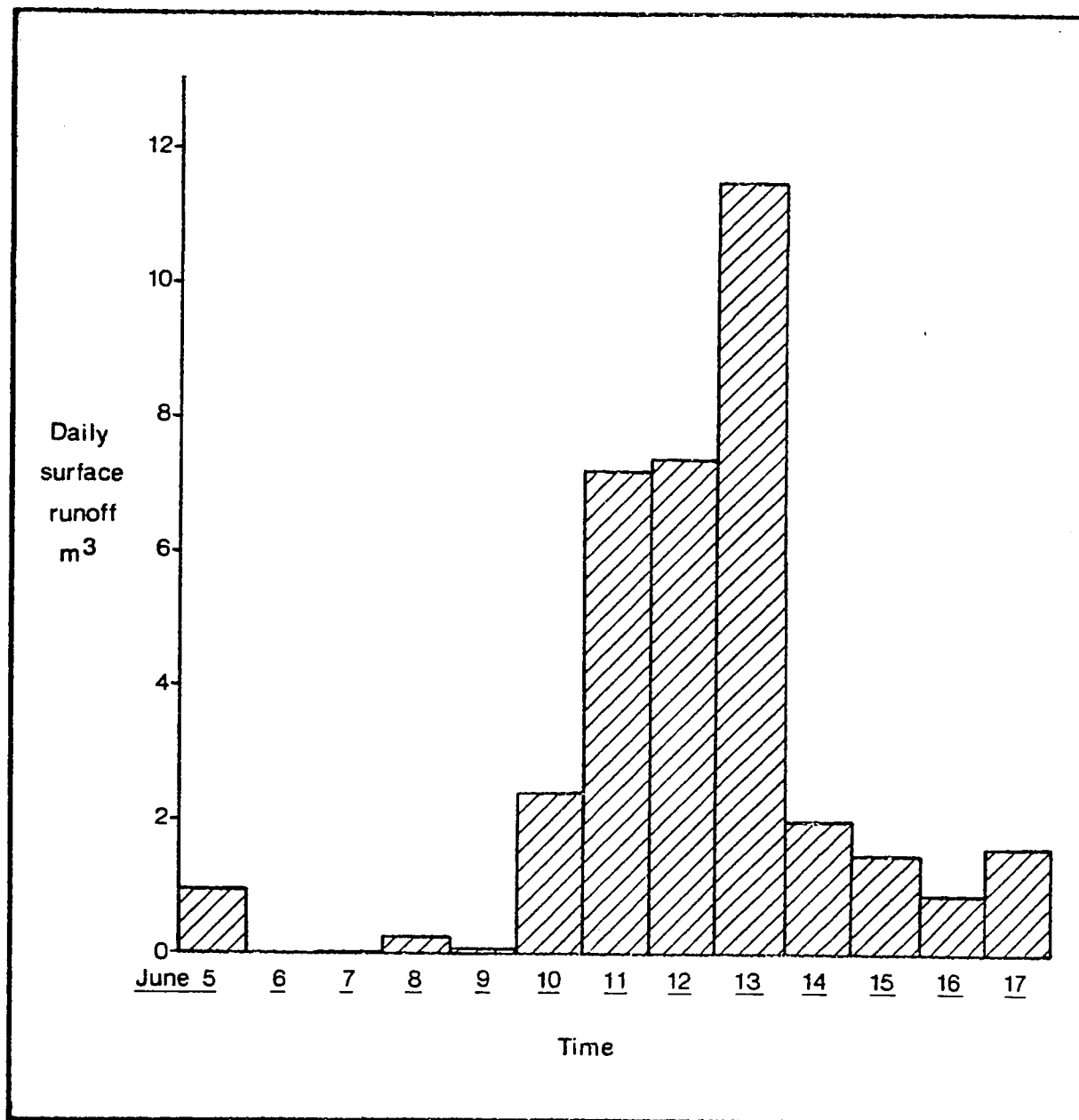


FIGURE 24: - Plot 2: bar graph of daily surface runoff tables.

Note: totals include real and simulated values.

Date	Lag time (hours)	r^2	Number of points
June 5	3.5	0.62	20
June 10	7.0	0.53	51
June 11	3.0	0.89	74
June 12 ^a	3.0	0.96	29
June 12 ^b	3.0	0.69	42
June 13	1.0	0.93	45
June 16	1.5	0.98	9
June 17	1.75	0.94	55
a Rising limb b Falling limb			

TABLE 10. - Plot 2: Coefficients of determination for
surface runoff vs lagged net radiation.

(Lags shown are those giving highest r^2 values. Regression not undertaken for days with complete hydrographs).

N.B. As the significance of determination coefficients cannot be evaluated,
the optimum lag times as identified above may not be correct.

of the snow, thereby counteracting the diminution of lag times incurred by the decrease in snow thickness. The very large lag time for June 10 was attributed to the combined deficits of radiation and snowpack water storage, caused by the poor weather of the preceding few days.

As at plot 1, non-random data collection precluded the use of significance tests on the regression and determination coefficients. Nevertheless, the high r^2 values at first suggested that the simulations of surface runoff would be accurate. However, the data are part of a time series, and it is known that serial correlation can significantly inflate r^2 values. Therefore, scatter diagrams were produced from data abstracted from surface runoff and net radiation graphs, and the regression residuals tested for serial correlation with the Durbin-Watson 'd' statistic (Durbin and Watson, 1950; 1951). No overall trends were observed in the scatter diagrams, that is, there was no general tendency for the rising or falling limbs of the surface runoff hydrographs to be over- or under-predicted. However, in every case (except June 16 when there were too few points for the test), the null hypothesis that serial correlation was not present, was rejected at the 0.01 level, and it was concluded that positive serial correlation existed. For example, for June 11 data (Figure 25), where $n = 49$ and d_1 (from tables) = 1.32 at the 0.01 level, d was calculated as 0.01. Since $d < d_1$, H_0 was rejected and it was concluded that positive serial correlation existed.

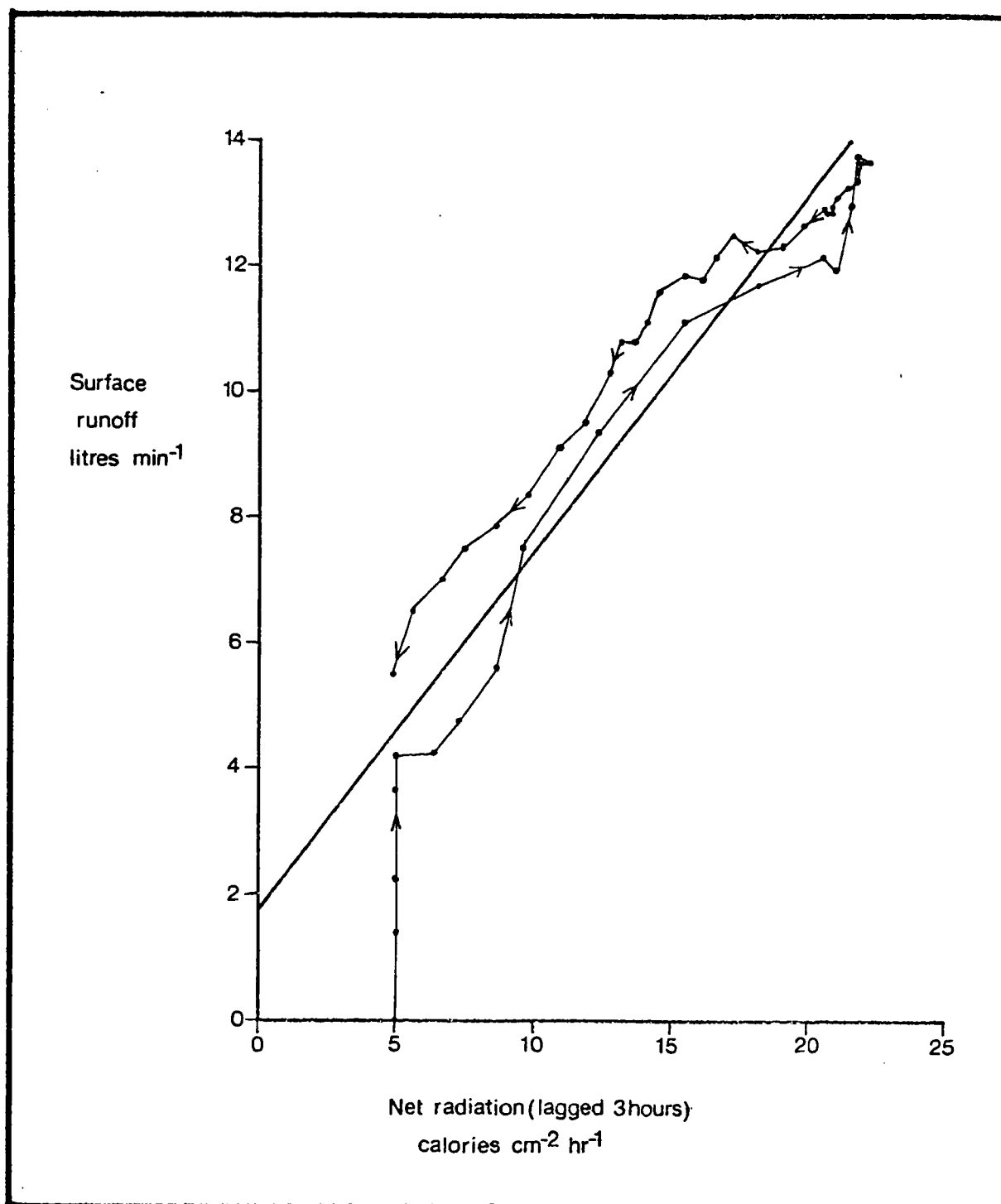


FIGURE 25: - Plot 2: surface runoff vs net radiation scatter diagram and regression line (June 11).

Note: Regression line is: surface runoff = $0.573 R_n + 1.815$, where R_n is lagged 3 hours.

Serial correlation meant that the data points were not normally distributed around the regression line. Simulation by regression, therefore, contained the possibility of considerable inaccuracies. However, this was still considered to be the best method to fill in the breaks in the surface runoff record and since the major point of the exercise was to obtain daily and seasonal totals of surface runoff, rather than instantaneous discharges, overall accuracy probably remained within the target range of $\pm 10\%$.

