CHANNEL DEVELOPMENT AND FLUVIAL PROCESSES IN SNOW-FILLED VALLEYS, RESOLUTE BAY, N.W.T.

CHANNEL DEVELOPMENT AND FLUVIAL PROCESSES IN SNOW-FILLED VALLEYS, RESOLUTE BAY, N.W.T.

by

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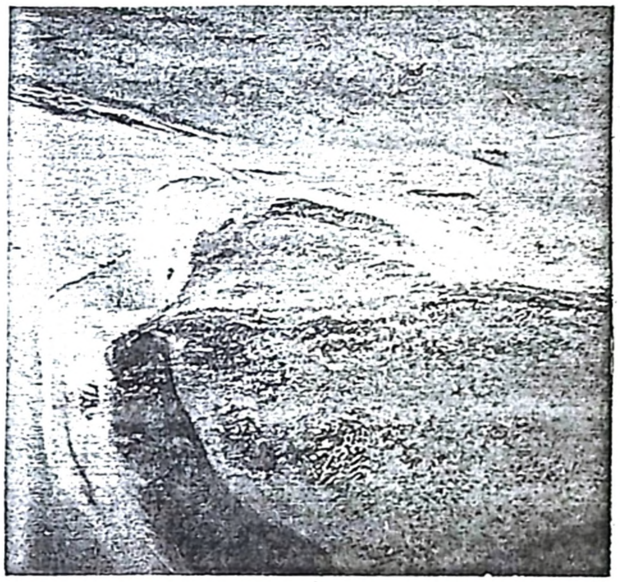
A Thesis
Submitted to the Faculty of Graduate Studies
in Partial Fulfilment of the Requirements
for the Degree
Master of Science

McMaster University
1978

A mes père et mère,

Avec toute ma gratitude.



Frontispiece: Aerial view of differential snow deposition in an incised valley; the photograph was taken in mid-May 1977.

MASTER OF SCIENCE (1978)
(GEOGRAPHY)

McMASTER UNIVERSITY
Hamilton, Ontario

TITLE: Channel Development and Fluvial Processes in snow-filled valleys, Resolute Bay, N.W.T.

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NUMBER OF PAGES: xiv; 132

ABSTRACT

In 1977, this study was carried out in a small drainage basin ($33\ km^2$) near Resolute (74°55'N, 94°50'W), Northwest Territories (1) to examine the manner in which meltwater runoff carves channels in the valley snowpack before the channels become stablised on their clastic beds, and (2) to assess the role played by valley snowpacks on fluvial processes.

Major factors controlling channel development in the snowpack include the distribution and the characteristics of the snow, which in turn are related to the local topography and the prevailing directions of winter snowdrift. Based on this relationship, an attempt was made to predict the sequences of channel development in terms of several processes including ponding, tunnelling, lateral and vertical shifting, and stream capturing.

Availability of water controls the rate of channel development sequences and hence the magnitude of fluvial processes over a flow season. In the case of substantial runoff, the rate of snowpack depletion is rapid. However, since the bulk of annual water discharge occurs while the snow is interposed between the running water and the bed material, little geomorphic work is performed during the early part of the flow season. For four selected sites, calculations suggest a protective effect of the snow in reducing the potential bed material transport.

ACKNOWLEDGEMENTS

I wish to thank my supervisor, Dr. M.K. Woo for his assistance and encouragement by his stimulating advices and guidance during my work at McMaster University. His availability was really appreciated.

The National Research Conncil of Canada, the Water Resources Research Grant from Inland Waters Directorate, Environment Canada, and a grant from the Presidential Committee on Northern Studies of McMaster University provided financial support for this study. The generous logistical support of Polar Continental Shelf Project, Department of Energy Mines and Resources is gratefully acknowledged.

I am greatly indebted to all my friends in Hamilton who permitted this stay to be so enjoyable, particularly to the presence of Cheryl. Special thanks to Barbara for the typing of the first draft of this paper, and to Rosanna for its final version.

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CHAPTER ONE

INTRODUCTION

1.1 Introduction

In the High Arctic, many small valleys and gullies are partially or totally filled by snowdrifts during the long polar winters. With the arrival of spring, annual peak flow occurs as the basin snowpack is rapidly melted. In the snow-filled valleys, however, "snow and ice constitute, for a long time, a screen between the flow of water, the streambed, and the banks" (Pissart 1967). The presence of snow and ice can therefore inhibit geomorphological activities because the energy of the water will be mostly spent in the removal and the melting of bottom ice and snow in the valley.

Several questions then arise, including:

(1) How does the protective effect of snow and ice affect the geomorphological action of the stream?
(2) How is the hydraulics of flow affected by the presence of snow and ice in the channels?
(3) How does the stream overcome its snow-filled condition?

This research will attempt to answer the above questions.

1.2 Literature Review

Various researchers in the High Arctic have observed the openings of stream flow in snow-filled valleys.

Cook (1967) identified five stages in the annual stream flow cycle of Mecham river, Cornwallis Island. Firstly, a premelt period in early June is characterised by sublimation. It is followed by a second period of intense melting where daily temperatures are above the freezing point. In the third stage, meltwater is collected in the valley floor and ponds up behind ice-jams. The break-up of the ponds produces catastrophic flood events which contribute a large proportion of the total annual runoff. The fourth period is of sharply decreasing flow until the river is frozen to the bottom, followed by an extensive period of snow accumulation, which constitutes the fifth period. Church (1972) observed in more detail the break up stage in Baffin Island and emphasised the role of "extended fair weather or, more efficiently, a rainstorm" as a triggering mechanism leading to the flushing of the snow in channels and hence initiating significant runoff. Similarly, Washburn (1973) considered slush flow as a common method of stream break up in Northeastern Greenland.

Although Pissart (1967) recognised the effect of the bursting of ponded waters as a triggering mechanism for the break up of some stream segments in snow-filled valleys, he described a more gradual opening of the channel following the snow saturation period. Sturges (1975) also reported a non-catastrophic opening of the stream occurring as "oversnow flow" in the presence of an intermittent stream channel in a middle latitude environment. He discussed the positive effect of differential thickness of depth hoar at the bottom of the snowpack and the influence of a low melt rate at the start of snowmelt in the development of subsnow-surface conduits.

Moreover, Pissart emphasised the protective role of the snow and bottom-ice. He reported that for a ten-day period, 75 percent of the total annual discharge merely flowed over an ice layer without incising it, thus "reducing the activities of running water". Inversely, after recognising the importance of the "ripening" of the snowpack which has to attain "its capacity level of free water storage" before the commencement of flow, McCann et al. (1972, p. 78) pointed out that "stream channels on the snow surface are soon abandoned for true alluvial channels". Although this is probably true for larger streams, Sturges (1975) presented a comprehensive study of the observed effect of a snow-filled valleys on the fluvial processes of a small stream in Wyoming, U.S.A. He reported that with a late and sudden start to the ablation season, the stream will flow oversnow "when the quantity of meltwater reaching the stream channel exceeds the rate at which water is transmitted through the dense snow filling the channel" (Sturges 1975). The oversnow flow tends to concentrate at the interface between the snowpack and the side of the channel which results in significant bank erosion. However, only a small proportion of the sediment attains the watershed outlet. This is shown by the great volume of sediment deposited on the snow surface which is caused by the filtration effect of the pack; this filtration is a function of the physical straining of sediments when water percolates through the snow.

Similarly, in high latitudes, though a certain geomorphological inefficiency is implied by partial flooding over the snow, an extensive movement of bed load, taking its source from unprotected reaches and banks, occurs during the period of high water velocities. The magnitude of this load component is shown by "the cast of a former channel course on a snow surface" which the channel had occupied "for only two days, before it was abandoned in favour of a course beneath the snow" (McCann and Cogley 1973, p. 123).

Fluvial processes in the Polar Regions are responsible for denudation and transportation activity, and, in the last decade, underestimation of these activities has diminished. The work of authorities such as St-Onge (1965) and Church (1972) and physical evidence of reasonably well organised drainage networks to substantiate their observations have removed any doubts as to the importance of fluvial processes. The study of valley morphology discussed by St-Onge (1965), Pissart (1967), and McCann and Cogley (1973) shows that the "removal of material from the landscape is ultimately effected at high latitudes by the same fluvial agents as the mid-latitude regions" (Cogley 1972). Despite low annual precipitation, a high runoff ratio will occur because of the concentration of the bulk of the flow in a short period, the presence of a frozen substratum, and the absence of a substratum, and the absence of a vegetative cover.

To date, however, there is no detailed work on Arctic channel development in snow-filled valleys. Moreover, despite the recognition of the importance of Arctic fluvial processes as geomorphic agents, no quantitative assessment of the protective effect of snow and bottom-ice in streams exists in the literature on the High Arctic.

This study of channel development and fluvial processes in snow-filled valleys will extend our knowledge of the differential openings of streams, relative to their geomorphological action, and hence offer an improvement of water resources management in the Canadian High Arctic.

1.3 Objectives

To improve understanding of the role of snow and ice in fluvial processes, the research has the following objectives:

(1) to determine the temporal and spatial sequences of channel development in snow-filled valleys

(2) to study the effect of snow and ice on flow conditions

(3) to assess the importance of the protective effect of snow and ice on the geomorphological action of High Arctic streams.

CHAPTER TWO

STUDY AREA

2.1 Location

Field study was carried out in a drainage basin at Resolute, Cornwallis Island, Northwest Territories, Canada (74°55'N, 94°50'W), 5 km northeast of Resolute Airport (Fig. 2.1).

The basin area was chosen because:

(1) a great variety of snow accumulation patterns was encountered in its valleys.

(2) of its proximity to Resolute, allowing easy access.

2.2 Climate

In general, the hydrological behaviour of a basin is strongly dependent on the climate. In the High Arctic, precipitation and temperatures are low, the latter allowing a ten month basin storage in the form of ground ice and snow accumulation. Despite their desert characteristics, these cold regions experience a significant snow-melt flood of short duration during the beginning of the flow season, (i.e. when daily temperatures permit a sudden release of the water storage).

From this climatological characteristic, the great significance of streams and rivers as geomorphological agents in the periglacial

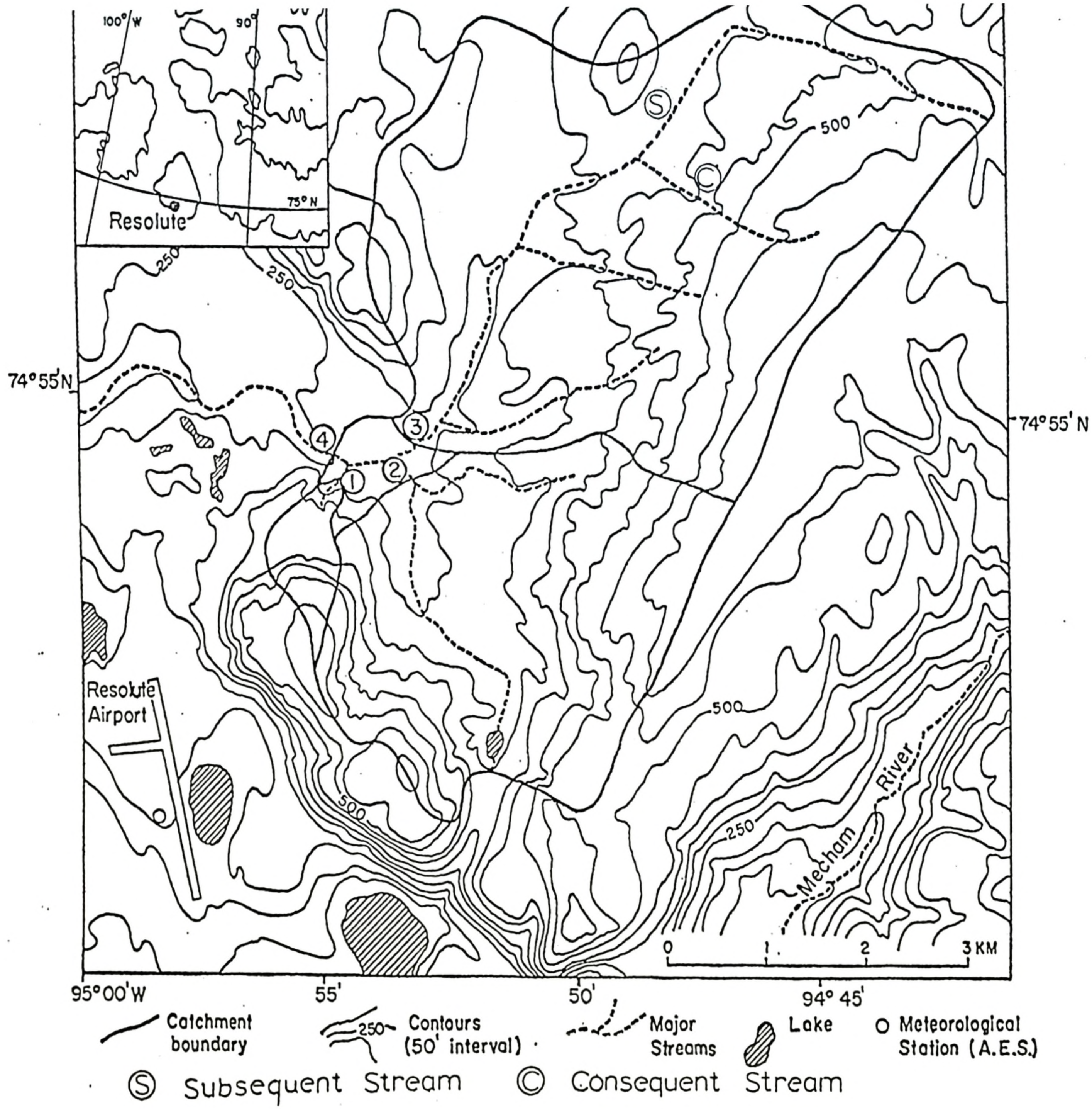


Fig.2.1. Topography of the study area. Numbers in circles indicate the outlets of Basin 1 (area 0.5 km^2), Basin 2 (area 10. km^2), Basin 3 (area 21 km^2) and Basin 4 (area 33 km^2).

environment was recognised(St-Onge 1965, Cogley 1972, McCann et al. 1972, Church 1972).

A comprehensive summary of climatic conditions for Resolute is found in Marsh (1978). However, out of all the climatic processes, emphasis should be placed on snow drifting which plays a major role during the winter months in the redistribution of the snow cover (Cook 1967, Sturges 1975).

2.3 Snow Distribution and Snow Characteristics

Snow and ice form an integral part of the landscape and they persist for nine or more months each year. During the long Arctic winters, snow is redistributed by drifting and the subsequent pattern of snow disposition is therefore strongly influenced by an interaction of wind direction, wind intensity and the basin terrain. The snowpack will be disposed in a different manner, i.e. accumulated in sheltered area (e.g. valleys), leaving hill tops and flat terrain snow-free. This disposition and peculiar characteristics of the snow, such as higher density and hardness acquired during the snow drifting process, will constitute the initial conditions of channel development in snow-filled valleys.

By the end of summer, most High Arctic streams experience low flow conditions. When freeze-back occurs, the channels are practically dry. Other than a sporadic presence of thin ice patches formed in the pools of the streambed, winter snowpack develops directly on the dry valley floor.

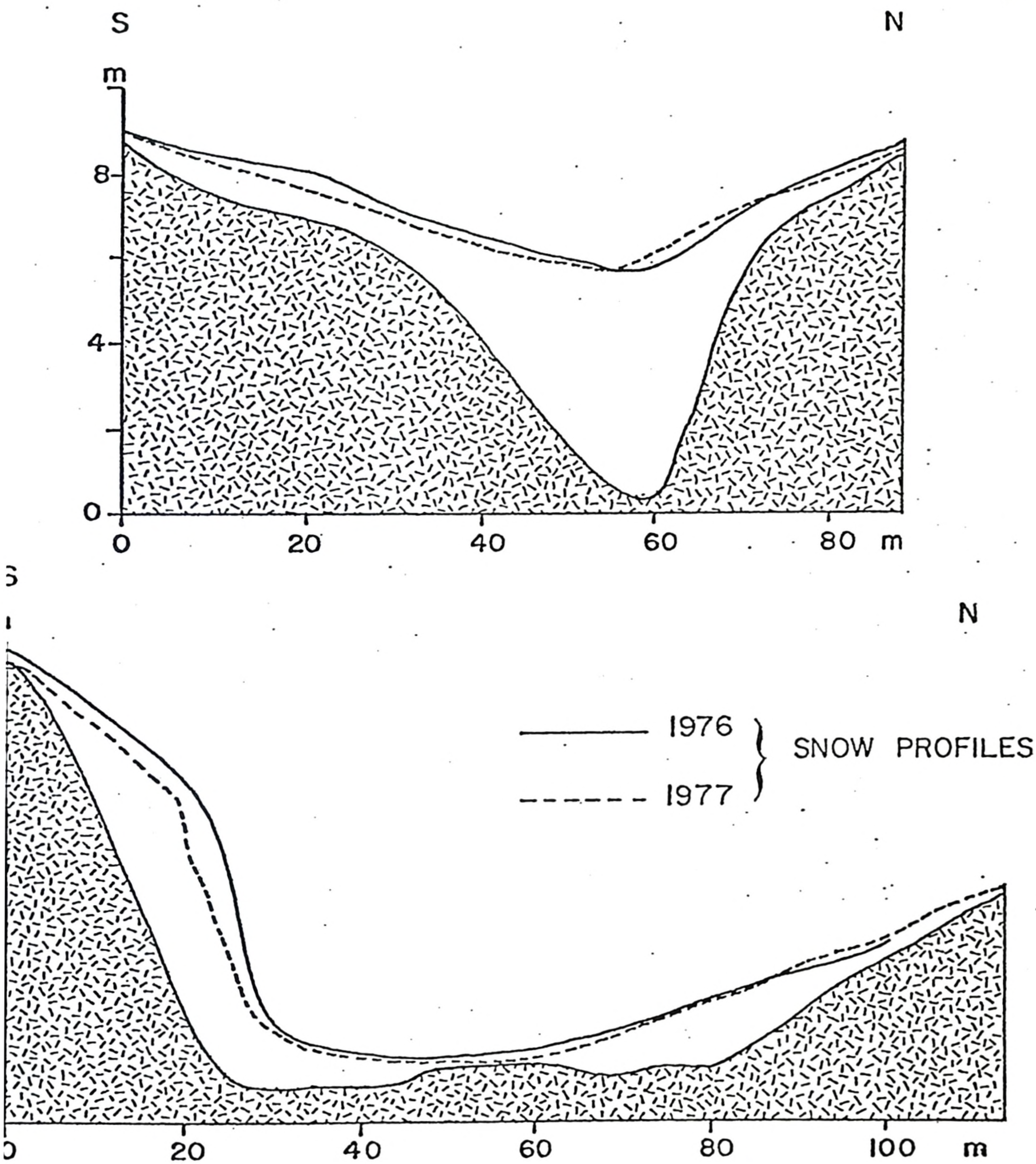


Fig.2.2. Snow surface profiles across two typical valleys. Note similarity of the profiles surveyed in May 1976 and in May 1977 although the latter winter deposited considerably less snow over the entire drainage basin.

The distribution of an Arctic basin snow cover is characteristically uneven (Woo and Marsh 1978). Larger valleys accumulate a substantial quantity of snow and depending on the prevailing wind and the local terrain, snow distribution across a valley is seldom symmetrical, nor does the snow profile resemble the cross-profile of the valley floor (Fig. 2.2). Along the longitudinal axis of a valley, the snow surface is uneven, with deflation hollows scooped out between snow ridges which traverse the valley (Fig. 2.3).

When the prevailing winter wind direction does not change from one year to another, similar drift patterns develop. At two sites in the basin, snow profiles were repeatedly surveyed in mid-May of 1976 and of 1977 (Fig. 2.2). Although the winter of 1976-77 received only two-thirds of the amount recorded in the previous winter (Woo and Marsh 1978), the snow accumulation pattern of both years remained similar at the two sites. This suggests that in terms of the valley snowpack, the process of snow drifting rather than the absolute amount of snowfall plays the dominant role in affecting snow distribution.

Snow drifting also results in snow compaction. In the High Arctic where the wind is not obstructed by vegetation, the snow of the pack tends to be denser and harder than the snow of sub-Arctic areas. Using seven years of data from Resolute and from Aklavik (68°14'N, 134°50'W), Williams (1957) found that the snow in Resolute was always denser and harder than that of the sub-Arctic site at Alkavik (Fig.2.4).

The present study confirms Williams' observation that the Resolute snow density exceeds 350 kg m^{-3}. From sample snowpits dug in 1976 and 1977, snow hardness was found to exceed 30,000 kg m^{-2} except

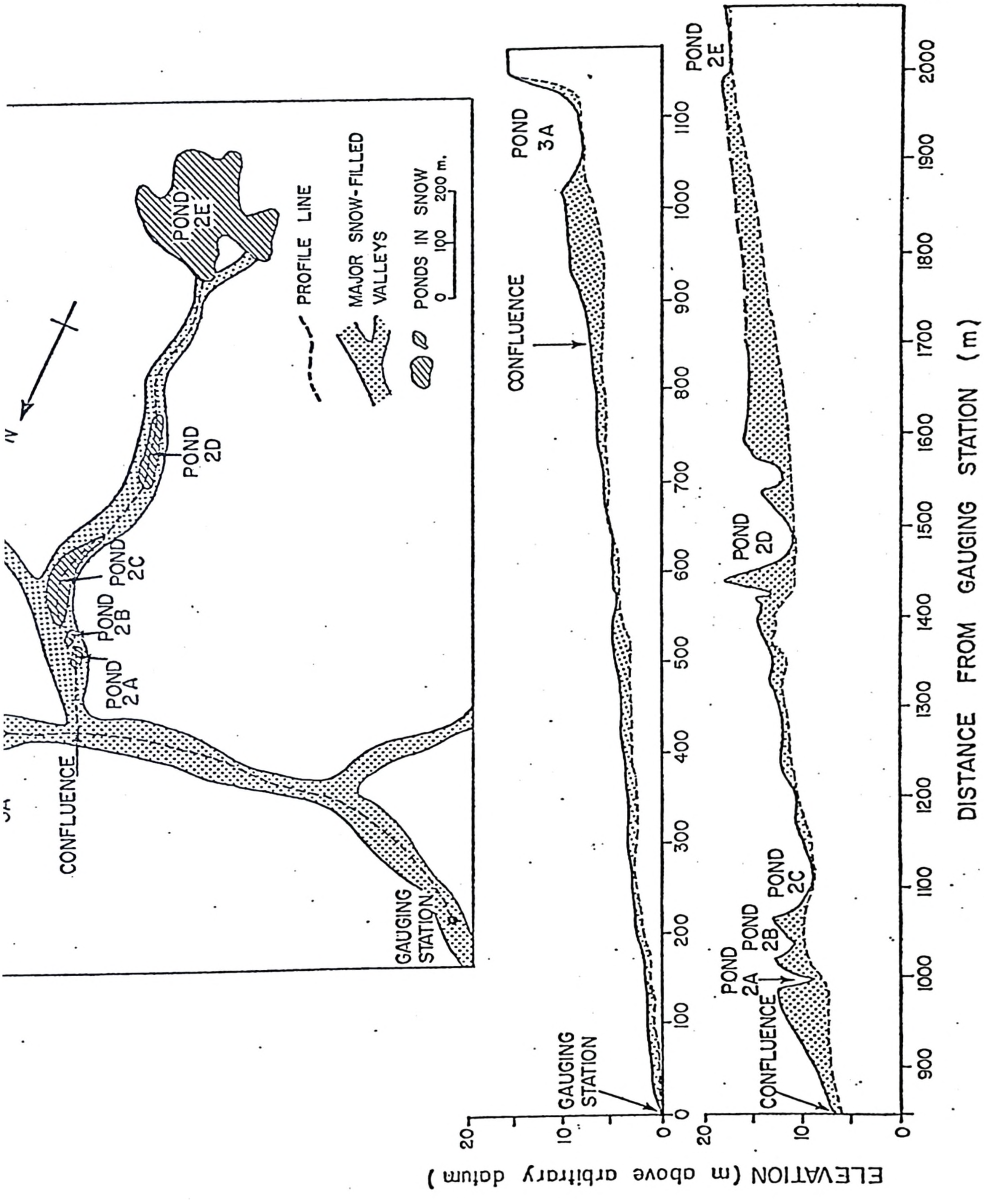


Fig.2.3 Longitudinal profiles of snow cover along two major valleys. An uneven snow profile caused meltwater pondage shown in the map above.

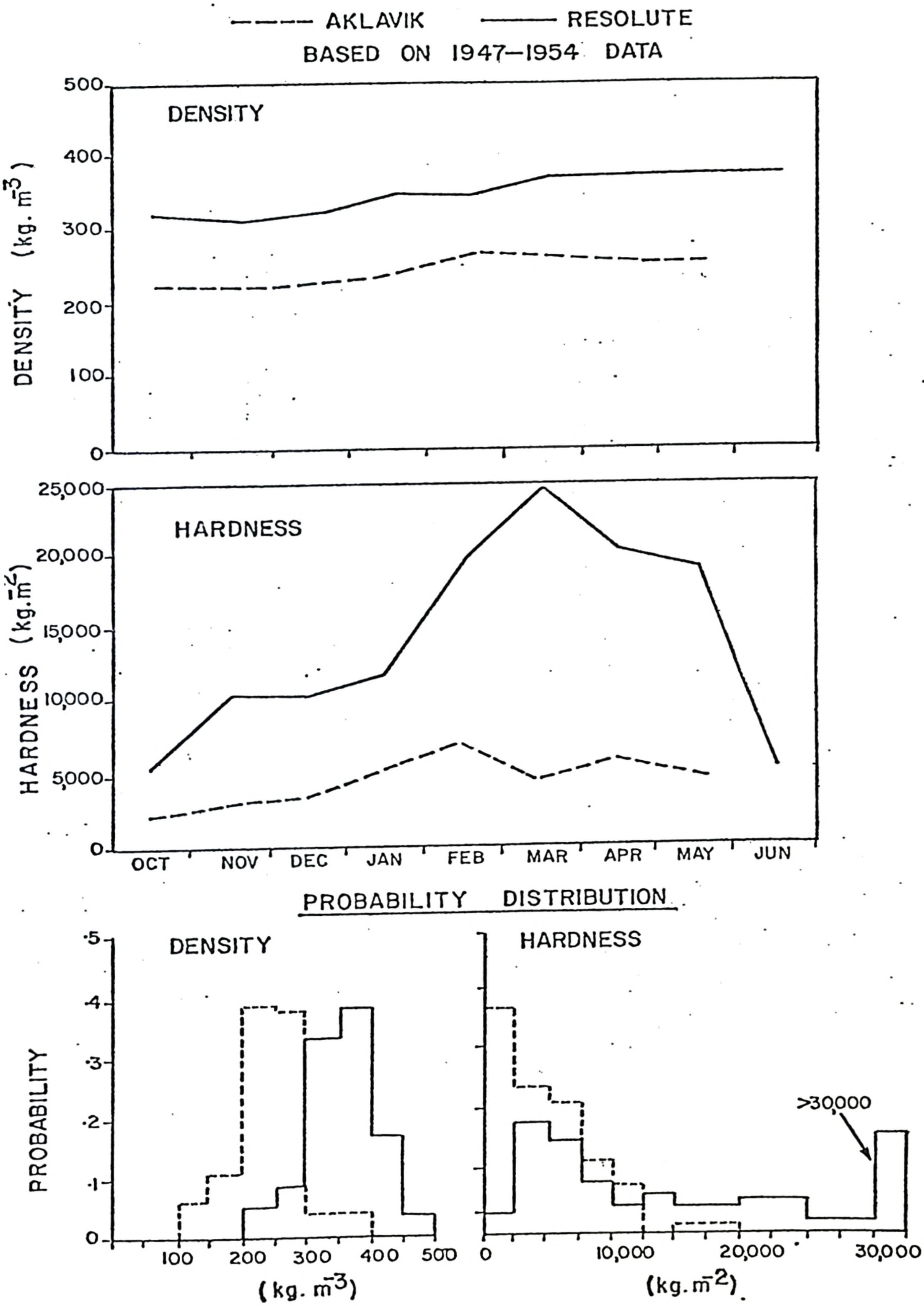


Fig.24. Mean snow density and snow hardness at a typical sub-Arctic site (Aklavik) and at a typical High Arctic site (Resolute). Also shown are the probability distributions of snow density and snow hardness (after Williams).

where depth hoar was encountered (the May 30th 1977 profile in Figure 2.5), or where the freshly fallen snow at the surface did not have sufficient time to be compacted (the May 22nd 1976 profile in Figure 2.5).

Thermally, the snowpack is quite cold at the end of winter. By the beginning of June 1977, snow temperature at the snow-ground interface of a typical site was -15°C. With the arrival of spring, the snowpack warmed up quickly. Although snowmelt did not take place until the air temperature rose above 0°C on June 10th, radiation from the 24-hour sun provided sufficient heat to the snowpack to raise the temperature of the entire pack to 0°C within four days (Heron and Woo 1978). Snowmelt followed and, between June 10th and June 24th, most of the basin was bare of a snow cover. The rapidity of snowmelt is partly due to the shallowness of the snowpack on the slopes and the hilltops. For the snow-filled valleys, the snowpack was much deeper and snowmelt runoff did not occur until the pack was saturated.