The daily graphs of surface runoff, net radiation and air temperature show recurring relationships between the variables. June 11 was a typical example. Net radiation values became positive at 04.00 h, increased rapidly when overcast skies cleared at 10.00 h, and reached a peak at 12.00 h (Figure 26). A more gradual decline then occurred until net radiation again became negative at 22.00 h. Air temperatures followed a similar pattern of build-up and recession, but the peak value of $+11.1^{\circ}\text{C}$ occurred three hours after that of net radiation. Surface runoff began at 10.00 h and increased steadily for five hours reaching the maximum recorded discharge at plot 2 of $13.85 \text{ litres min}^{-1}$. Flow then declined and the site was abandoned at 22.15 h. Simulation of the remaining curve suggested that flow ceased at 01.15 h on June 12. Estimation of the water equivalents of the snowpack revealed that surface runoff represented only 52.9% of the calculated changes that took place between June 11 and June 12 (Table 11): the remainder must have been lost as intra-vegetational wash,

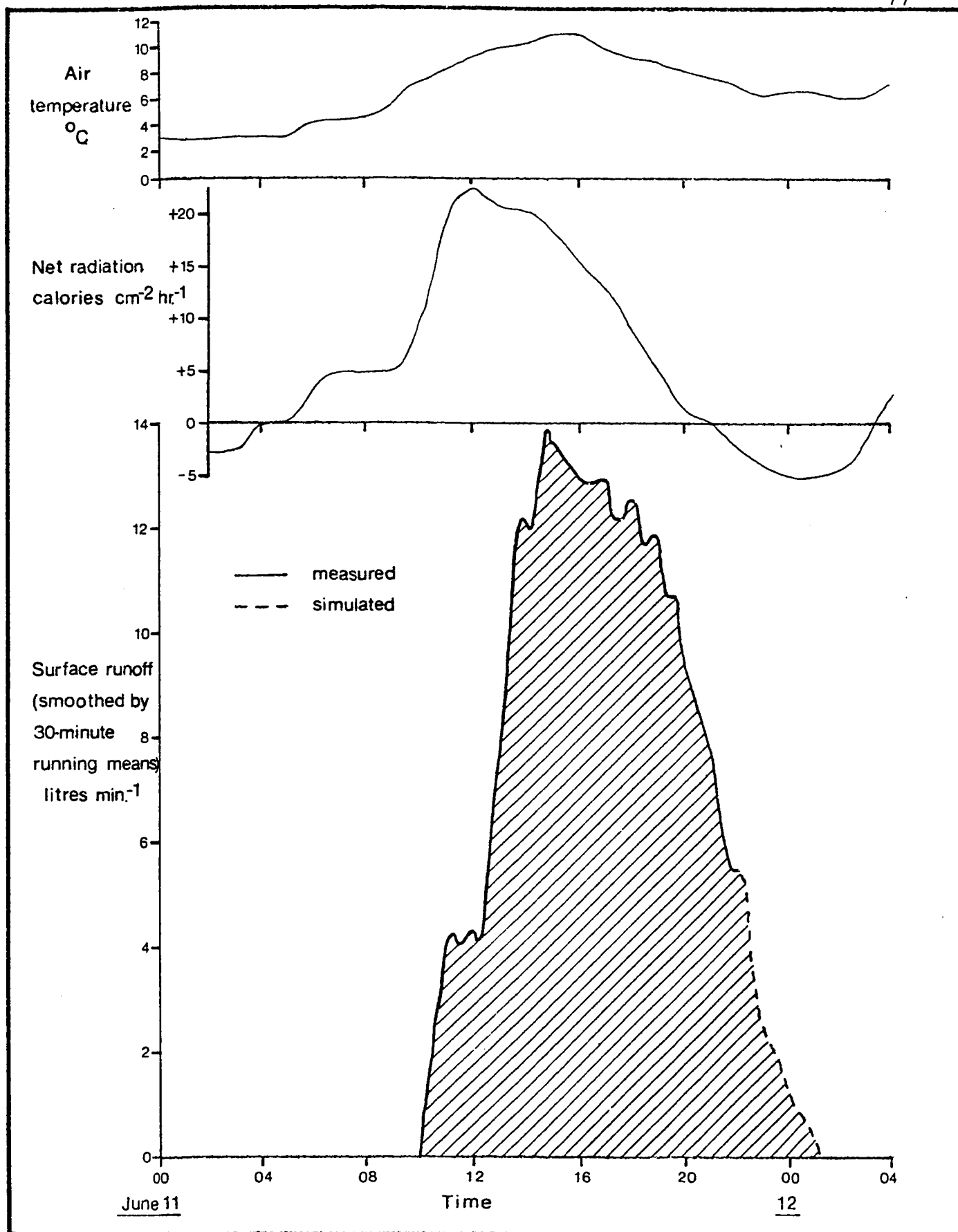


FIGURE 26: - Plot 2: surface runoff, net radiation and air temperature (June 11).

Simulation of surface runoff used the following formula: surface runoff = $0.573 R_n + 1.815$, where R_n is lagged 3 hours, $N=76$.

Timing of Observations	Time elapsed (hours)	Loss of snowbank water equivalent (m ³)	Surface runoff (recorded and simulated) (m ³)	Surface runoff/loss of water equivalent (%)
13.00 June 5 - 13.00 June 7	48	17.5	0.93	5.2
13.00 June 7 - 11.00 June 11	94	37.4	2.78	7.4
11.00 June 11 - 17.00 June 12	30	20.9	11.08	52.9
17.00 June 12 - 19.00 June 13	26	13.4	13.40	100.0
19.00 June 13 - 17.00 June 14	22	14.8	2.93	19.7
17.00 June 14 - 15.00 June 15	22	12.1	0.99	7.5
15.00 June 15 - 18.00 June 16	27	5.8 ^a	1.60 ^b	27.4
18.00 June 16 - 18.00 June 17	24	5.6	1.78	31.7
Total water equivalent of snowbank = 127.5 m ³ (including 2.4 m ³ rain).				
Total surface runoff = 35.4 m ³				
Percentage water loss as surface runoff = 27.8%				
a Includes 2.4 m ³ rain				
b Includes 0.64 m ³ derived from rain				

TABLE 11. - Plot 2: surface runoff production (June 5 - June 17).

throughflow and evaporation. Table 11 shows the other surface runoff coefficients compared with snowmelt. The figure of 100% for June 12-13 is clearly anomalous, and may be the result of either the underestimation of water equivalent changes, or the overestimation of the simulated surface runoff. Peak production of surface runoff per unit area of snow did occur on June 13, but as a result of the apparently high runoff coefficient, it is impossible to say whether this was the result of a melt-rate maximum or more of the surface runoff produced reaching the collector (Table 12).

The lag time between net radiation and surface runoff was attributed to the vertical and horizontal movements of water in the snowbank, combined with the flow of water over the ground surface to the collector. While it was not feasible to measure the speed of intra-snow water movement, the utilisation of rhodamine dye allowed an estimate to be made of average ground surface water velocities. The speed of flow was recorded as 1.3 cm sec^{-1} . This was low in comparison with normal surface runoff velocities ($5\text{--}8 \text{ cm sec}^{-1}$ (Kirkby, 1969)). Visual observations suggested that it was the result of a diminution of velocity as the water passed through the moss bands on the plot. Using the figure obtained, only 14 minutes out of the 3-hour lag on June 11 could be attributed to the time taken for water to pass over the ground surface from the snow edge to the collector. Thus over 90% of the lag-time probably was associated with the passage of water through the snowbank.

Date	Peak one-hour flow (litres)	Snow area (m ²)	Peak flow/snow area (cm ³ cm ⁻² hr ⁻¹)
June 5	265.6	377.2	0.070
June 7	3.9	366.1	0.001
June 8	214.8	356.1 ^a	0.060
June 9	14.2	346.1 ^a	0.004
June 10	266.1	336.1 ^a	0.080
June 11	814.8	326.8	0.249
June 12	760.1	299.6	0.254
June 13	748.5	272.5	0.273
June 14	359.8	252.5	0.144
June 15	80.9	224.9	0.035
June 16	134.5	199.4	0.070
June 17	186.2	119.5	0.154
a Estimate			

TABLE 12. - Plot 2: peak surface runoff per unit snow area (June 5 - June 17).

(c) Plot 3

Plot 3 was monitored for surface runoff between June 20 - June 25, but the amounts collected represented only a very small proportion of the melt at the site during the season. Daily surface runoff totals are not presented as regression against lagged net radiation produced low r^2 values (for example, the best r^2 value to be calculated from the June 22 data was 0.23), so that gaps in the surface runoff graphs could not be filled in. By themselves, therefore, the data from plot 3 are not very useful, but they still provide a basis for comparison with other plots and processes.

The comparison of peak hourly surface runoff per unit area of snow surface at all three plots (Tables 9, 12 and 13) reveals that plot 3 was subjected either to a much more rapid surface lowering, or had a greater runoff coefficient than the other two. Unfortunately, the low r^2 values also prevented the calculation of runoff coefficients for plot 3. However, no obvious reason suggested itself as to why this plot should possess a runoff coefficient significantly different from those of the other two, and it was hypothesised therefore that the variations were the result of different energy transfers. This, together with the poor correlation of melt and lagged net radiation at plot 3, was in keeping with the results of Gray and O'Neill (1974) who consider turbulent heat transfer to be the most important energy supply to isolated snowbanks. It was probably not a coincidence that the maximum rate of surface runoff per unit area of snow occurred on June 22, when air

Date	Peak one-hour flow (litres)	Snow area (m ²)	Peak flow/snow area (cm ³ cm ⁻² hr ⁻¹)
June 20	99.3	25.4	0.391
June 21	224.7	23.4	0.958
June 22	250.6	18.4	1.364
June 23	34.8	14.3	0.245
June 24	57.0	11.5	0.494
June 25	27.2	8.85	0.308

TABLE 13.- Plot 3: peak surface runoff per unit snow area (June 20-June 25).

temperatures were high ($+9^{\circ}\text{C}$), the sky was clear and there was a strong wind (estimated at 30 km per hour): these combined to give large heat fluxes both of radiant energy and turbulent heat.

(d) Rainfall

In the High Arctic, snowmelt is the major source of water for surface runoff, with the other possibilities being active layer thaw and summer precipitation. As noted in Chapter 2, the former is not thought to be an important contributor of water, but the latter did give rise to surface runoff on two occasions at the study sites. For example, a storm of 6.25 mm in five hours occurred on June 15 during the period when plot 2 was being monitored. The resultant hydrograph (Figure 27) was separated by regression into runoff thought to be result of snowmelt (the hydrograph peak) and that derived from precipitation ('baseflow'). Evaluation of the amounts involved (Table 11) showed that 640 litres of runoff were due to rain, representing a runoff coefficient of 26.4%.