2.4 Topography

The basin area is 33 km^2 and the elevation ranges from 84 to 200 m. Topographically, the basin is fringed by plateaus or rounded ridge-tops, with rolling to hilly terrain occupying the central parts of the basin. A post-Pleistocene stream occupies the drainage basin which is composed of limestone of Ordovician and Silurian age (Cook 1967) underlying a relatively thin veneer of glacial deposits. Most of the ground surface is barren, with surficial materials ranging from

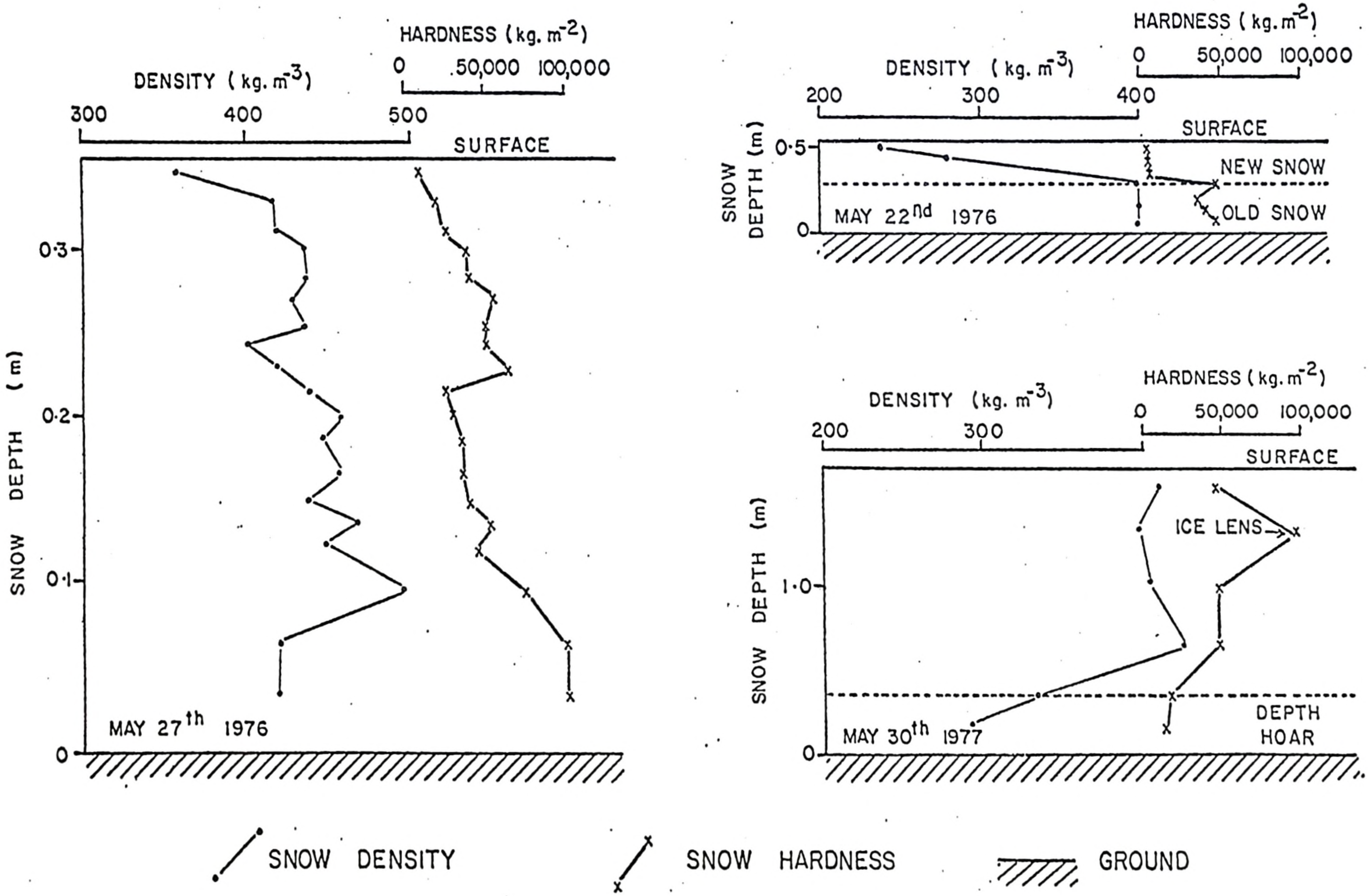


Fig.2.5. Snow density and snow hardness characteristics shown at three snow pits in the study area. Note that snow hardness and snow density decreased with new snow and decreased with depth hoar formation.

clay to boulders (Cruickshank 1971).

Permafrost underlies the entire basin at a depth of 0.3 to 0.7 m below the surface. It restricts all the hydrological activities to take place at the surface during the flow season.

The region has been recently glaciated, leaving many topographic depressions occupied by ponds in summer. Isostatic rebound since deglaciation has enhanced rapid incision of coastal streams to produce gorge-like valleys (St-Onge 1965). For this reason, alternate sequences of valley shapes are found over the basin, including non-incised valleys with narrow floor and with wide floor, and incised valleys with narrow floor and with wide floor (Fig. 2.6). The spatial distribution of these streams is also the result of geological controls; the stream network has a trellis drainage pattern commanded by its underlying eroded sedimentary layers gently dipping towards a west-northwest direction. Accordingly, subsequent and consequent streams were developed on this cuesta landscape (Fig. 2.1).

2.5 Data Collection

The data collection is composed of three components, which are as follows:

(1) survey of the snow disposition characteristics and observation of the temporal and spatial variations of channel development in snow-filled valleys.

(2) water discharge measurement of the streams, with some hydraulic considerations.

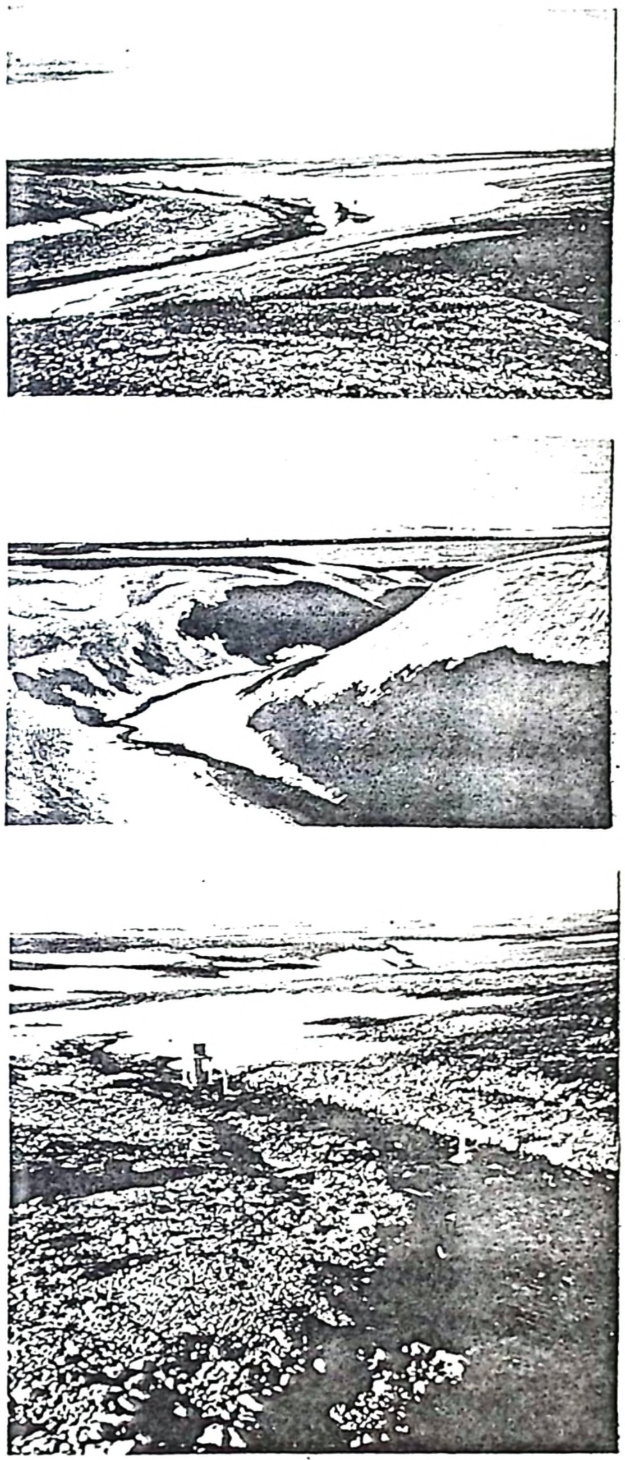


Figure 2.6 Examples of valley shapes: (a) incised-wide floor valley; (b) incised-narrow floor valley; (c) non incised-narrow floor valley.

(3) assessment of the magnitude of fluvial processes as indicated by the levels of sediment loads, with special reference to suspended, dissolved and bed loads.

2.5.1 Snow Distribution

Initial conditions of snow disposition in snow-filled valleys were investigated by the survey of longitudinal snow profiles (snow thalweg), and of potential ponding areas; the long profile was resurveyed at the end of the season under stable (i.e. snowpack free) channel conditions. The profiles were computed with a level and a surveying rod to the accuracy of 5 mm. A similar method was used for an assessment of a stage - water capacity relationship in potential ponding areas, where profiles were surveyed at different levels incremented by 1 metre from the lowest point, with measurements of the width of the pond at the height of every surveyed point. The monitoring of the accumulation and release of water in ponded areas was then reduced to a manual recording of water levels. Temporal variations of channel development in snow-filled valleys were illustrated by a sequence of snow cross profiles along typical snow accumulation sections. Using a Brunton compass, the inclination of the slope of the snow surface was measured to a sixth of a degree by a 2.44 metre long surveying rod averaging the gradient(Fig. 2.7).

At weekly intervals in summer, extensive surveys were carried out to observe channel development characteristics at all the snow-filled valleys of the basin. A comparison of such field information with the

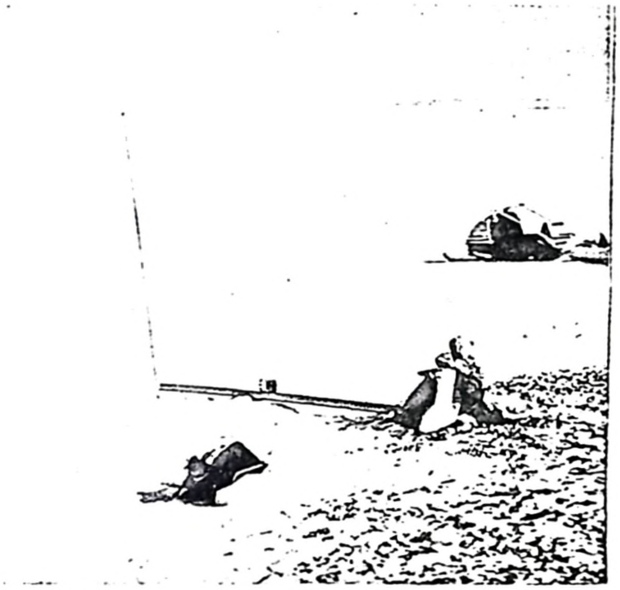


Figure 2.7 Cross sectional profile measurements of the snow surface.

late-lying snow conditions shown in a set of 1969 aerial photographs enables useful conclusions to be drawn about the recurring sequence of channel development patterns in the late-lying snowpack.

2.5.2 Streamflow

Prior to any channelled streamflow on the snow surface, the rise of the water level in the snowpack was monitored by a Leopold-Stevens type "F" water level recorder set in an in-snow stilling well. Moreover, a series of five 105 mm diameter, 0.6 metre long plastic pipes perforated by eight holders of 10 mm diameter was spread over the even snow surface at Site 4 (Fig. 2.8). Water level was manually recorded to a precision of one millimetre at a time interval of approximately one minute. This hydrostatic water level rise in the snowpack illustrated the enlarging contributing areas with time. Then stream discharge was measured in four locations in the basin. Until the period of stable channel was reached, no stage-discharge relationship could be considered because of the changing shape of snow channels. During that period, discharge was measured directly from the velocity - area method, where the velocity was determined by a "mini OTT" type current meter; diurnal fluctuations of the snowmelt flood were recorded twice daily per site. Once the channel had stabilised, a stage-discharge relationship was considered, where variations of stages were recorded by an OTT at Site 2, and a Leopold-Stevens type "F" water level recorder at each of Sites 1, 3, and 4.

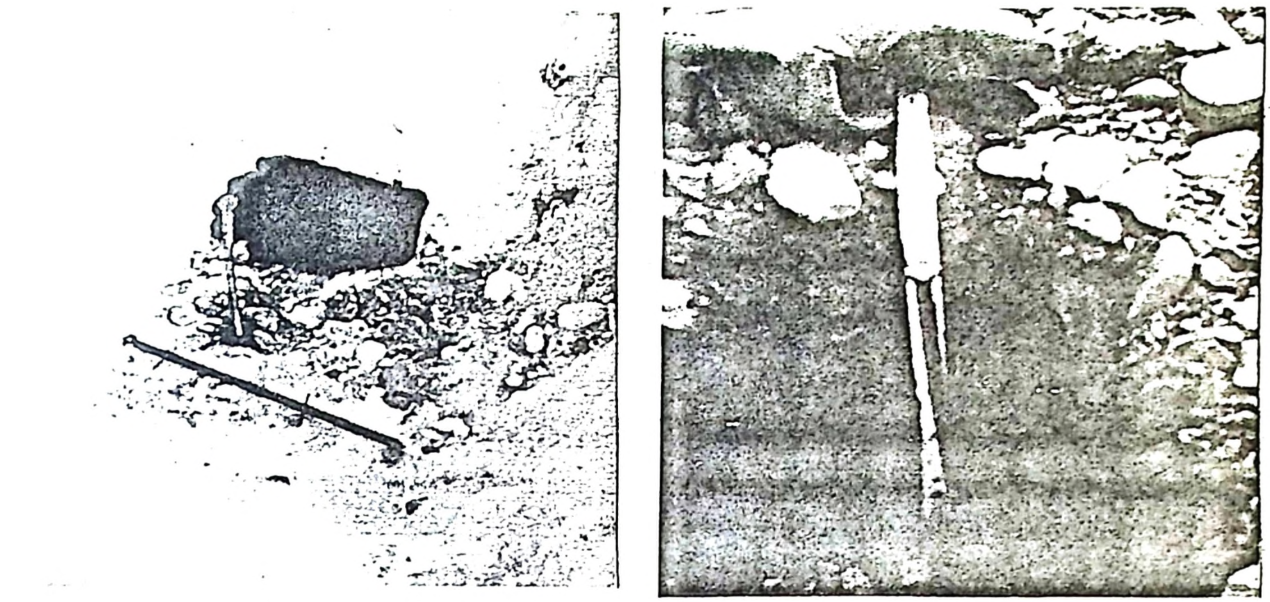


Figure 2.8 Hydraulic conductivity pipes used at Site 4 to measure the rise of the water level in the snowpack.