It was thought that the plot source areas for surface runoff resulting from rainfall would consist of the snowbank itself and the saturated area between it and the overland flow collector: on June 15 these sections of the plot accounted for 85% of its area. As the runoff coefficient was less than one third of this figure, it follows that rain falling on the snowbank failed to cause runoff and/or that the water was 'lost' as evaporation and increased subsurface runoff. The rapid

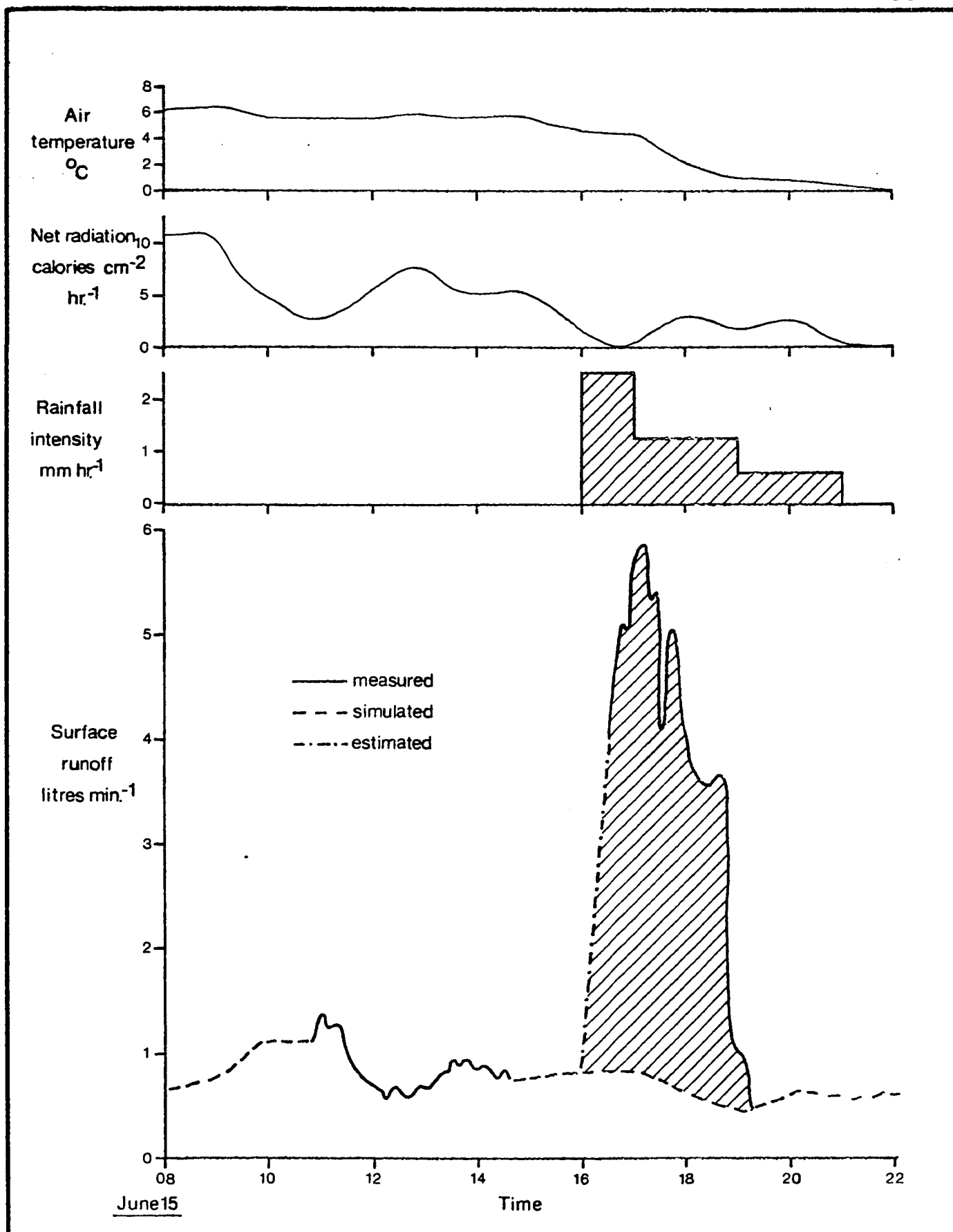


FIGURE 27: - Plot 2: rainfall and runoff (June 15).

Simulation of surface runoff used the following formula: surface runoff: $0.06 R_n + 0.48$, where R_n is lagged 2 hours, $N=25$.

response time of runoff to precipitation at the beginning of the storm and the equally rapid fall-off at the end suggested that most water was stored in the snowbank rather than passing through.

The largest rainstorm of the summer (7.5 mm) occurred on June 23 between 00.00 h - 03.00 h. At plot 2, surface runoff had ceased four days earlier on June 17 and the remaining snow had ablated on June 18. In spite of this, surface runoff occurred at the plot. The flow, which was not recorded, was related probably to antecedent conditions: the continuing high soil moisture contents on the plot allowed saturated overland flow to take place. In contrast, neither the precipitation event of June 15 nor that of June 23 induced surface runoff at plot 1 due to the large soil moisture deficits and the thicknesses of the thawed layer (38 and 48 cm respectively). The instance of surface runoff at plot 2 after the end of snowmelt may have been relatively rare: if the storm had taken place a few days later or had been of a smaller magnitude, it is probable that surface runoff would not have occurred.

The theory of partial area contribution to runoff (see pp. 3 - 4) states that only parts of a drainage basin contribute water to the storm hydrographs of streams. These contributory areas are generally contiguous with the channels themselves, as surface runoff is considered to be the major component of the resultant channelised flow. The source areas are variable in extent, depending on soil moisture conditions and may expand during a storm. Saturated, rather than Hortonian,

overland flow is the norm in humid environments: rainfall intensity does not usually exceed infiltration capacity, and surface runoff occurs where the regolith water-table reaches the surface. The infiltration of precipitation is prevented and subsurface water moving downslope is forced on to the surface.

The observations of surface runoff that were made at the three plots in this study did not support the hypothesis that Hortonian overland flow is normal in permafrost regions (e.g. Freeze, 1972). Clearly, it would have been very rare for surface runoff derived from precipitation to be produced at interfluvial locations such as plot 1, where only a thin snow cover is present at the beginning of summer and the active layer thickens rapidly. Rather, as might be suspected intuitively, the production of surface runoff and consequent streamflow in the Thomsen Fly Camp area is strongly influenced by snow distribution. Snowmelt itself produces the vast majority of surface runoff and any remaining amounts are the results of rain falling on areas already in wet conditions. The contributory areas usually are made up of the snowbanks themselves and the saturated areas downslope from them. Only in very exceptional storms and aided by the presence of impermeable permafrost, do other areas of the basin become saturated.

Subsurface runoff

One of the aims of the study was to provide information regarding the relative magnitudes and frequencies of surface and subsurface wash processes operating in a permafrost

environment. This objective was only partially achieved, however, for two main reasons. First, the results obtained for subsurface runoff may have been influenced considerably by disturbance associated with the installation of the measuring apparatus itself. Second, all of the subsurface runoff may not have been successfully diverted laterally into the collecting pipes. Further, the interpretation of subsurface flow is more complex than that relating to surface runoff, necessitating a consideration of the interaction not only of water supply factors, but also of hydraulic gradients and permeabilities.

Subsurface runoff was collected at a number of depths at sites 1 and 3, contemporaneously with surface runoff (Table 4). A lesser degree of confidence can be placed on the results obtained from the collectors at site 1 since these were the first to be installed and greater surficial disturbance occurred.

(a) Site 1

The collectors were inserted on May 31 and records began on June 1. In spite of the guttering being in place until July 17, only 3 days of flow (June 1-3) were recorded in the upper collector (Figure 28). Amounts were small (142.0 litres), but these were not the total flows that occurred, since gaps existed in the record. There was no feasible way of filling in the data breaks, nor was it possible to estimate the amount of flow that had taken place before June 1. For these reasons, few detailed comparisons with surface runoff are possible.

In general terms, amounts of subsurface runoff collected at an average depth of 6 cm at site 1 (gutter 1) followed a

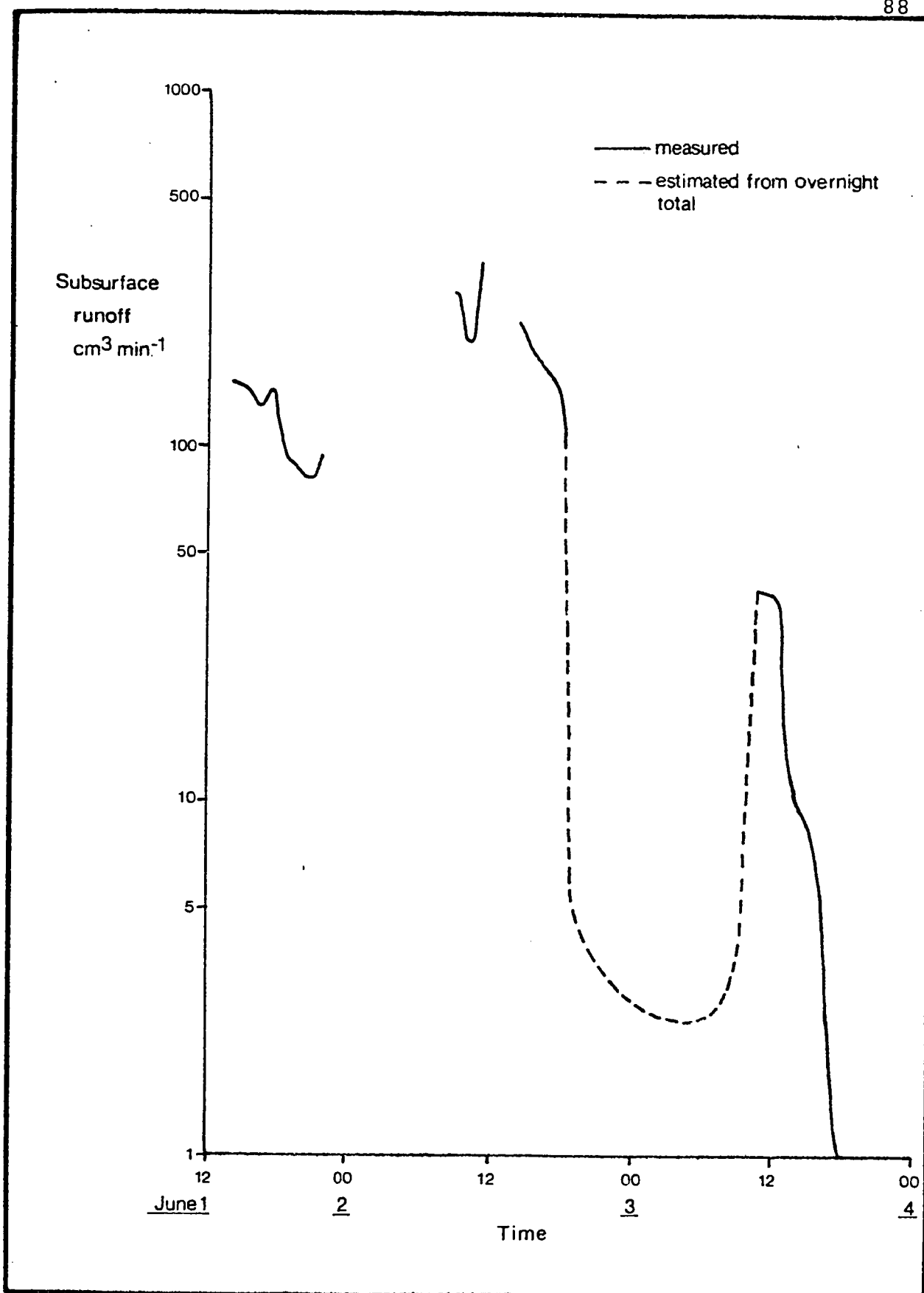


FIGURE 28: - Site 1: subsurface runoff, gutter 1 (June 1 - June 3).

diurnal pattern, being greatest during the day and decreasing at night. This suggested that subsurface flow rates were responding to snowmelt. Further support for this hypothesis was found in the rapid decline and cessation of throughflow on June 3, as the remaining snow upslope of the collector ablated and disappeared. The steep hydrograph recession was indicative of a further point: after the end of snowmelt, influenced by a rapidly dropping frost-table and surface evaporation, the water-table quickly dropped below the lowest point of the top collecting gutter. Despite rainfall later in the season, this layer did not become saturated again.