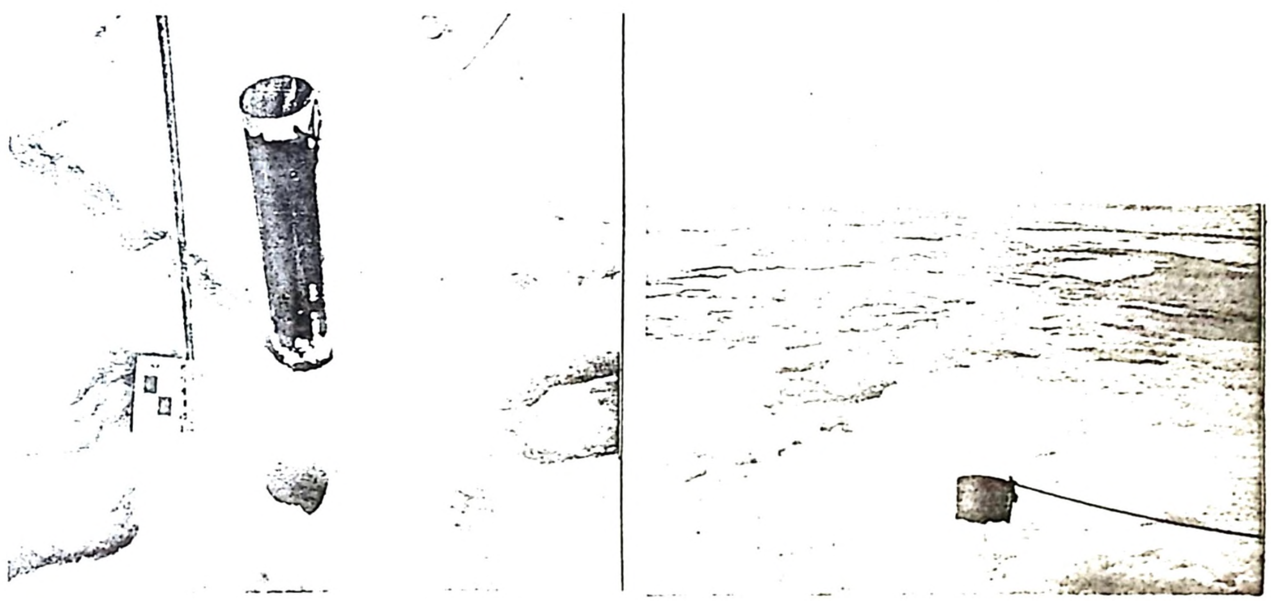


Figure 2.9 Erosion-deposition devices; (a) at set-up; (b) after the snowmelt peak runoff.

The hydraulic geometry of the opening channel was computed from the components of the velocity-area method. Moreover, the hydraulic slope was measured daily, with a greater frequency during the flood events. The hydraulic head was measured following Church and Kellerhals (1970)'s procedure, between two fixed levelled locations in the streambed.

2.5.3 Fluvial Processes

The magnitude of fluvial processes was measured as rates of sediment loads. Suspended and dissolved loads were assessed by a "width-depth-integrating" sampling program, using a hand-held DH-48 sampler, at sections 1, 2, 3 and 4. Frequency of sampling was dependent on daily transport fluctuations, i.e. twice daily during the snowmelt period and once for the post-snowmelt duration. An analysis of the suspended sediment content of these samples was done by the filtration method (Dingman 1966). The titration of the filtrates for calcium and total carbonates (Schwarzenbach 1957) yielded the bulk of the dissolved sediment concentration since the stream as located in a limestone basin. Water temperature and its conductivity measured in mOhms were also recorded to establish a conductivity - dissolved load relationship.

Bed load being the most difficult load component to measure was indirectly assessed by driving several 0.3 m spikes into the stream bed to a depth of 70 mm. Enclosing each spike was a steel washer which lay flush with the bed. These spike and washer units were deployed at different points along the channels. The principle of the device is that, at a point on the channel bed, degradation is monitored by the distance of the steel washer below its initial position shown by a grooved mark on the spike. Any subsequent aggradation is obtained as the thickness of the sediment above the steel washer level (Fig. 2.9). Two series of spike and washer measurements were recorded to differentiate between the snowmelt and post-snowmelt flow periods. Surveys were also undertaken of bed load material that had been mobilised during a short part of the snowmelt flood and deposited as a cast a former channel course over the snow surface. Also a line of painted pebbles of uniform size was spread across the channel bed to indicate bed load movement during the flow season period. Detailed monitoring of the rate of movement of the overloose material were performed by daily downstream distance measurements from the initial location of the line of painted pebbles.

2.5.4 Meteorology

Meteorological records were maintained at a site near the basin outlet. Such records include air temperature, humidity, radiation over snow surface, wind speed and rainfall.

CHAPTER THREE

CHANNEL DEVELOPMENT IN SNOW-FILLED VALLEYS

This study examines the manner in which the flow of meltwater carves channels in the valley snowpack before these channels are finally stabilised in the clastic bed. Channel development processes in snow-filled valleys are controlled by the distribution and the characteristics of snow, which in turn are related to the local topography and the prevailing directions of winter snowdrift. Enclosed herein is a description of channel development processes followed by an attempt to model the channel development sequences.

3.1 Channel Development Processes

To understand the mechanism of meltwater storage and release from the valley snowpack, the various processes of channel formation and development in the snowpack were observed. On the whole, the development sequence begins with snowpack saturation, and terminates when the channel is stabilised on a streambed of clastic materials.

3.1.1 Saturation of the Valley Snowpack

In spring, the thinner snowpack of the hilltops and the slopes begins to melt, but snow temperature within the valleys snowpack remains below 0°C. As meltwater from the slope gathers in the valley, some heat is dissipated in raising the valley snowpack temperature, and then heat is transmitted to the frozen ground beneath the snow. Before streamflow commences, it is common to find a layer of ice at the bottom of the snowpack. This bottom ice layer is either produced by the freezing of water remaining in the streambed during the previous winter, or it is due to a refreezing of the spring meltwater in contact with the frozen streambed, or both.

With a delay in the ripening of the valley snowpack, streamflow lags behind the melting of snow on the hillslopes. In 1977, hillslope snowpacks began to melt on June 11th. This meltwater moved downslope and was collected along sections of the valley. Eventually, downstream movement of this water took place. Flow progressed in steps and halts, depending on whether additional water was available en route, or whether water diffused in the snowpack to saturate its unripened portions. It was not until June 16th that runoff was observed at station 4 (Fig. 3.1).

3.1.2 Streamflow in Snow

Initial movement of meltwater in the valley takes place as subsurface flow in the snow, as sheetflow on the snow surface or as slush flow of wet snow. These flow conditions are then altered to channelled flow on the snow surface, in the snowpack as tunnels, or between the snowpack and the valley slope.

Non-channelled subsurface flow occurs when water in the saturated valley snowpack diffuses downstream, collecting additional meltwater from the ripen snowpack along the flow path. At any site in the valley,

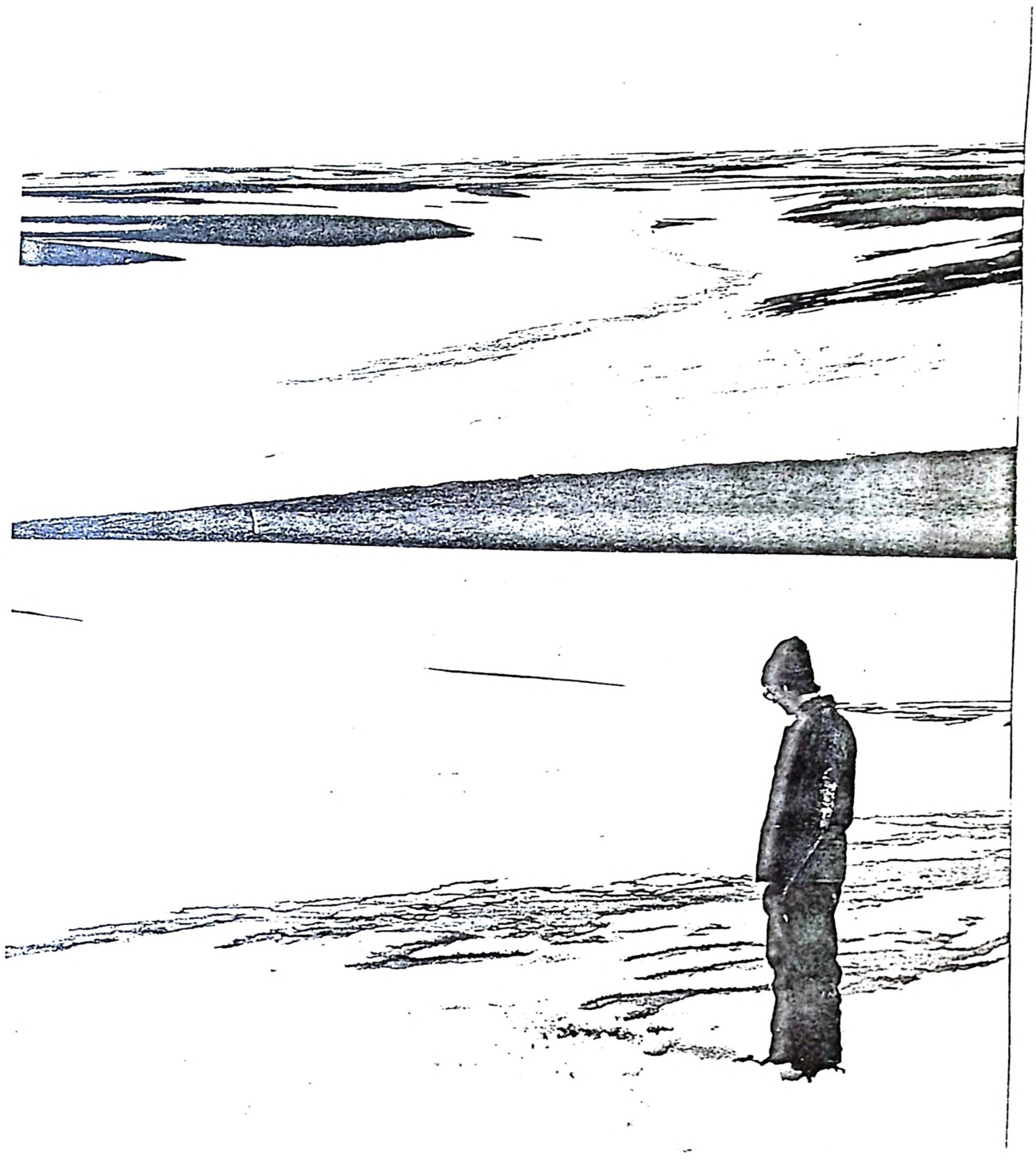


Figure 3.1 Upper photograph shows incipient downstream water movement along a snow-filled valley near Station 4 (see Fig. 3.3 for location). Lower photograph is a close-up at the flow front. Surface runoff occurs as sheet flow on the snow, but water infiltrates continuously at the leading edge to saturate the unripened snowpack.

the arrival of such downstream flow is indicated by a sharp rise of the hydrostatic water level in the snow. One example of this process was observed along a 1 km reach above the stage recorder at site 4. Between June 14th and 16th, the front of subsurface flow travelled at an average rate of 5×10^{-3} m.s^{-1} over a channel slope of 8.3×10^{-3}. Sharp rises in hydrostatic water level corresponded with the passage of this front(Fig.3.2). Based on the time of hydrograph rises, isochrones are obtained (Fig.3.3). In detail, the front was not straight, but was locally deflected by the uneven depths of the valley snowpack.

When the hydrostatic water level rises above the snowpack surface, water moves as sheet flow over the snow (Fig.3.2). Since surface runoff travels faster than subsurface flow, sheet flow often proceeds as a broad but shallow water layer ahead of the subsurface flow. At the leading edge of sheet flow, water infiltrates continuously into the non-saturated snow beneath. Downstream progress of sheet flow is arrested when infiltration loss exceeds surface water supply from upstream.

Slush flow is defined by Washburn (1973) as the predominantly linear flow of water-saturated snow. He commented that it was "the common method of stream breakup" in places like the Mesters Vig district of northeastern Greenland. Such mode of flow was not too common at the study area. The possible difference with the Mesters Vig district is a lower valley floor gradient at Resolute. Although several cases of slush flow were documented at the upper parts of our study basin,with generally steeper slopes (Fig. 3.4), the prevalence of gentle slopes appears to favour downstream movement of water along rather than a mixture of water and snow.

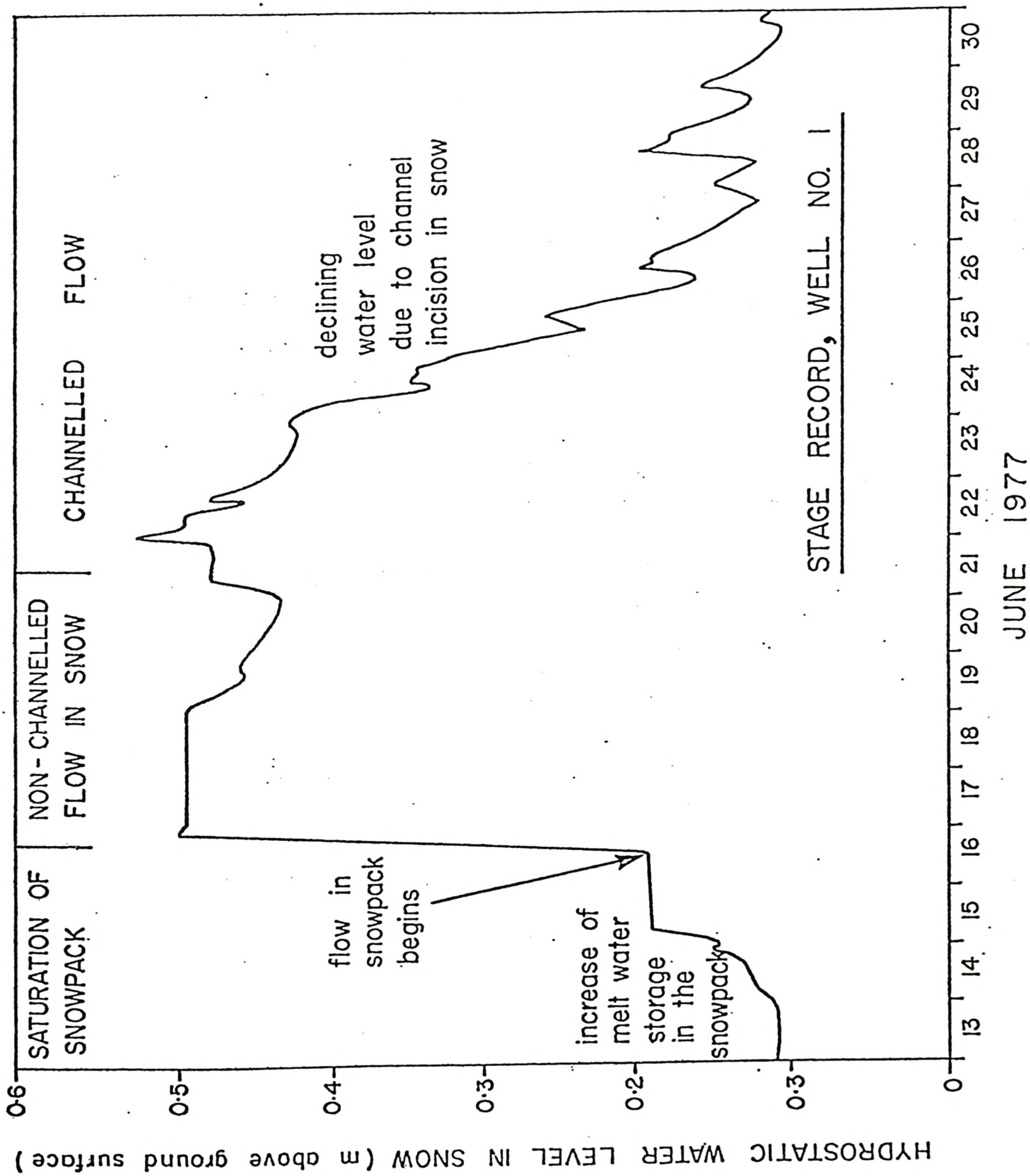


Fig. 3.2 Stage record at the outlet of Basin 4 showing hydrograph responses to (1) snowpack saturation, (2) non-channelled flow in snow and (3) subsequent streamflow in a channel carved in the snow. Note an abrupt rise in the hydrostatic water level when non-channelled flow from upstream reaches the sites.