The maximum rate of subsurface runoff observed was 0.33 litres min^{-1} on June 2 but this value is not considered meaningful without an examination of the contributory slope width. The subsurface runoff collector was 0.5 m wide, but an examination of frost-table contours (Figure 29), revealed that the contributory width of slope was actually 2 m on June 3. Assuming that the situation had been the same on June 2, the maximum rate of subsurface runoff per unit width of slope was 0.17 litres $\text{min}^{-1} \text{ m}^{-1}$, compared with a maximum surface runoff rate of 0.4 litres $\text{min}^{-1} \text{ m}^{-1}$.

Very little subsurface runoff was collected from the second level of guttering (average depth 16 cm) at site 1. Only 1.4 litres had been recorded by June 4 when flow stopped. The small amounts probably resulted from the limited drop in height of the flexible tube that ran between the gutter and the storage container. However, even if some proportion of

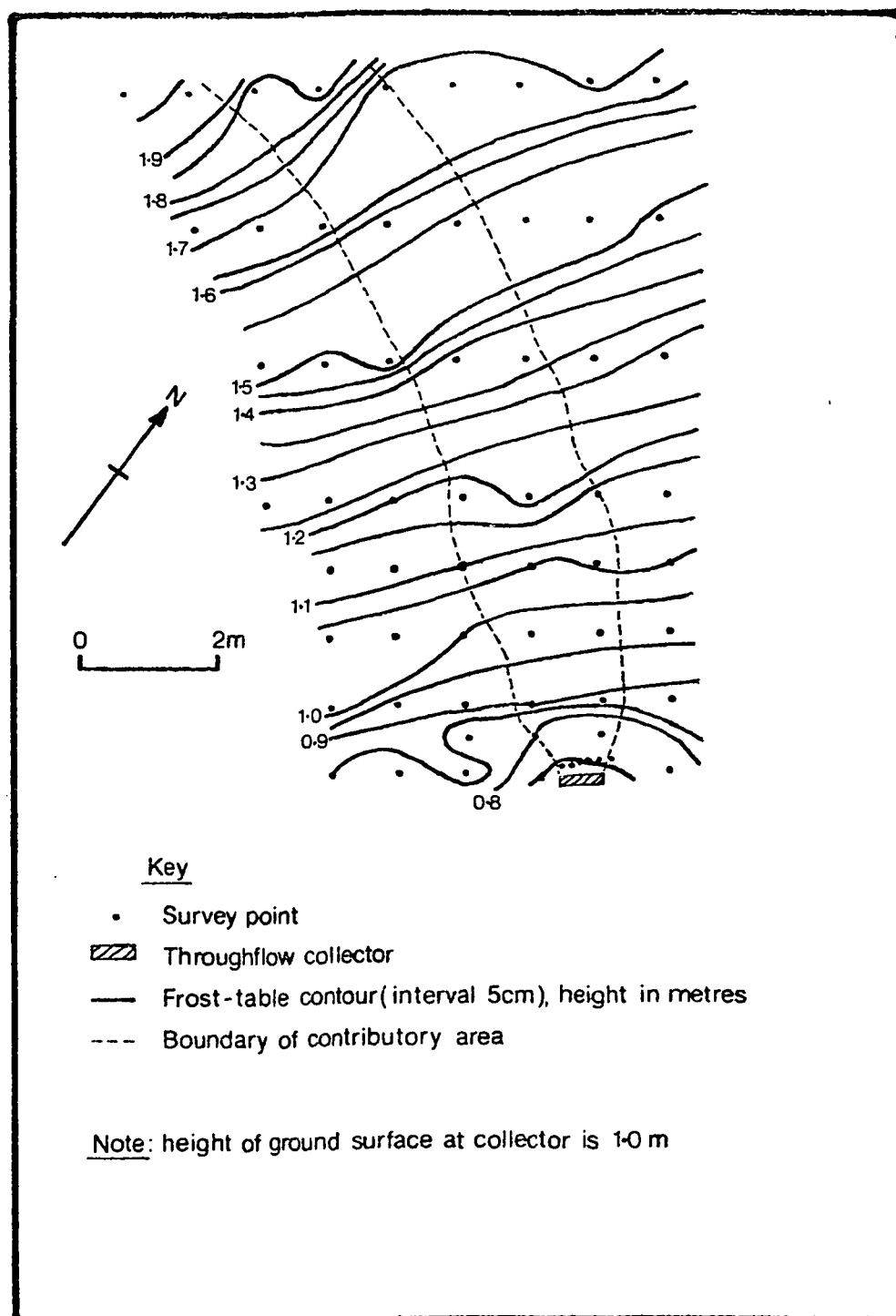


FIGURE 29: - Site 1: throughflow collector contributory width (June 3).

flow was not collected, its cessation on June 4 indicated that rapid recession was normal at depth, as well as close to the surface.

No flow was recorded at a depth of 31 cm during the snowmelt period because this soil layer was still frozen. Even after it had thawed, however, only 0.01 litres were recorded up to July 17. This single event followed the rainfall of June 27. No flow was recorded at the upper levels either at that time or in response to any of the other precipitation events, indicating that as a result of large moisture deficits, these zones did not become saturated after the snowmelt period.

If breaks in the record are taken into account, it is probable that the magnitudes of surface and subsurface runoff are similar at site 1. However, their relatively small amounts and short durations suggest that at interfluvial locations, neither is important in the generation and sustenance of streamflow.

(b) Plot 2

Subsurface runoff was not measured directly at plot 2, but through the assessment of some of the other factors in the slopewash system, estimates could be made of its likely importance. Evapotranspiration from the snow surface (Chapter 2 and Table 14) and evapotranspiration from the ground surface (Table 15) appeared to be of similar orders of magnitude. Both seemed to be relatively unimportant except at the beginning and end of the melt period, and accounted for losses of only 5.1 m^3 (4%) of the water in the snowbank. Since 28% of the snowbank was

Time	Average loss of weight in containers (g)	Average area of snow (m ²)	Average evapsublimation rate (mm/day)	Total evapsublimation (m ³)	Percentage of total snow melt
13.00 June 5 - 13.00 June 7	53.5	374.2	1.70	1.27	7.2
13.00 June 7 - 11.00 June 11	24.1	343.9	0.77	1.06	2.8
11.00 June 11 - 17.00 June 12	17.0	313.2	0.54	0.17	0.8
17.00 June 12 - 19.00 June 13	n.d.	286.1	n.d.	n.d.	n.d.
19.00 June 13 - 17.00 June 14	7.0	262.5	0.22	0.06	0.3
17.00 June 14 - 15.00 June 15	n.d.	238.7	n.d.	n.d.	n.d.
15.00 June 15 - 18.00 June 16	n.d.	212.2	n.d.	n.d.	n.d.
18.00 June 16 - 18.00 June 17	71.0	159.5	2.26	0.36	6.1
18.00 June 17 - 18.00 June 18	158.6	119.5	5.05	0.60	6.4
Total calculated				3.52	2.7

TABLE 14. - Plot 2: evapsublimation losses (June 5-June 18).

Time	Average saturated area ^a (m ²)	Average unsaturated area ^b (m ²)	Saturated area evaporation rate (mm/day)	Unsaturated area evaporation rate (mm/day)	Total evaporation (m ³)	Percentage of total snowmelt
13.00 June 5 - 13.00 June 7	3.8	9.5	1.62	1.29	0.04	0.2
13.00 June 7 - 11.00 June 11	25.4	18.2	1.73	1.37	0.07	0.2
11.00 June 11 - 17.00 June 12	46.2	26.1	3.16	2.51	0.26	1.2
17.00 June 12 - 19.00 June 13	58.9	42.5	1.89	1.50	0.19	1.4
19.00 June 13 - 17.00 June 14	68.7	56.3	1.85	1.47	0.19	1.3
17.00 June 14 - 15.00 June 15	83.8	65.0	0.65	0.52	0.08	0.7
15.00 June 15 - 18.00 June 16	87.8	87.5	1.33	1.05	0.23	4.0
18.00 June 16 - 18.00 June 17	109.2	118.8	2.46	1.95	0.50	8.9
Total					1.56	1.2
a Area downslope of snowbank and upslope of overland flow collector ($\alpha = 1.26$).						
b Area upslope of snowbank and downslope of plot edge ($\alpha = 1.0$).						

TABLE 15. -Plot 2: ground evaporative losses (June 5-June 17).

removed as surface runoff (Table 11) it follows that approximately 65% of the water left the plot as subsurface runoff. This figure, if correct, was much higher than that obtained for site 1 where approximately 10% of the water equivalent of the snowbank was thought to have taken a subsurface routing.

(c) Site 3

Subsurface runoff was collected at site 3 with one gutter next to the runoff plot and three gutters located 150 m downslope. The ground had thawed to depths that allowed free flow to all gutters for the whole period of study (June 21 - July 27) except for the lowest gutter at the downslope site. The latter site is considered first.

The most obvious feature of the hydrograph of the upper gutter (gutter 1) (Figure 30) is its rapid decline. Flow decreased by a factor of 10^4 , from a maximum of 6.2×10^{-2} litres $\text{min}^{-1} \text{ m}^{-1}$ slope width (9.3×10^{-2} litres min^{-1} recorded) to 6.6×10^{-6} litres $\text{min}^{-1} \text{ m}^{-1}$ (1.0×10^{-5} litres min^{-1} recorded) between June 22-29. The contributory slope width to the collector was 1.5 m. The decline was attributed to the reduction in the supply of water derived from the snowbank upslope: by June 27 the snow had ablated completely.

Superimposed over the general downward trend were daily fluctuations. Although there was a general correlation between the maxima and minima of surface runoff recorded at plot 3, and subsurface runoff collected downslope for June 22 and 23, this was not the case for later dates.

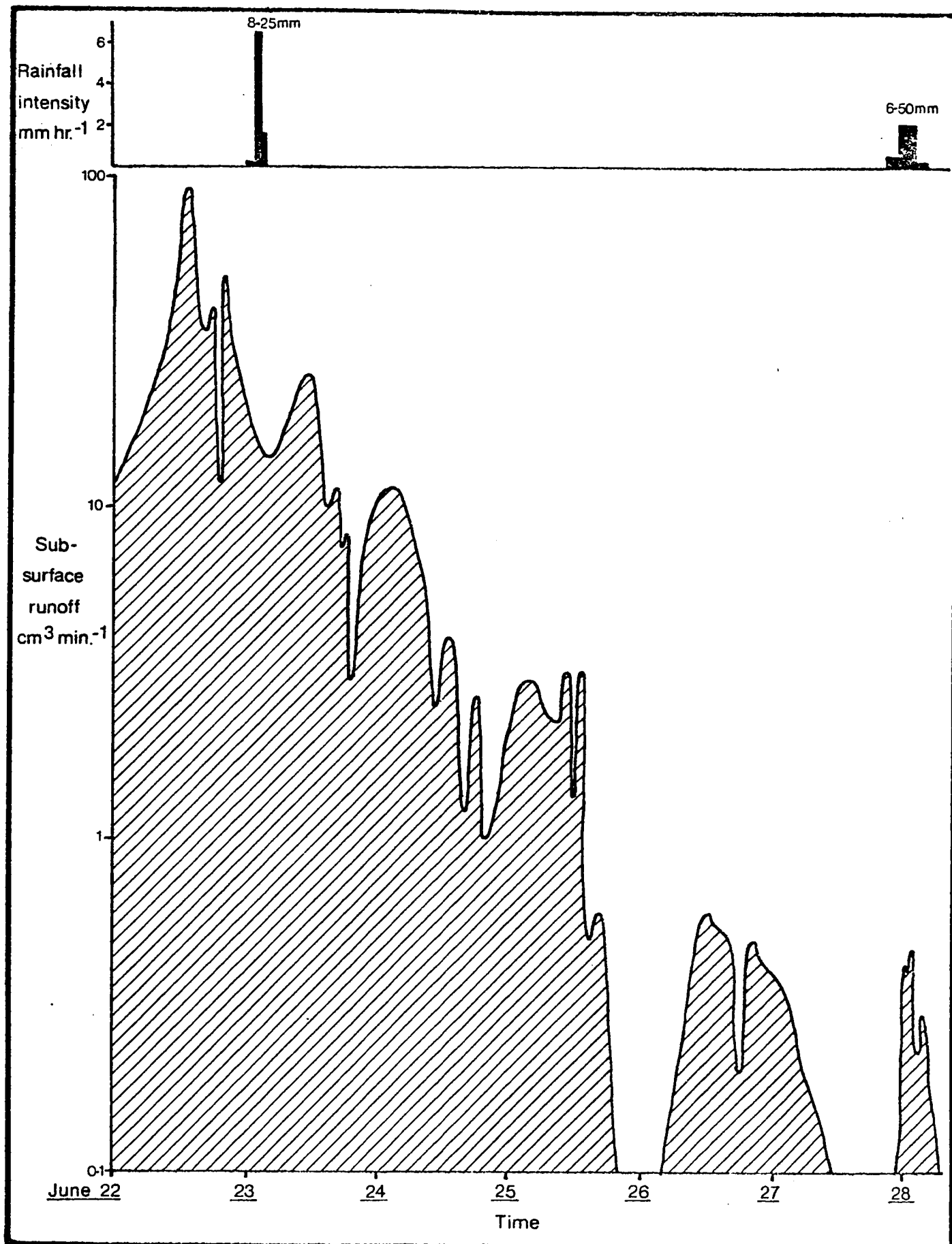


FIGURE 30: - Site 3: subsurface runoff, gutter 1 (June 22 - June 28).

The subsurface runoff hydrograph rose on June 28, contrary to the general trend (Figure 30). This was presumably related to the precipitation input. The response was rapid, indicating that the thawed layer water-table was close to the gutter level. Water must have reached the collector as a 'wave', beginning with that derived from the area immediately upslope, and continuing with water from further away. The apparent lack of response to the earlier precipitation event of June 23 was simply a function of the very different magnitudes of flow that were taking place.

Gutter 2 at the downslope throughflow site collected only 0.24 litres during the period of record, compared with 80.23 litres from gutter 1. Excavation of the trench containing the flexible tubing at the end of the season revealed that there was sufficient gradient for flow to occur and no constrictions were evident in the pipe. No explanation is offered for this low flow.

Gutter 3, at an average depth of 31 cm, produced the longest record of continuous flow, but at low rates. Over a 35 day period, a total of only 6.85 litres of water was collected. The graph of the average daily rates of flow (Figure 31) showed a broadly increasing trend between June 23 - July 3, and a decline between then and July 23. On the latter date, precipitation caused flow rates to increase and they remained at a high level until the record ended four days later. Superimposed over the trends were considerable daily variations, but no diurnal pattern was detectable. The maximum average

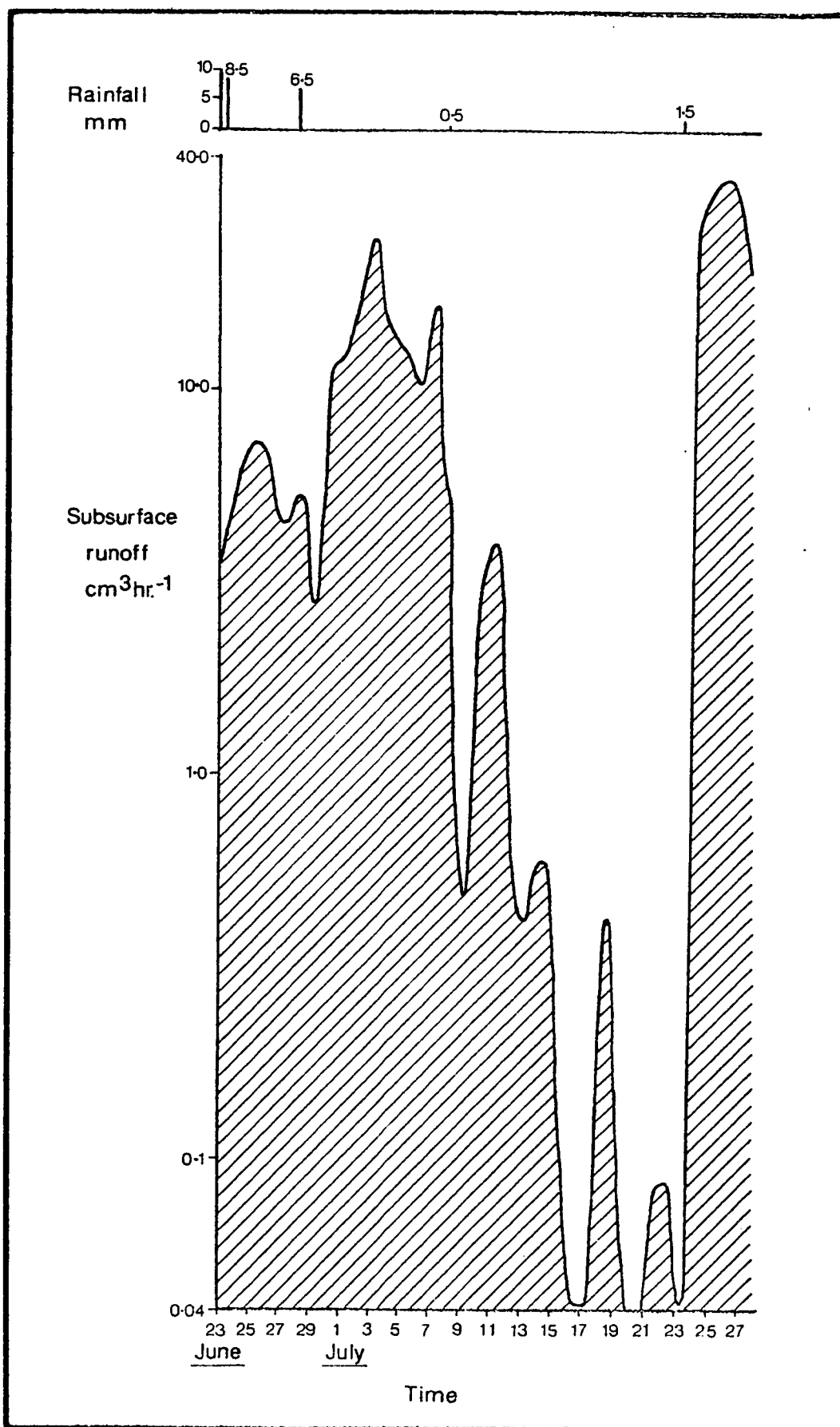


FIGURE 31: Site 3: subsurface runoff, gutter 3 (June 23 - July 28).

hourly flow rate of 4.1×10^{-4} litres $\text{min}^{-1} \text{ m}^{-1}$ slope width (6.1×10^{-4} litres min^{-1} recorded) occurred on July 1, but the maximum daily total of flow was on July 26 (0.85 litres recorded) in response to rainfall.

The low rates of flow at the 31 cm depth may reflect a low coefficient of permeability existing in the material. However, a further reason had to be sought to explain the general rise and decline in the hydrograph. Three possibilities were examined. The first was that the trend was merely a function of the technique of collection. Although the average gutter depth was 31 cm, the maximum depth was 34 cm, the minimum 28 cm, and the outlet pipe was at 39 cm. Thus it was possible for part of the gutter to collect water while other parts of the collection apparatus were still frozen. The rise of the hydrograph then, represented the progressive downward movement of the frost-table, from a partial thawing of the collector to full thaw. The decline occurred as the thawed-layer water-table also moved downward, leaving first the upper part and ultimately all the gutter above the saturated zone.

The second possible explanation of the overall rise in the trend was that it was caused by the slow passage downslope of water derived from the snowbank. This contained the implication of very low velocities of water movement since the last snow ablated on June 27, but the peak subsurface runoff rate at that time, was not reached until July 3.

The third hypothesis was that the rise and fall in the hydrograph were responses to variations in the release of

water by ground ice thaw. Neither this nor the second hypothesis, however, could explain the lack of response to the two rainfall events of June 23 and June 28. On the other hand, the first hypothesis provided the interpretation that this lack of response was the result of most of the flow passing through the regolith above gutter 3 which was still partially frozen. Hence the first hypothesis that the rise and fall was the result of the collection technique was considered to be the most probable explanation.

The variations superimposed on the trend were the result of the interaction of a number of factors: the slow passage of varying amounts of snowmelt-derived water during the period up to June 27, the possible melting of ground ice upslope, evapotranspiration variations, and inadequacies in the measurement method.

The collector at the top of the slope at site 3 (gutter 4), was situated next to the runoff plot, at an average depth of 6 cm. The flow probably resembled that taking place under the plot itself since the source areas were only slightly different. There were several interesting points about the data record, the most important being the small quantities of water collected, 6.62 litres between June 22 - July 19. This represented $12.61 \text{ litres m}^{-1}$ slope width (contributory width was 0.52 m) compared with recorded surface runoff on the plot of $1089.36 \text{ litres m}^{-1}$ slope width over the same time period (no estimate made for missing data). Further, peak average one-hour subsurface runoff was $2.69 \times 10^{-3} \text{ litres min}^{-1} \text{ m}^{-1}$

slope width compared with $2.98 \text{ litres min}^{-1} \text{ m}^{-1}$ for surface runoff. Clearly subsurface runoff at this point on the slope and at this level was relatively unimportant, and accounted for only about 1% of total downslope water movement.

A trend of relatively stable subsurface runoff rates was exhibited in the records for gutter 4, followed by a rapid decline from June 28 onwards (Figure 32). The variations superimposed on the trend were diurnal in form, and flows were generally greater during the day than at night. However, it appeared that supply factors were not the only ones influencing the results.

The movement of water through the soil depends on its permeability and applied hydraulic gradient (Darcy's Law), as well as the change in moisture storage in the soil (Continuity equation) (Knapp, 1974). It was hypothesised that the maximum possible hydraulic gradient above collector 4 slowly decreased between June 22-27. With a relatively constant permeability, this would have led to a slow reduction in the maximum possible amount of subsurface runoff. The nightly drops below these levels, occurred as water drained out of the soil while incoming supplies of water had ceased. The decrease in the hydraulic gradient may have been due to an increase in the thickness of the thawed layer or may have been caused in some way by the retreat and ablation of the snowbank.