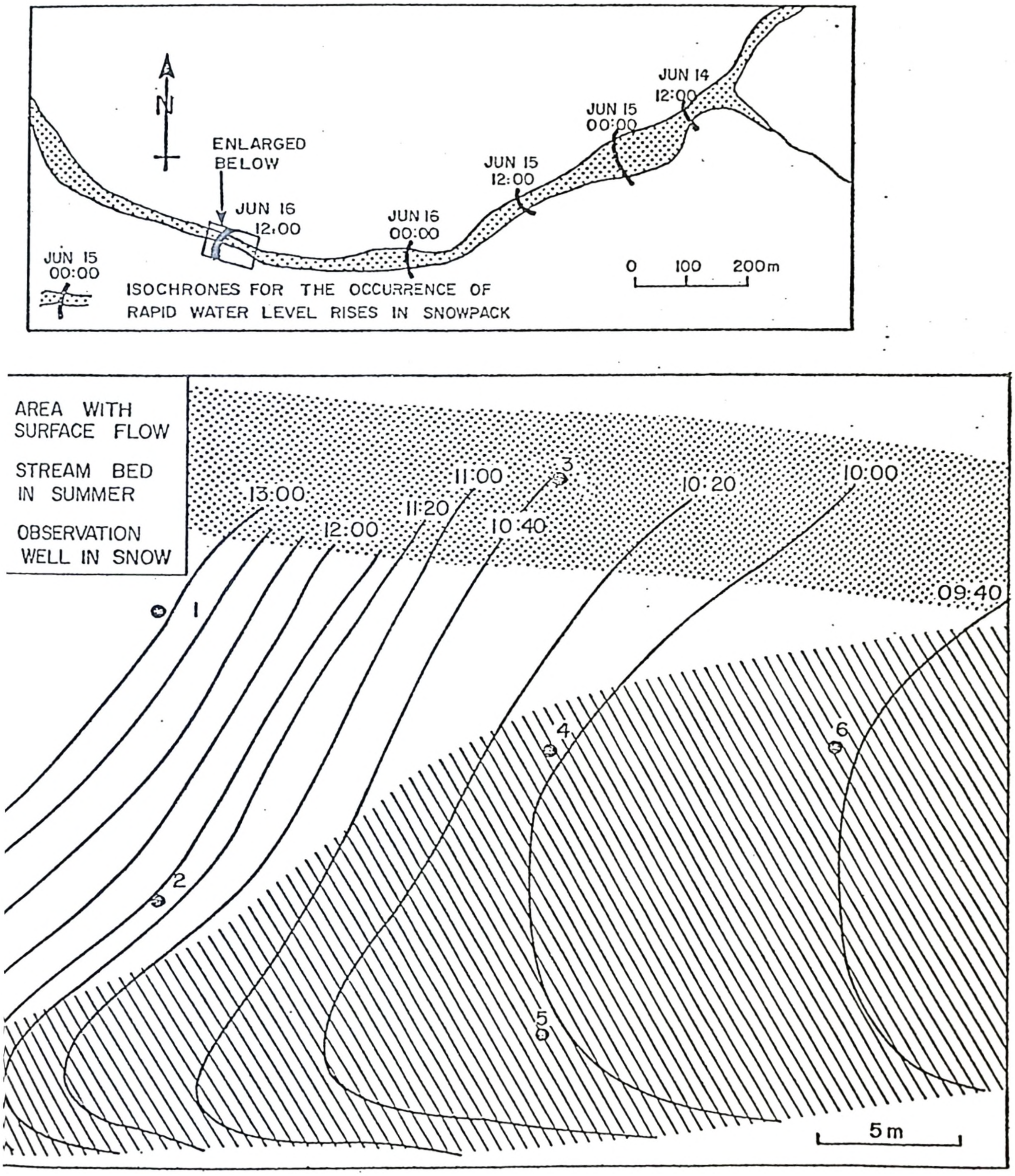


Fig. 3.3 Isochrones showing the movement of non-channelled flow within the valley snowpack at the outlet of Basin 4.

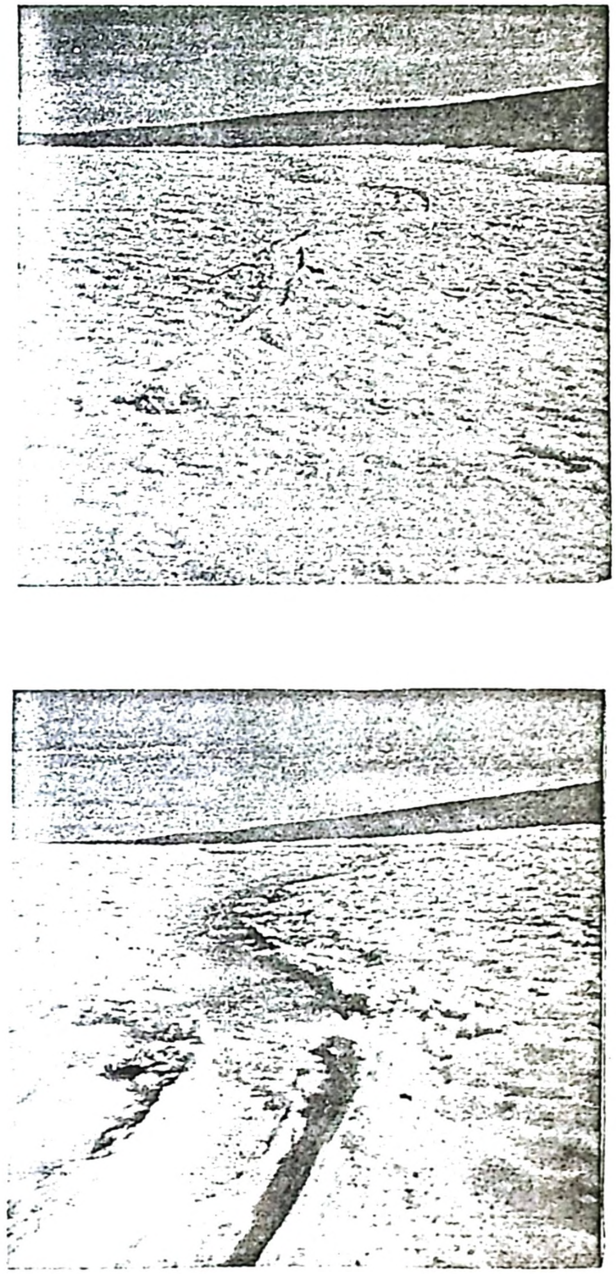


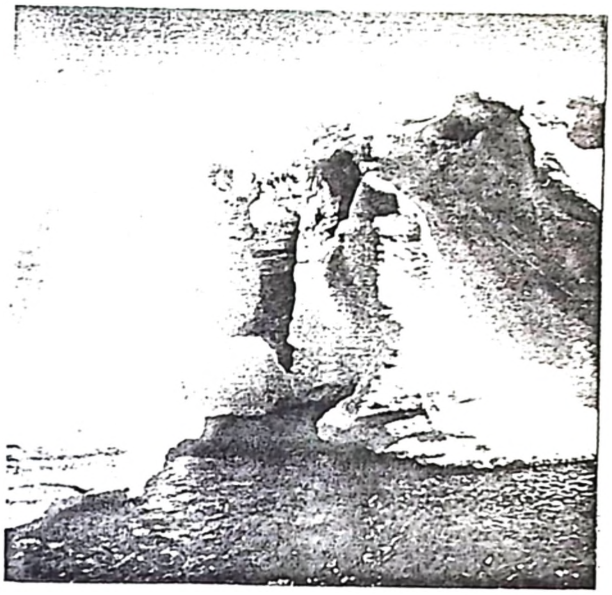
Figure 3.4 Small scale slush flow in tributary at upper part of the basin.

Channelling of flow often follows the initial phase of subsurface flow, sheet flow or slush flow. In deep snowpacks, a channel develops on the snow surface or within the pack. When open channel flow occurs on the snow, continual incision of the snowbank causes lateral or vertical shifting of the channel until the valley floor is reached (Fig. 3.5 - 3.6). In shallower packs, channels soon develop partially or entirely on clastic beds (Fig. 3.7). These channels may be located at the site of the valley where the local snow surface is the lowest. Channels thus created are sometimes flanked by the valley slope on one bank and by the valley snowpack on the other (Fig. 3.8).

One special form of channelled flow occurs in snow tunnels (Fig. 3.9). Tunnel formation is favoured by valleys deeply filled with snow. In these cases, the hydrostatic water level in the snowpack will not reach the upper portion of the pack. When subsurface flow has melted or removed the bottom snow to form a tunnel, the cohesive but non-saturated snow ensures that the roof does not collapse. Within the tunnel, water can flow on snow, on a layer of bottom ice, or on a streambed of clastic materials. The distribution of tunnel reaches is shown in Figure 3.9. The occurrence of snow tunnels is not restricted to the High Arctic. In northern Japan, Naruse et al. (1976) also reported a tunnel developed in a patch of avalanche snow which covered a small, steep valley.

3.1.3 Formation and Draining of Ponds in Valleys

Along some gorge-like valleys, winter snow drift produces troughs and ridges in the valley snowpack. In spring, meltwater from the hillslope



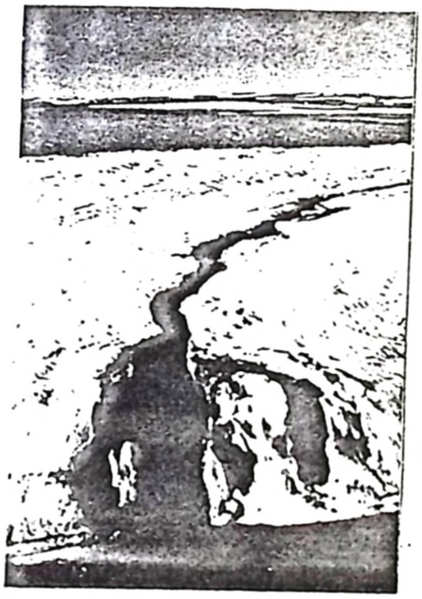


Figure 3.5 Vertical shifting of the channel until the valley floor is attained in a deep snowpack reach, small tributary joining Site 2.

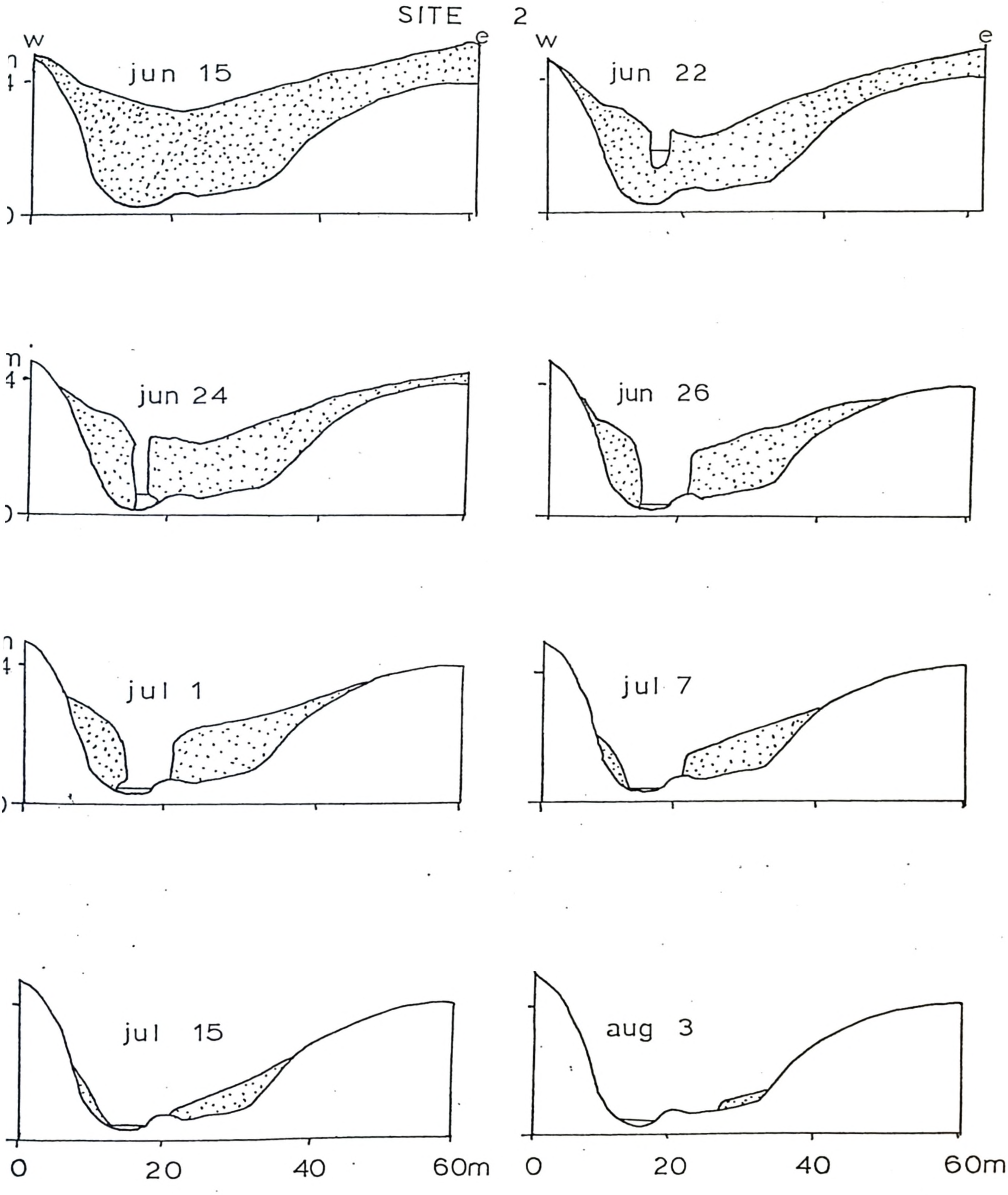


Figure 3.6 Time series showing vertical incision at Site 2.