A comparison of totals of subsurface runoff collected in gutters 4 and 1, the former at the top of the slope and the latter at the base, revealed that the upper level subsurface

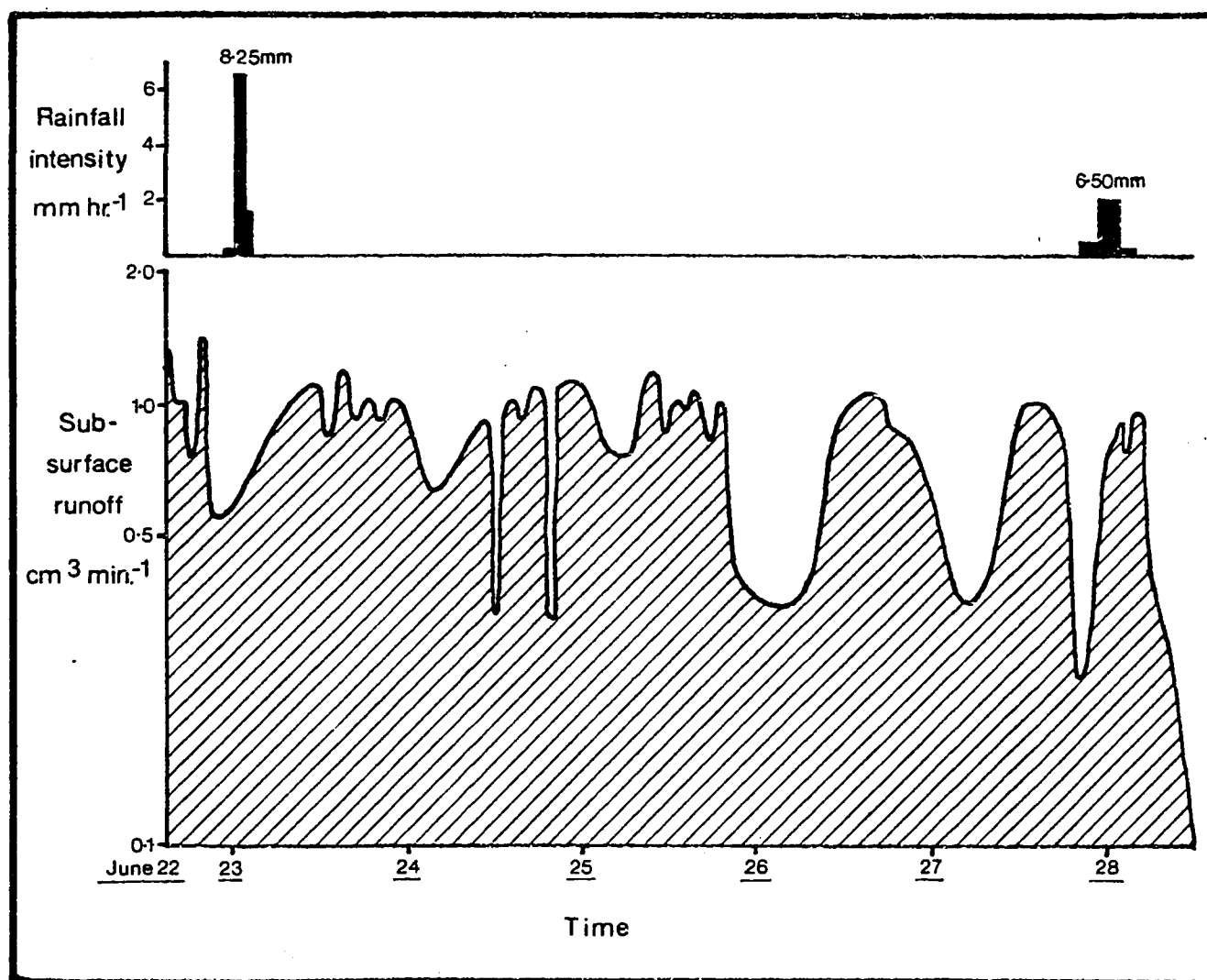


FIGURE 32: - Site 3: subsurface runoff, gutter 4 (June 22-June 28).

flow volumes increased more than four times downslope, from 12.6 - 53.4 litres m^{-1} slope width. The most probable explanation of this was that downslope from gutter 4, the thick and continuous vegetation mat enabled the progressive conversion from surface to subsurface runoff, thus illustrating the difficulty of distinguishing in the field between the two types of slope water movement.

Runoff

The complex nature of the slope runoff system is revealed through the data collected at the slopewash sites. Subsurface runoff rates vary vertically at any one position on the slope. These variations occur during and after active layer thaw, due to frost-table and accompanying water-table movements, and moisture recharge after precipitation events. The relative proportions of surface and subsurface runoff rates also vary with time at any one position on the slope. Further, subsurface runoff as a percentage of total downslope water movement, varies from place to place on the slope. Clearly more and better comparative data are needed before predictive modelling of the slope runoff system can be undertaken.

Sediment movement

The removal of sediment from, or its addition to, a slope, is not usually the result of a single process but the interaction of a number of different ones. For the purposes of this study, however, the removal of sediment in slopewash was the only form of transport that was intensively investigated.

Nevertheless, it was probable that other processes, particularly aeolian activity, effected considerable amounts of sediment transport.

(a) Suspended sediment

Overland flow was sampled 66 times for suspended sediment at the three runoff plots: sediment amounts were expressed as concentrations in terms of parts per million of water (p.p.m.). No obvious temporal trends were observed in the concentrations, indicating that source factors rather than discharge were the key variables. It would have been possible to split up the discharge records into segments corresponding to the individual sediment samples in order to obtain total sediment removal. This was not undertaken, however, since it was felt that such a method would appear to give the results a degree of accuracy that was unwarranted. As an alternative, the mean concentration for each plot was multiplied by the total flow volume originating on it (Table 16). The volumes of overland flow included both real and simulated values where necessary. It was not felt valid to undertake the calculations for plot 3, first because regression against lagged net radiation did not produce accurate estimates of overland flow, and second, because the vast majority of the snowbank had ablated before records began.

The mean concentration of suspended sediment was greatest at plot 1, but the denudation estimate for that site was much lower than at plot 2 due to the smaller flow volume. The higher concentration probably reflected the sparser vegetation cover and more tractable particles: grains were observed moving

Site	Time period	Number of samples	Mean concentration (p.p.m.)	Standard deviation	Total flow volume ^a (litres)	Denudation estimate ^b ($\text{g m}^{-2} \text{ yr}^{-1}$)
Plot 1	May 28 - June 5	17	130	15.4	964	0.92
Plot 2	June 5 - June 19	36	110	6.1	35,400	10.04
Plot 3	June 20- June 27	13	111	11.5	-	n.d.
a Includes simulated values where actual data are missing. b Denudation estimate = (Total flow volume x Mean concentration) / (Plot area) .						

TABLE 16. - Suspended sediment concentrations in overland flow.

as bedload and forming a miniature delta on plot 1. The overland flow collector had trapped 13 gm of sediment by the end of the observation period and this amount was added to the value obtained by multiplying concentrations by flow volumes.

The sediment concentrations obtained from plot 2 showed a low degree of variation, the maximum being 129 p.p.m. and the minimum 100 p.p.m. It was hypothesised that this uniformity of concentration was attributable to the ponding of water on the plot in between cross-slope bands of vegetation. This was thought to prevent the movement of any larger particles as bedload, and concentrations were based therefore on the finer sediment fraction which presumably was relatively constant.

Throughflow was sampled 27 times for suspended sediment at sites 1 and 3, and concentrations averaged 250 p.p.m.. Considering the much smaller amounts of throughflow compared with overland flow, concentrations even of this magnitude were relatively insignificant. Additionally, it was felt that the concentrations may have been inflated by ground disturbance associated with the installation of the throughflow collectors. No detailed analysis was undertaken, but it seemed probable that the process of thermo-planation was unimportant or non-existent at the study sites.

(b) Aeolian activity

Two samples of snow were analysed for sediment content at plot 1, giving an average of 9.46 gm m^{-2} of plot surface. If only 0.92 gm m^{-2} were removed by the slopewash processes

measured (Table 16), there was the possibility that deposition was actually the dominant process. More likely, however, the particles deposited from the snow on the plot surface were later eroded by aeolian activity. The overall denudation balance on the slope, therefore, was impossible to estimate.

At plot 2, the snow surface was dirty and 1 m^2 was cleared and filtered. Here, there was a concentration of 17.35 gm m^{-2} compared with the slopewash denudation estimate of 10.04 gm m^{-2} . Again it was not possible to estimate the annual slope denudation balance as aeolian activity was not confined to deposition, nor to the study period.

No slopewash denudation estimate was made for plot 3, but qualitative observations suggested that the ablating snow-bank exhibited a number of standard features over its length of approximately 200 m (Figures 33 and 34). At the top and bottom of the bank, zones of white broken ice existed, consisting either of discrete pieces, or a cracked, cloudy ice mass (thought to be the 'feather edge' of Washburn (1973, p. 204)). Below the layer of broken ice at the top of the bank, and above the broken ice at the bottom, were zones of 'solid' ice, grey in colour and relatively clear and smooth; the solid ice zones appeared to be continuous at depth under the rest of the snow-bank. Between the solid ice zones was 'clean' or 'dirty' coarse granular snow at the surface. The latter was at the upper interface between solid ice and snow and continued at depth. The former made up the remainder of the snowbank. At the upper interface between the solid ice and the snow, sedimentary

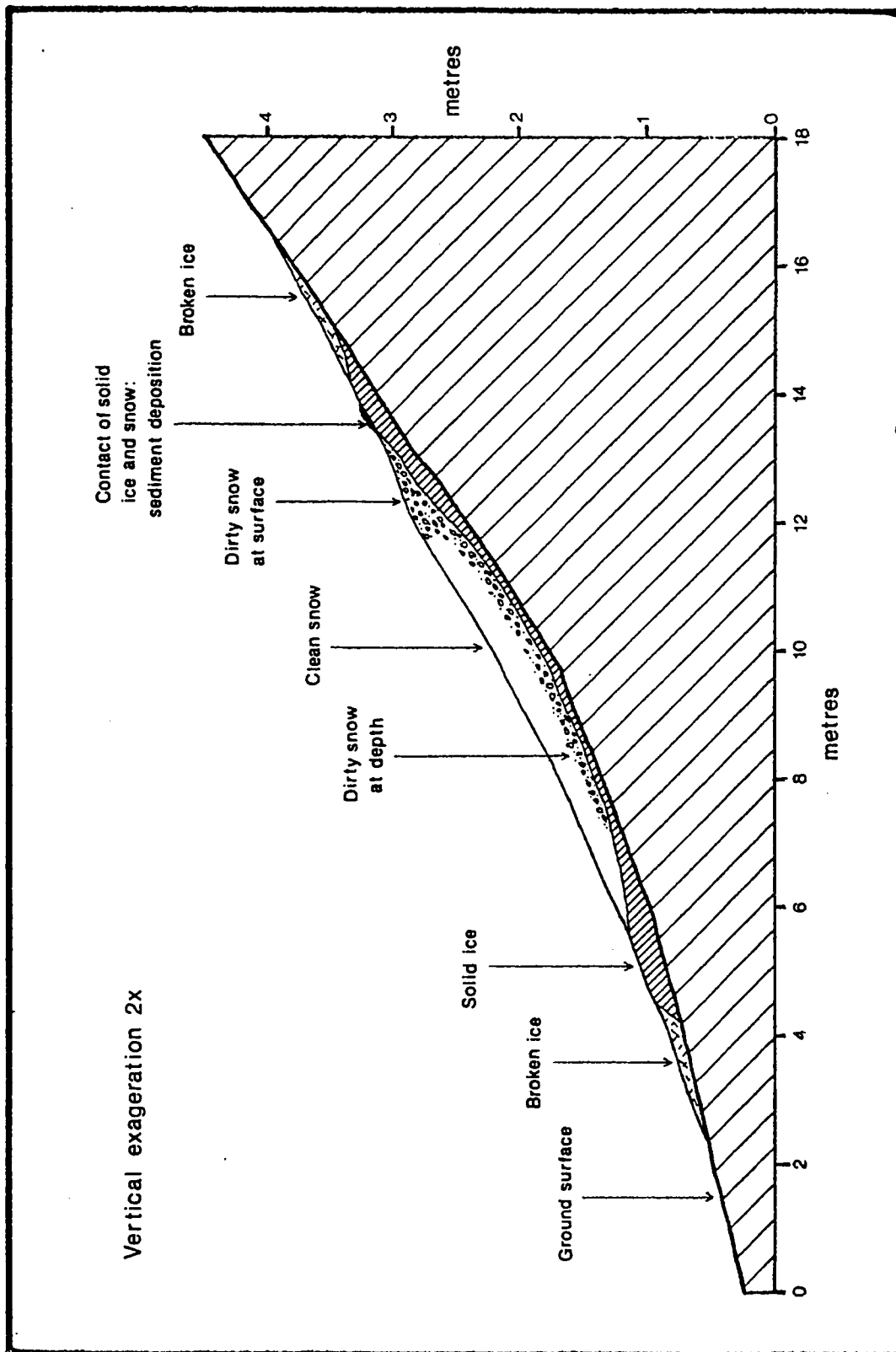


FIGURE 33: - Plot 3: cross-section of ablating snowbank (June 23).

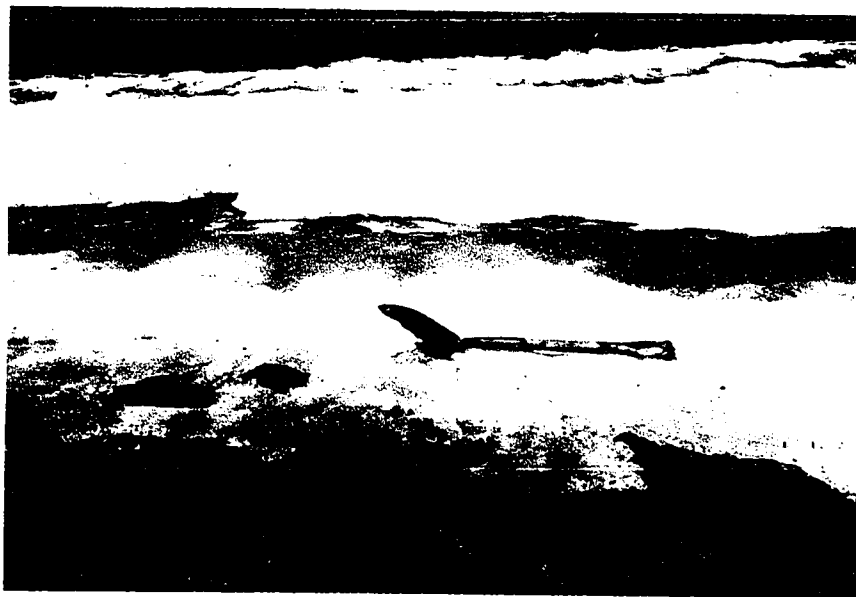


FIGURE 34: - Ablating snowbank, site 3 (June 21).

Note: view is upslope.



FIGURE 35: - Sedimentary deposits on snow, site 3 (June 22).

deposits were observed (Figure 35): these were being eroded by water flowing over the ice surface, and being redeposited inside the snow zone, presumably when the speed of flow was reduced due to impedance by the snow grains.

As the snowbank at site 3 ablated over a number of days, the area of clean snow contracted and the line of sediment deposits moved downslope in unison with the upper edge of the bank. Ultimately, when all the snow had disappeared and only ice was left, most of the sediment was carried down off the ice and deposited as patches on the slope (Figure 36) although some remained on the ice and was dropped in situ. The actual position of deposition was considerably affected by weather conditions on the final day of melt: on days with high insolation rates and/or high air temperatures, rapid melt rates were induced and the sediments were carried further downslope than they were on cool, cloudy days.

The sediments consisted of fine sand and were laid down as discontinuous patches up to 5 cm thick. Deposition occurred both on top of small hummocks and in the gaps between them. A total of 1.5 kg of sediment (equivalent to 1.07 kg m^{-1} slope width) was collected by carefully scraping the surface of the runoff plot. This figure is considered accurate to $\pm 20\%$. Assuming that the plot site was representative, 0.21 metric tons ($\pm 20\%$) were deposited along the base of the 200 m wide slope (Figure 37).

The mechanism of downslope movement of the sediment on the snowbank surface is relatively clear, but the origin of the



FIGURE 36: - Sedimentary deposits on the slope, site 3 (June 24).
Note: lens cap is 55 mm in diameter.



FIGURE 37: - Line of sedimentary deposits, site 3 (July 3).
Note: deposits are light-coloured and run in an irregular line across photograph.

sediment is more problematic. It is thought that the immediate source was the dirty basal snow (Figure 33) which possessed a sediment content of 2800 p.p.m.. As the snowbank ablated, this sediment-rich snow was progressively exposed at the surface and further sediment concentration occurred. The basal sediment could have been derived from the slope itself or from aeolian sources. Concerning the former, the only position where the ground appeared erodable upslope of the deposits was in the broken ice zone at the top of the snowbank. There the ground surface was thawed and unprotected by the solid ice layer. However, there appeared to be no mechanism for particles to be transported from the broken ice region, over the surface of the solid ice and into the basal snow band. Therefore it seemed impossible for the sediment to have been derived from the slope itself, and it was concluded that the sediment was aeolian in origin.

The aeolian activity may have occurred in winter or summer. Winter activity normally produces a series of sediment stratifications in the snow and these were not observed at site 3. The dirty basal snow itself was 15-20 cm in thickness and unstratified. By the time the snow at site 3 was examined, however, sediment banding probably would have been destroyed by water movement through the snow so that its absence was inconclusive. Further observations, however, were available from sites 1 and 2, and banding was not seen at either site. Therefore, if winter aeolian activity was the cause of the sediment it must have produced low concentrations distributed

throughout the snow. The ablation that took place would have been responsible for a redistribution of the sediment resulting in the dirty basal snow.

Summer aeolian activity certainly occurs in the Thomsen Fly Camp area and can produce considerable sediment concentrations on snow surfaces. At site 3, however, the cleanliness of much of the snow surface (Figure 33) excludes spatially homogeneous sediment deposition. Nevertheless it is possible that summer sediment deposition did take place and that the majority occurred on the upper part of the snowbank (already ablated when observations commenced). This distribution could have been caused by a decrease in wind-speeds immediately below the crest of the slope, since it is in the lee of winds carrying particles from the sand plain. The sediment deposited could then have been transported downslope and concentrated in the basal snow by melt-water.

On the above evidence, neither a winter nor a summer origin for the sediment can be exclusively accepted or rejected, and further investigation is required. This is particularly important since the amount of sediment is of a magnitude sufficient to be significant to the development of the slope.

(c) Dissolved sediment

Conductivity measurements were made for 60 of the 66 overland flow samples at the three plots (Table 17), and the values were converted into p.p.m. of calcium carbonate (CaCO_3). No attempt was made to analyse variations in the within-plot

Site	Number of observations	Mean conductivity (micro-mhos cm^{-1})	Standard deviation	Mean concentration ^a of dissolved solids (p.p.m. CaCO_3)	Denudation estimate ^b ($\text{g m}^{-2} \text{yr}^{-1}$)
Plot 1	16	296.8	86.4	127.8	0.82
Plot 2	33	64.3	30.0	26.1	2.38
Plot 3	11	145.0	88.2	58.9	n.d.
^a Derived from instrument manufacturer's values. ^b Denudation estimate = (Total flow x Mean dissolved solids)/(Plot area).					

TABLE 17. - Dissolved solid concentrations in overland flow.

values of dissolved solids: instead, average values were calculated.

As with suspended sediment, values of dissolved solids were greater at plot 1 than at plot 2. Due to the amount of flow at the latter, however, the denudation estimate is considerably higher. The concentrations for plots 2 and 3 are within the range (18-95 p.p.m.) recorded by Smith (1972) from snowbanks on Somerset Island, N.W.T., but the values at plot 1 are considerably in excess. Reasons for this are not known.

The mean conductivity of throughflow at site 1 was 296.0 micro-mhos cm^{-1} (10 samples), virtually the same as that of overland flow (Table 17). Since approximately half of the runoff at site 1 had occurred on the surface and the other half had passed through the active layer, denudation by solution was accomplished equally by overland and throughflow. Further, the amounts of sediment removed by overland flow in suspension and solution were almost equal: $0.92 \text{ g m}^{-2} \text{ yr}^{-1}$ (Table 16) and $0.82 \text{ g m}^{-2} \text{ yr}^{-1}$ (Table 17) respectively.

At site 3, the average conductivity (12 samples) for all depths was 432.9 micro-mhos cm^{-1} (186 p.p.m. CaCO_3), three times as much as the value obtained from overland flow. This high concentration was influenced by figures from the deepest gutter which were twice those of the upper levels (maximum of 365 p.p.m. CaCO_3). In terms of total denudation, however, solution was insignificant in the lower levels due to the small amounts of throughflow, and could have been important only in gutter 1

where the possibility existed that high flow rates had occurred for some weeks prior to the commencement of the data record.

The differences in throughflow solute concentrations between sites 1 and 3 and the intra-site variations with depth may have been caused by (1) the growth of different types and amounts of vegetation, (2) the varying times that it took for throughflow to move through the regolith, and (3) differences in the mineral composition of the substratum. More data are needed before any conclusions can be reached regarding these possible influences. However, the results were similar to those obtained from 'soil water' by Woo and Marsh (1977) on Ellesmere Island, and greater than those obtained by Cogley (1972) on Devon Island or Smith (1972) on Somerset Island.

Solute concentrations in overland flow at plots 2 and 3 were low in comparison with those obtained in humid temperate and tropical regions, where total hardness values are commonly 100-300 p.p.m. (Smith and Atkinson, 1976). Only the values for overland flow at plot 1, and the concentrations in throughflow at both sites 1 and 3, were within this range.