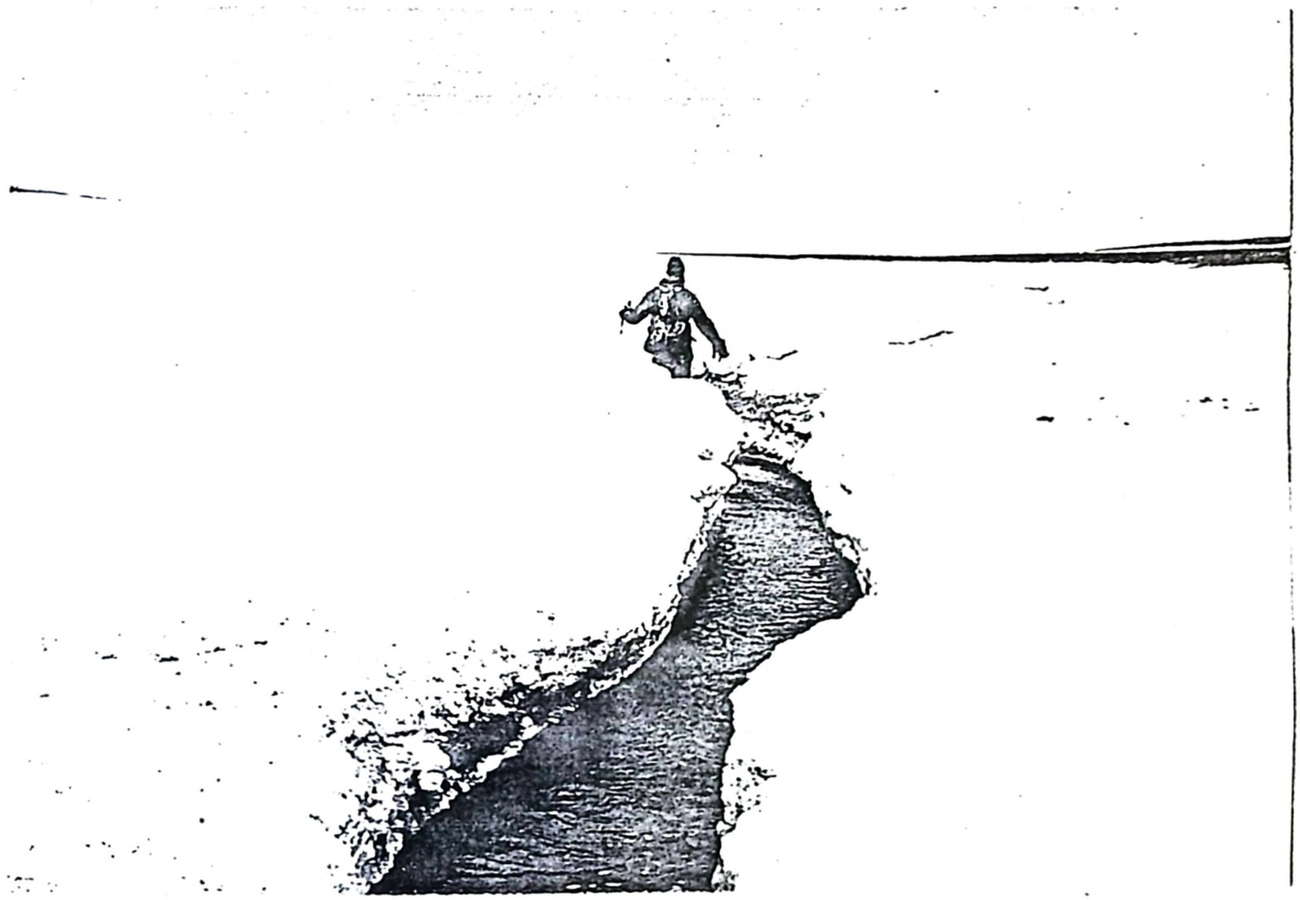


Figure 3.7 Channel developed in a relatively shallow snowpack in non-incised valleys. Downcutting brought the channel to the streambed four days after initial flow occurred.

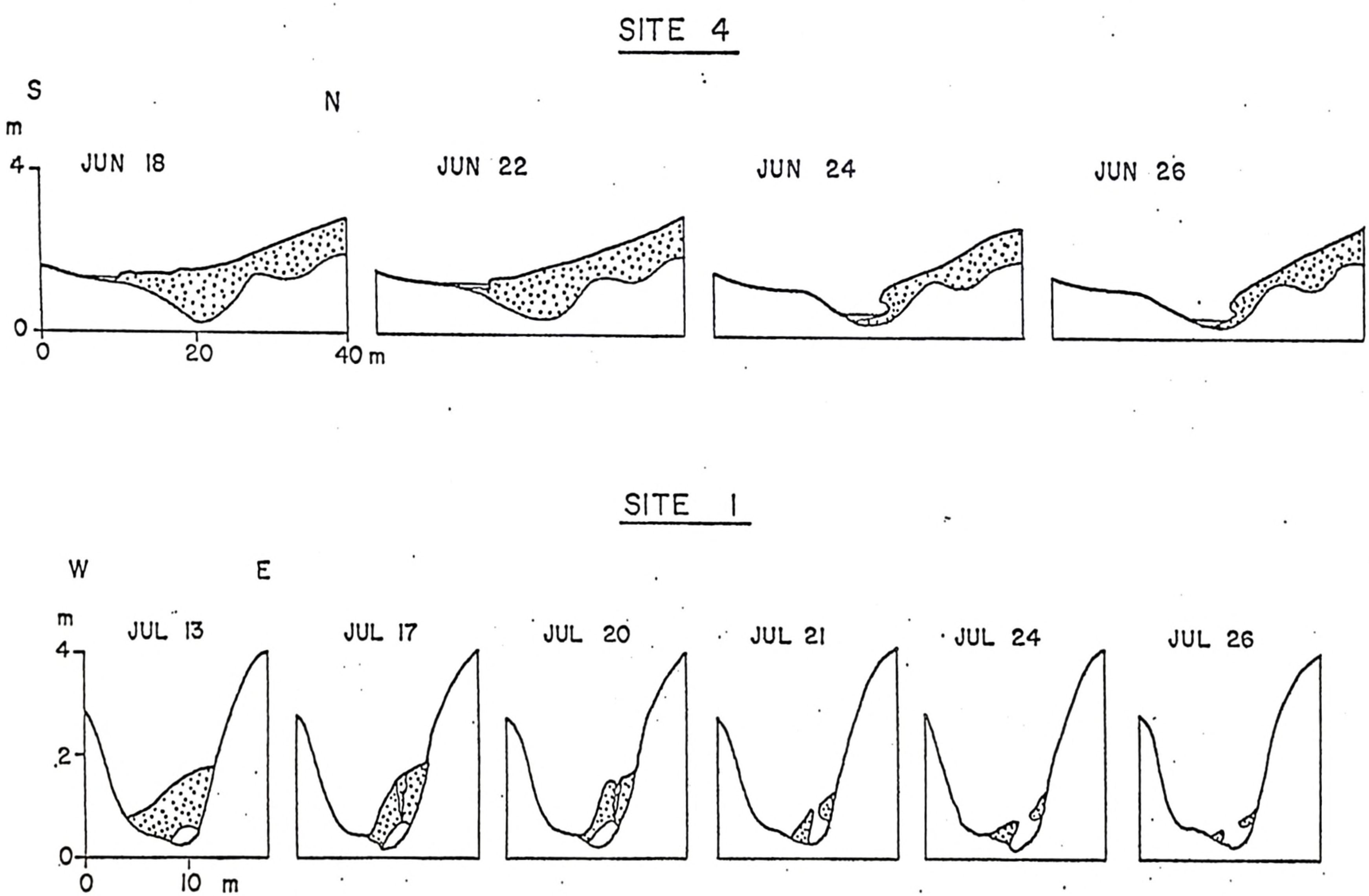


Fig. 3.8 Site 4 shows a lateral shifting sequence of a channel formed at the edge of a valley snowpack. Site 1 gives an example of snow tunnel formation and decay.

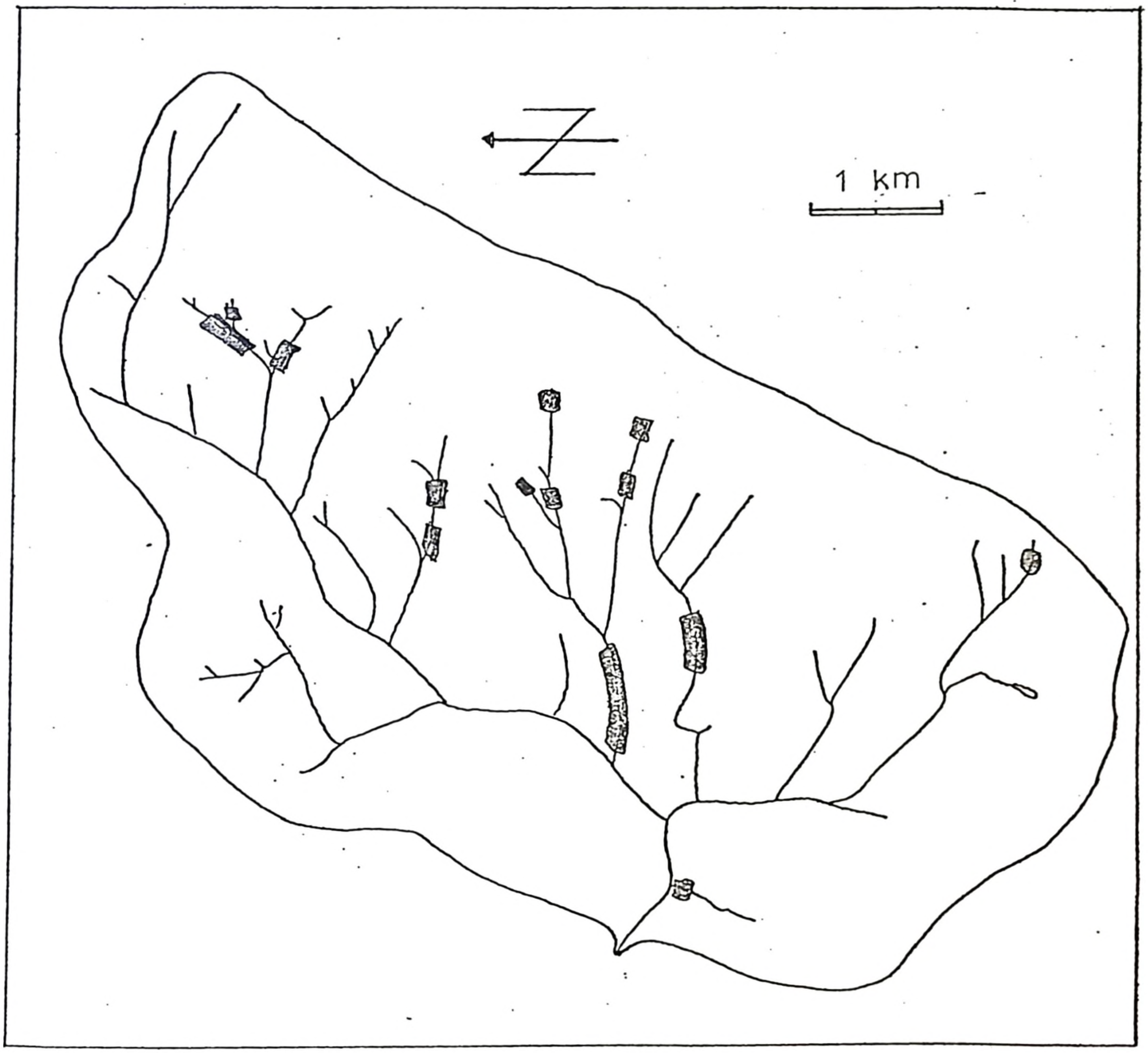


Figure 3.9 Distribution of tunnel reaches in the snow-filled channel of the basin.

and runoff from upstream are collected in depressions behind the snow ridges. This creates ponds whose infilling continues so long as the snow ridges remain (Fig. 3.10).

One such sequence of ponds along a tributary valley was observed during the snowmelt period of 1977. Infilling of these ponds began in mid-June. As shown in Figure 3.11 ponds 2A and 2B were small, but ponds 2C and 2D had maximum storage capacities of 5,600 m^3 and 5,900 m^3. Pond 2E was impounded by a snow ridge across the valley constriction below an alluvial flat, thus a large volume of meltwater (over 12,000 m^3) flooded an area of 35,000 m^2 (Fig. 3.11).

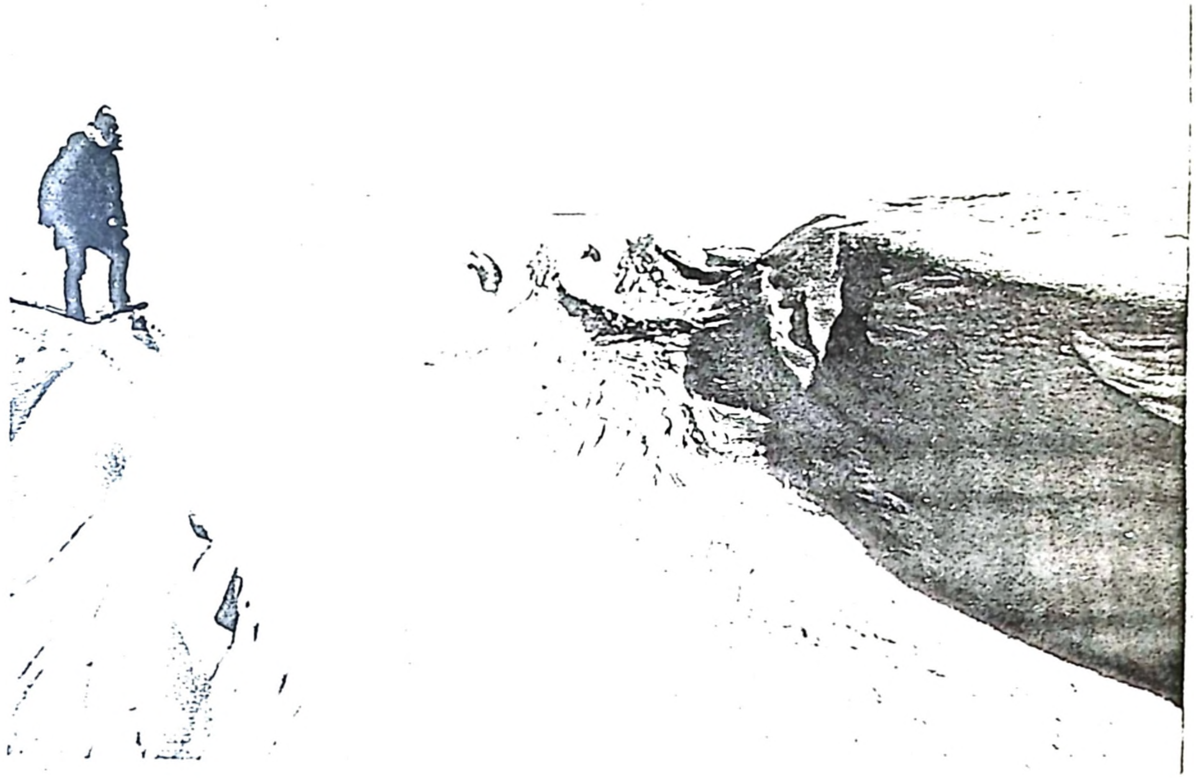
Ponds thus created often drain rapidly when the snow dams disintegrate. Disintegration involves one or both of the following processes:

(1) a seepage of water through the snow dam until it fissures and finally collapses

(2) pond water spilling over the snow dam followed by vertical cutting through the snow to produce an outflow channel.

The draining of ponds 2E to 2A took place in June 1977. On the 19th, a maximum capacity was reached in pond 2E but ponds 2D and 2C were partially filled. On the following day, the snow dam of pond 2E began to fail and water cascaded downstream, quickly filling the lower ponds. The snow drift dam of pond 2D then collapsed by fissuring and the snow dam of pond 2C was carved by a channel formed in the snow. By June 24th, ponds 2D and 2E were totally drained (Figs. 3.12 and 3.13).

Hydrologically, the breakup of snow drift dams produces a sudden release of water, with a resulting rapid clearance of the snowpack in the valley below the dam.