(e) Summary

The quantities of data collected on the transportation of sediment by slopewash were insufficient to reach any detailed conclusions concerning slope denudation. However, three main points can be made. First, overland flow was more important than subsurface wash in locations where large snowbanks existed, simply due to the much greater amounts of water leaving the slope by this means. Second, with regard to

overland flow, suspended sediment concentrations did not vary significantly amongst the sites (Table 16). On the other hand, flow volumes did vary considerably, so that in terms of total denudation, the latter is the key variable. Third, as a consequence of the two other points, attention must be directed towards the factors concerned with the formation of snowbanks (for example aspect, prevailing winds, and breaks of slope).

CHAPTER IV

CONCLUSIONS

This study attempted to investigate the temporal and spatial variability of slopewash in a permafrost environment. The results obtained, however, become more meaningful if two types of comparisons are made: first, with the mechanism and importance of slopewash processes in other environments, and second, with other processes operating in permafrost regions. Generalised conclusions only may be reached, since it is wrong to assume that the Thomsen Fly Camp area is representative of all permafrost environments, and that the summer climate of 1977 was necessarily typical.

Slopewash processes

The existence and type of slopewash processes are closely integrated with models of runoff development. Horton's (1945) model relating overland flow and consequent stream hydrograph rise to the exceedance of infiltration capacity by rainfall intensity, has been shown to be invalid in temperate climates (for example, Kirkby and Chorley, 1967). It was hypothesised, however, that the model might remain correct in hot arid and semi-arid environments. Recent work has tended to disprove this suggestion (De Ploey et al., 1976; Yair and Klein, 1973; Yair and Lavee, 1974; 1976) and even in these regions,

surface runoff only develops on part of the slope. This also proved to be the case for slopes examined in the present study.

A previous study in the Canadian High Arctic (Woo, 1976b) revealed that during the snowmelt period, the main cause of variability in runoff-producing areas is the availability of meltwater. Further, it was concluded that after snowmelt, a catchment in the discontinuous permafrost zone of Alaska exhibits partial area contribution to runoff (Dingman, 1971). The present research provides further support for these observations.

A major conclusion of the research is that Hortonian overland flow is unlikely to occur in the High Arctic due to low precipitation intensities. Post-melt surface runoff is rare and dependent on antecedent conditions. The source areas are confined to locations where snowbanks have recently ablated, rather than to the channel-bank areas per se. Similarities in surface runoff production therefore exist between permafrost and non-permafrost environments, in that source areas are discontinuous and variable in extent, dependent on antecedent conditions.

Two principal differences are apparent, however, between slopewash processes in permafrost and non-permafrost environments. The first is related to the water source. In permafrost environments, free water in a liquid form can be present at the surface for only 3-4 months of the year. The majority of this water is winter precipitation that has been stored as snow and then released during the melt period. The volume of slopewash

produced by summer precipitation appears to be insignificant in comparison with that derived from snowmelt. In contrast, in most other environments, slopewash is not confined to one period of time. Further, due to the geomorphic effects of rainsplash (De Ploey and Savat, 1968; Ellison, 1944), the fact that most precipitation in non-permafrost environments is in the form of rain, is of itself significant.

Rainsplash processes were not quantified and are thought to be relatively unimportant in the study region due to the small amounts and low intensities of rainfall. It may be possible, however, for a considerable amount of erosion to be accomplished by a storm with a large recurrence interval, as much of the ground surface possesses a low percentage vegetation cover and is therefore poorly protected.

The movement of sediment downslope by surface flow may be weathering- or transport-limited (Carson and Kirkby, 1972, p. 190; Young, 1972, p. 66). A consequence of the lack of rainsplash in the Thomsen Fly Camp area is that the sediment concentrations in overland flow were probably weathering-limited: increases in measured concentrations were not related with hydrograph peaks and absolute values were low (maximum of 175 p.p.m.). This was in contrast with hot semi-arid regions where flows are often transport-limited and sediment concentrations can become very high (for example, 10,425 p.p.m. (Yair, 1972)).

Young (1974, p. 69) provides a list of surface wash rates, with values of ground lowering given in Bubnoff (B) units

(1 unit = 1 mm/1000 years). On this scale, those rates observed by Jahn (1961) are among the lowest on the table. No such erosion rate conversion was made for this study for two reasons. First, the results of one season were not adequate to allow their extrapolation over 1000 years. Second, due to the influence of aeolian transport, it was not clear whether the average removal of sediment per square metre of plot actually represented a surface lowering. In spite of these limitations, it is suggested that the presented data are in broad agreement with those of Jahn and indicate a low position on the table.

The second principal difference in the way slopewash processes act in permafrost and non-permafrost environments is related to the presence of frozen ground. Although its effects are most significant at the beginning of thaw, they remain important for the whole season on slopes which are well-vegetated. The frost-table acts as an impermeable barrier, preventing the percolation of water and causing lateral movement instead. In this way it is possible for slopes to become saturated during the snowmelt period and to produce surface runoff. Further, slopes below ablating snowbanks are often well-vegetated and may remain saturated for several weeks, since the layer of thawed ground thickens slowly.

The presence of permafrost per se is important only at the end of summer when the frost-table coincides with the permafrost table: the system then acts much the same as an impermeable bedrock covered by a shallow regolith. This analogy cannot be made for the beginning of the season when the rapid

daily thickening of the thawed layer critically affects the relative quantities of surface and subsurface runoff.

Overall, the effect of frozen ground is to increase the possibility of surface runoff, thereby reducing the importance of subsurface water movement and eliminating the movement of water at depths below 1.0-1.5 m.

Slopewash and other processes

The importance of slopewash in relation to other processes in permafrost regions is difficult to assess since few quantitative studies have been undertaken. That by Rapp (1960) of the Karkevagge Valley in northern Lapland is one of the most comprehensive process studies available. Unfortunately, Rapp did not investigate slopewash so that figures directly comparable with the present study do not exist.

Rapp found that the most important transport process was dissolved salts, in which form $26 \text{ tons km}^{-2} \text{ year}^{-1}$ ($26 \text{ g m}^{-2} \text{ year}^{-1}$) of material were removed. This was almost equal to all the other quantified processes put together. The figure is an order of magnitude larger than that obtained at Thomsen Fly Camp. However, there was a moderate precipitation at the Karkevagge, and evidence of considerable chemical weathering (for example, weathered rocks, avoidance of acidic areas by plants) and this did not appear to be the case for the Thomsen Fly Camp region.

Rapp found the rapid transfer of material (for example, earth-slides, avalanches and rockfalls) to be the next most

important transportational process. Similar processes are probably of little importance in the Thomsen Fly Camp area since there are very few steep slopes and active layer failures and rapid slope movements are not apparent.

Surprisingly, Rapp found solifluction and talus creep to be of minor significance. Although no measurements are available for the study area, data have been obtained for low-angled slopes ($2-4^{\circ}$) at Sachs Harbour (French, 1974b; 1976a). The data show that subsurface downslope movement takes place to a depth of at least 30 cm and averages $1.0-1.5 \text{ cm year}^{-1}$. This gives a volumetric downslope movement of between $22-30 \text{ cm}^3 \text{ cm}^{-1} \text{ year}^{-1}$. It seems likely that this magnitude of mass movement occurs in the study area since non-sorted stripes and solifluction lobes are widespread.

Although the slopewash data collected in this study and the solifluction data from Sachs Harbour are not directly comparable, it is suggested that solifluction is a more important transport process than slopewash. Slopewash appears to be effective only in limited areas, the snowbank locations, and at the surface only, while solifluction occurs on virtually all slopes and down to depths of 30 cm or more. Hence it seems probable that solifluction produces a higher transport rate in terms of $\text{cm}^3 \text{ cm}^{-1} \text{ year}^{-1}$.

Suggestions for future research

Many problems were encountered in this study and although field solutions were available for some, others required further thought or additional equipment.

The greatest weakness of the study was its short duration: the results remain questionable until they are known to be typical for the location. This problem can be solved only through the creation of a longer-term data base. A second general uncertainty, relates to the spatial representativeness of results. Further studies in permafrost areas with climatic characteristics both similar and dissimilar in terms of precipitation, heat fluxes and active layer thaw are required to assess this factor.

One of the ways of improving future slopewash studies is through an increase in data accuracy. In some cases (for example, evapotranspiration), this may require the use of different data collection techniques, while in others (for example, evaposublimation), extra care in certain parts of the technique already used is all that is necessary.

The most important technical improvement that should be made, relates to the measurement of surface runoff. Data collected concurrently at a number of sites are needed, and this goal requires that automatic measurement devices (for example, weirs and water-level recorders) are employed. In addition to rendering unnecessary the simulation of gaps in the data, automatic measurement would allow a much better evaluation of the effects of site on the relative proportions of snowmelt, surface and subsurface runoff.

The study has revealed a number of other factors that should be monitored in the future. Data on aeolian inputs and outputs to the slopewash system are needed to allow evaluation

of the slope denudation balance. Information on turbulent and latent heat exchanges to the snowbank is necessary to make better predictions of snowmelt, and hence surface and subsurface runoff. In relation to the latter, coefficients of permeability or values of hydraulic gradients could aid the understanding of subsurface hydrograph variations. Finally, the hypothesis that rainsplash is unimportant in permafrost regions, requires testing.

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ABSTRACT

The magnitude and frequency of slopewash processes in a permafrost environment were examined during the summer of 1977. Small instrumented plots were set up and monitored for surface runoff at one interfluvial and two valley-side locations. Subsurface water was collected at two of the sites using guttering positioned at various depths in the active layer. Snowmelt was the major source of both surface and subsurface runoff, and summer precipitation was of little importance. Volumes of surface runoff were generally greater than those of subsurface flow, although both varied with position on the slope. Partial area contribution to streamflow was usual, the source areas being those conducive to snowbank formation. Measurement of the amounts of suspended and dissolved sediment indicated that surface wash was a more important transport agent than subsurface wash. Both processes, however, appeared to be of limited importance at the interfluvial location, and only transported significant amounts of sediment at the valley-side sites where large snowbanks ablated.

RESUME

Au cours de l'été de 1977, l'importance et la fréquence du ruissellement ont été étudiées dans un milieu de pergélisol.

Pour cela nous avons établi des parcelles de mesure en trois sites: un sur un interfluve et les deux autres sur deux flancs des vallées. Les eaux souterraines ont été recueillies grâce à des pertuis disposés à différentes profondeurs dans le mollisol. Les résultats montrent que l'eau de fonte de neige est la composante principale de l'écoulement aussi bien en surface que dans le sol, alors que les précipitations estivales ne jouent qu'un rôle mineur. En volume, l'écoulement superficiel était en général plus important que l'écoulement souterrain, bien que nous ayons observé des variations selon la localisation sur la pente. L'alimentation de l'écoulement concentré provient presque exclusivement de certaines parties bien définies des versants, plus précisément celles où se forment les bancs de neige.

Les mesures des quantités de sédiments en suspension ou dissous montrent que le ruissellement est un agent de transport plus important que l'écoulement souterrain. Il apparaît toutefois, que l'action des deux processus est limitée sur l'interfluve et ne joue un rôle significatif dans le transport des sédiments que dans les sites de vallée, là où existaient de grands bancs de neige.