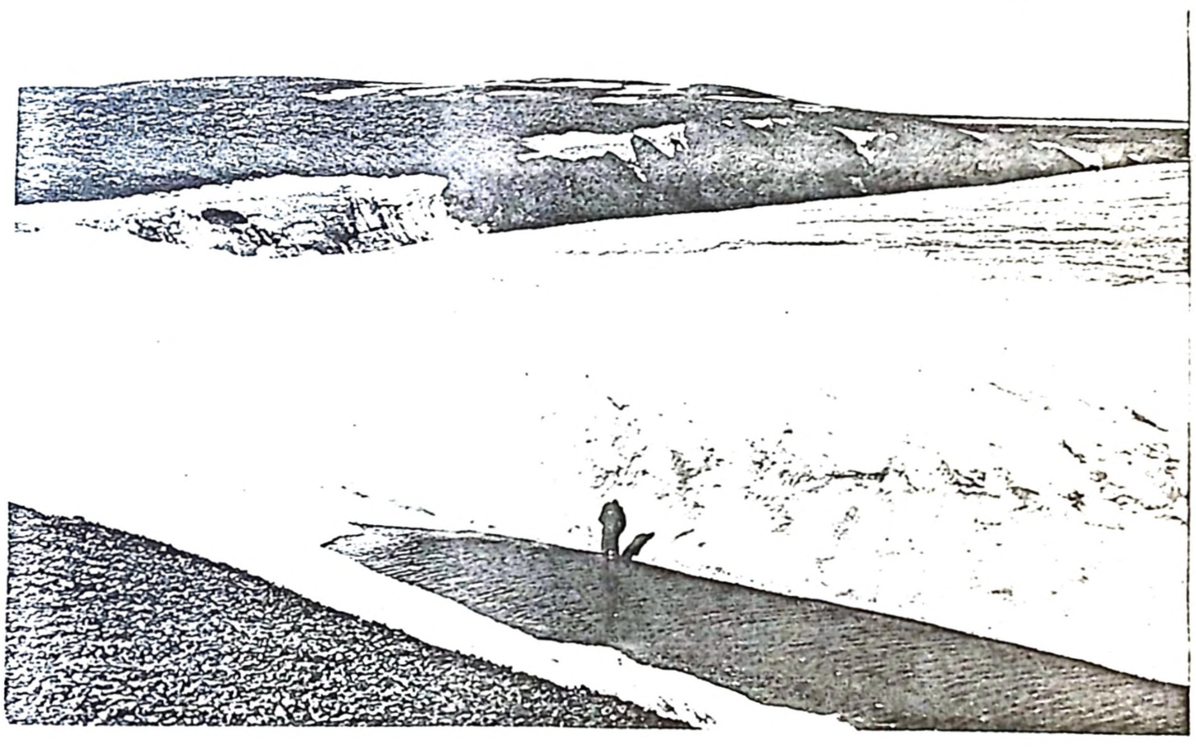


Figure 3.10 Upper photograph shows a snow dam spanning across a valley, leaving a hollow at the foreground which was eventually filled with melt-water in late June to form Pond 2D (Lower photograph).

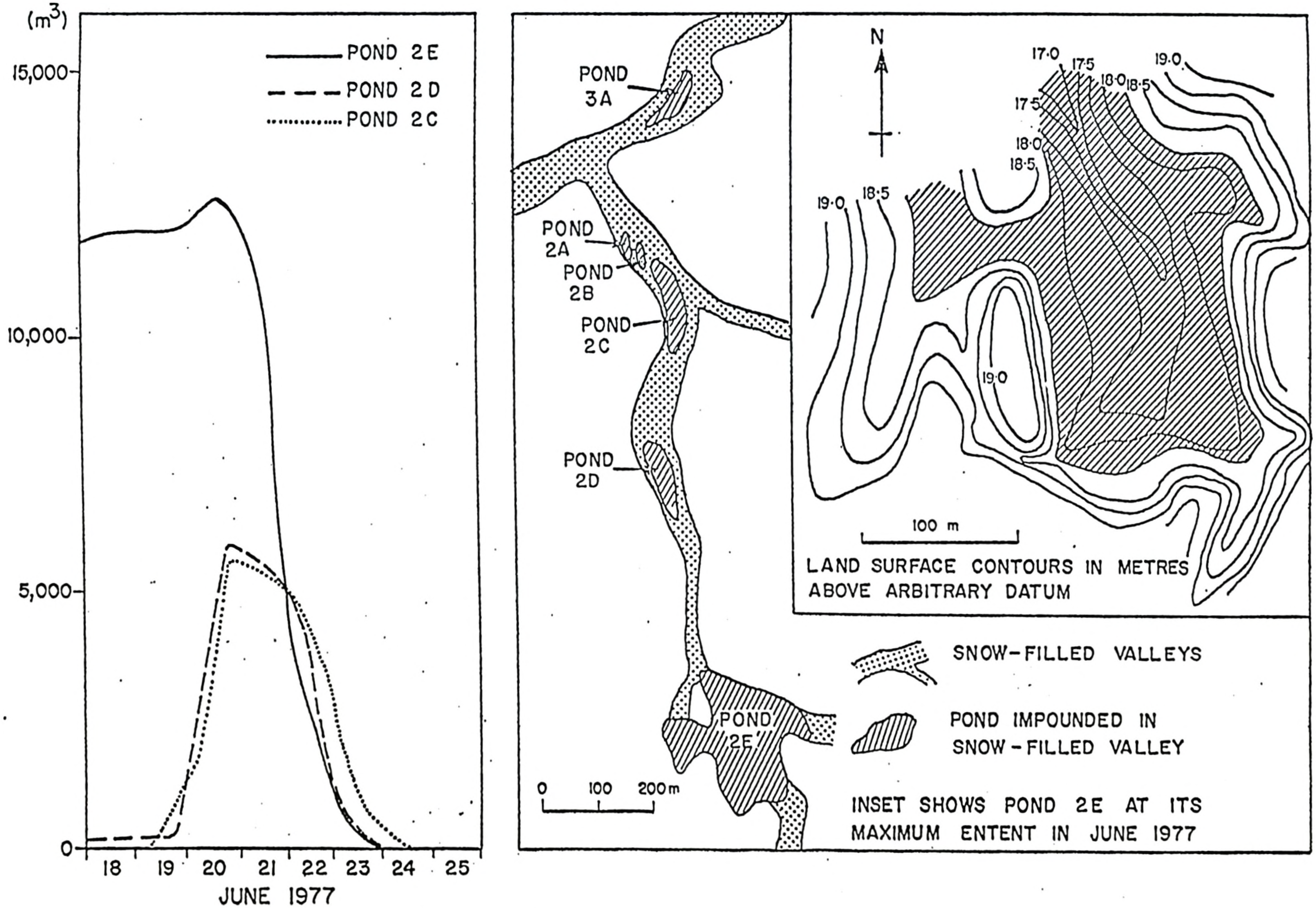


Figure 3.11 Change in the volume of storage in three snow-dammed ponds along a snow-filled valley. The location of these ponds and a detailed contour map of the largest pond are also included.

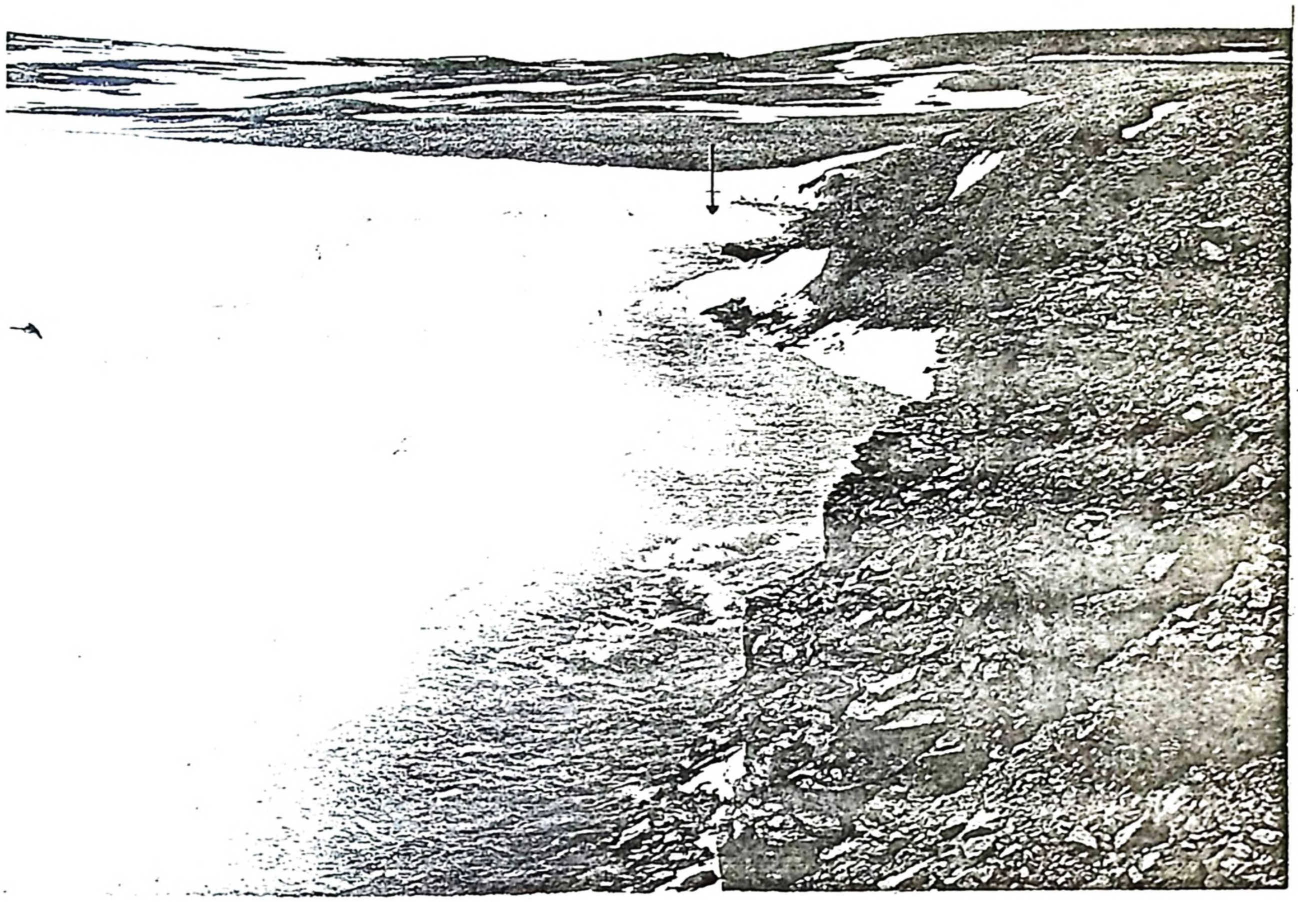


Figure 3.12 Snow ridge damming Pond 2D (shown by an arrow) broken on June 20 producing flash flood downstream.

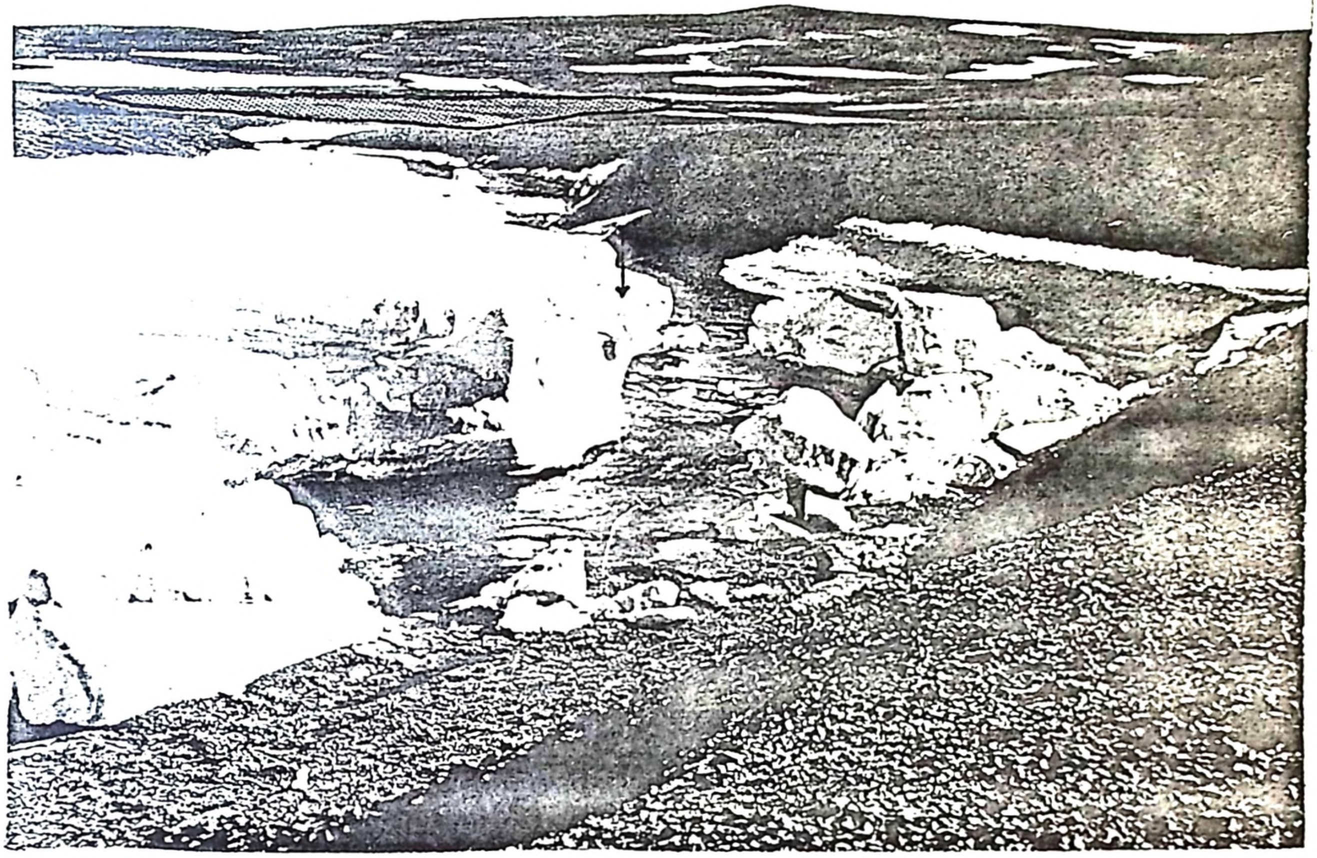


Figure 3.13 The decaying snow dam of Pond 2D (shown by an arrow) observed in mid-July. The maximum extend of Pond 2E is shown as shaded areas. After the draining of the ponds, lateral shifting of the channel produced snow blocks lying across the valley, with snow wall on the left exceeding of 8 m.

3.1.4 Disintegration of Channel Snowpacks

Valley snowpacks undergo continuous melting or erosion by streamflow. Melting causes a gradual reduction in the size of the pack while lateral cutting accelerates snow bank collapse.

In gorge-like valleys where the snowpack is deep, lateral undercutting leads to the calving of massive snow blocks (Fig. 3.14). Very often, the snow blocks fallen across the channel impound pools of water, causing a sudden decrease in streamflow downstream of the blockage. Streamflow decrease lasts from several minutes to three hours when the block is finally removed or when a channel is formed to bypass the blockage. The pool then drains, and this is accompained by a surge of discharge downstream. Such hydrologic phenomenon is comparable to those of proglacial streams which show short-term discharge fluctuations due to localised calving of the glacier ice across the channel (Wendler et al. 1972).

Streamflow irregularities caused by fallen snow blocks were observed in detail using the hydrograph record for Basin 2. The gauging site is indicated in Figure 2.1 and the study period was from June 25th to July 11th, after all the snow-dammed ponds in the valley were drained. During this period, the stream above the gauging site shifted its channel laterally, undercutting the snowbank and causing extensive snow block collapses (Fig. 3.14). This produced irregularities in the stage record at the gauging site (Fig. 3.15).

A qualitative comparison of solar radiation intensity, stream discharge and the relative magnitude and frequency of stage fluctuations suggests that:

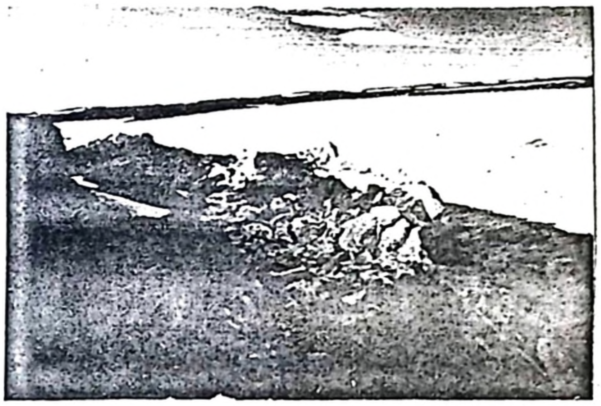


Figure 3.14 Calving of massive snow blocks during lateral undercutting of snowpack near Site 2.

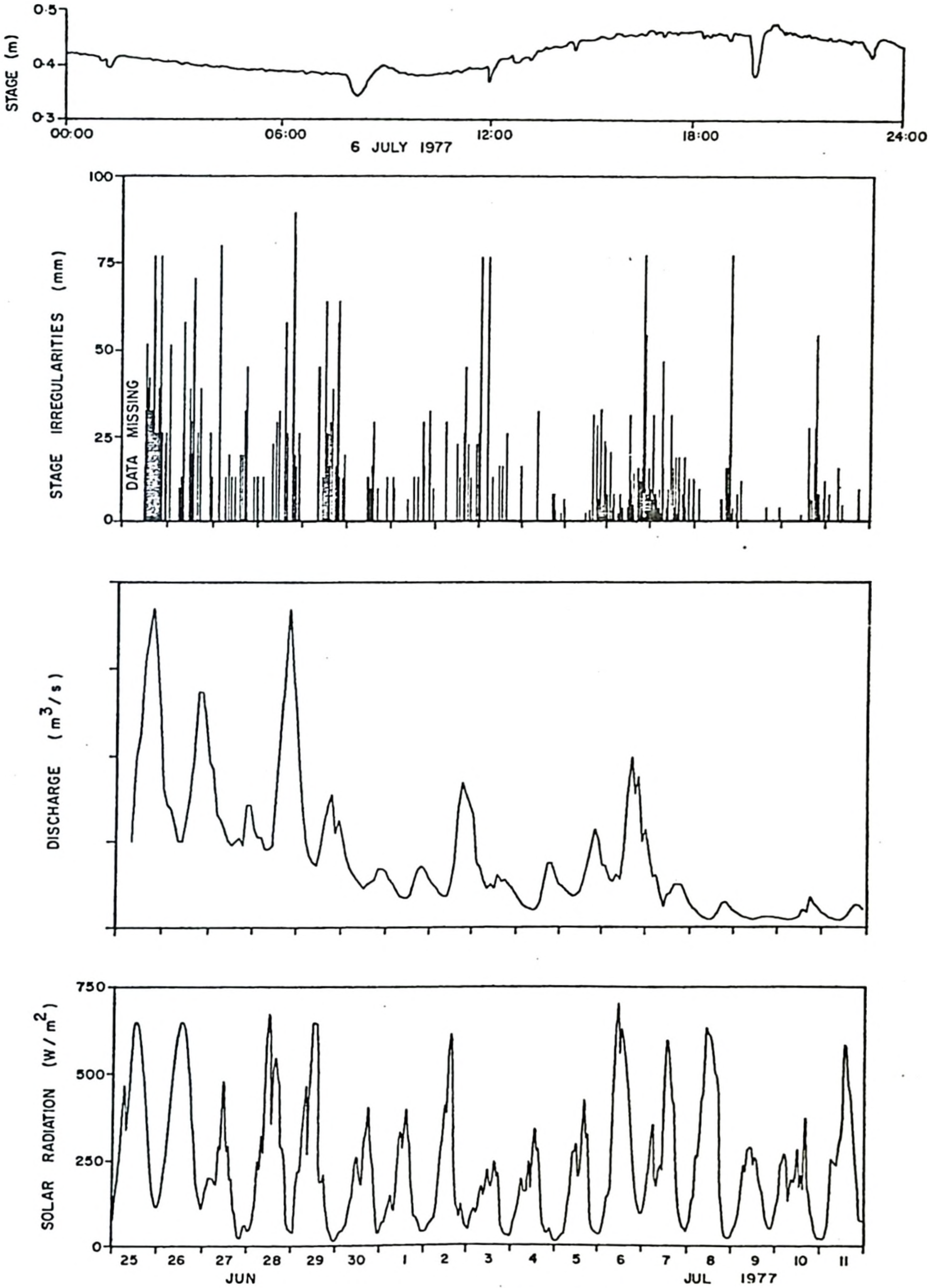


Figure 3.15 Top graph shows hydrograph irregularities appearing on the streamflow record of Basin 2. The other graphs show the magnitude and the time of occurrence of hydrograph irregularities during the snowmelt runoff period and compares their occurrence with the diurnal variations of discharge and solar radiation.

(1) the largest stage irregularities of each day occur about six to eight hours after solar noon, corresponding with the time of daily high flow

(2) a reduction in the frequency and the magnitude of stage irregularities occurred during the days of low radiation intensity

(3) the frequency and the magnitude of such irregularities decreased as summer advanced.

These observations indicate an acceleration of snow block collapse in relation to the energy of streamflow which in turn is related to radiation via snowmelt. A general decline of snow block collapse in mid-July shows combination of snowpack depletion and the cessation of lateral cutting activities.

Snow collapse is also common on the roofs of tunnels developed in snow. Snow roof collapse is often preceded by the formation of crevices (Fig. 3.16). Alternatively, the roof can be removed by continuous melting. In both cases, the tunnel degenerates into several sections of snow arches straddling the stream, representing remnant segments of the tunnel. The lower half of Figure 3.8 is a time sequence of tunnel roof ablation at a typical stream segment and Figure 3.17 shows the same tunnel after considerable ablation has reduced the thickness of the snow roof.

3.1.5 Channel Shifting

The initial position of a channel formed in the snow seldom corresponds with its ultimate position at the end of the melt season. Channel shifting can be gradual if it is accomplished by downcutting and lateral cutting in the snow. When flow diversion takes place, however, channel shifting becomes very sudden.

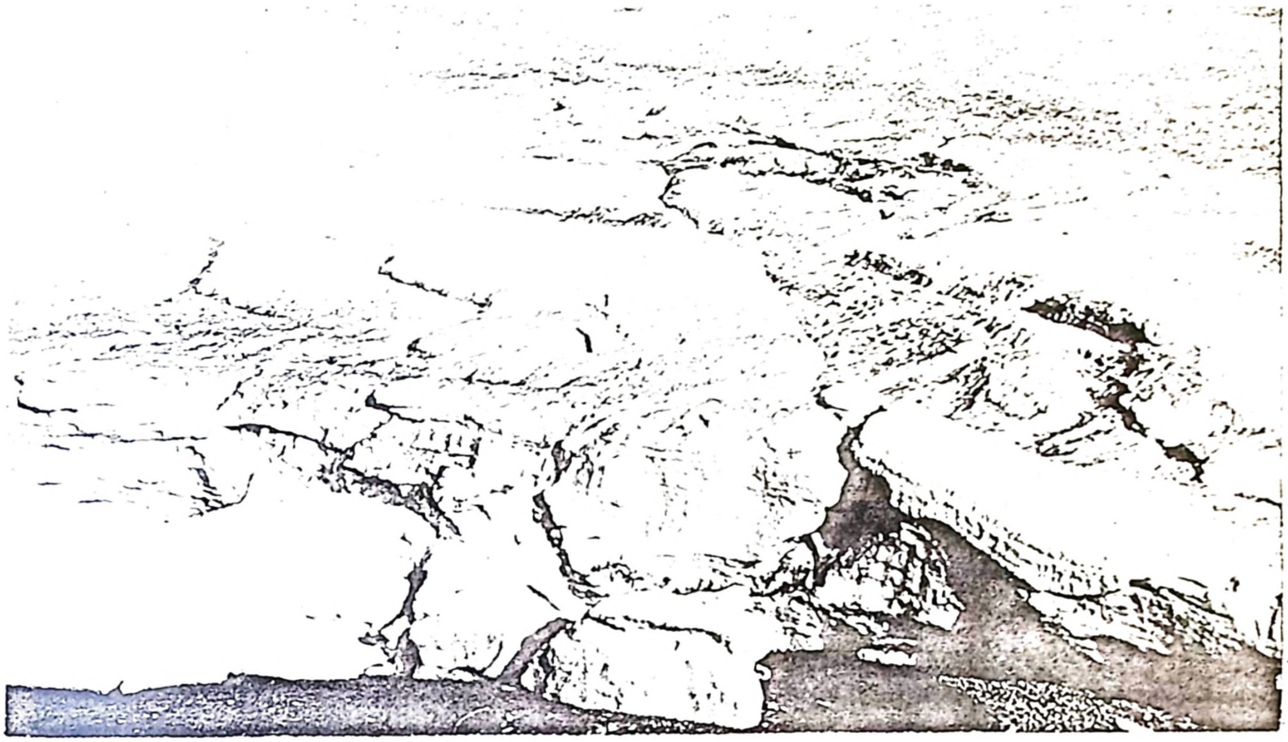


Figure 3.16 Snow roof collapse preceded by the formation of crevices.

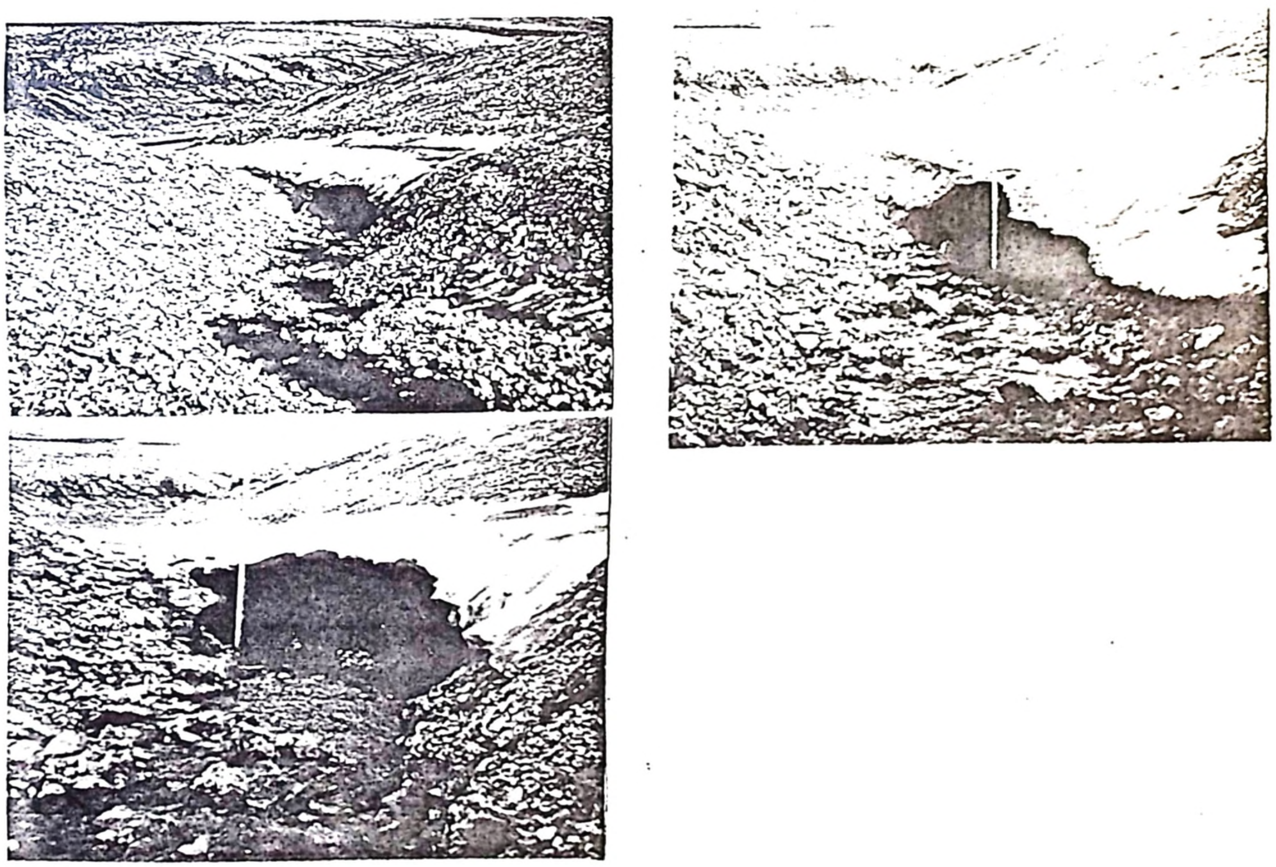


Figure 3.17 Three-fold sequence showing the thinning of the snow roof of a tunnel at Site 1.

The most common mode of channel development is by undercutting and by lateral cutting in the valley snowpack. Stream flowing on the snow incises continuously until it reaches the ground, with one or both channel banks fringed by snow. If the channel thus developed is not located at the lowest point of the valley floor, the stream will undermine its adjacent snowpack, thus causing lateral shifting. When the snowbank is thick, lateral shifting will induce snow block collapses as described in the previous section.

Downcutting and lateral shifting are exemplified by site 3 (Fig.3.18). On June 22nd, a channel was developed in the snow. After four days of downcutting in the snow, the stream reached the ground, but not at the lowest point of the valley. Lateral cutting followed as the channel shifted southward. A stable channel was finally established in mid-July and the residual snow on the banks decayed by melting.

An example of channel shifting by flow diversion is afforded by a small stream above site 1. In mid-June, a channel was formed on the snow at the western edge of the valley. This was abandoned on June 25th when a tunnel was formed in the snowpack (Fig. 3.19), leaving a substantial amount of bedload in the abandoned channel on the snow. Upstream of the tunnel reach another example of flow diversion by capture is shown in Figure 3.20.

3.2. Channel Development Sequences in Snow-filled Valleys

In view of a close relationship between topography and snow accumulation pattern, the disposition of snow in the valleys tends to recur

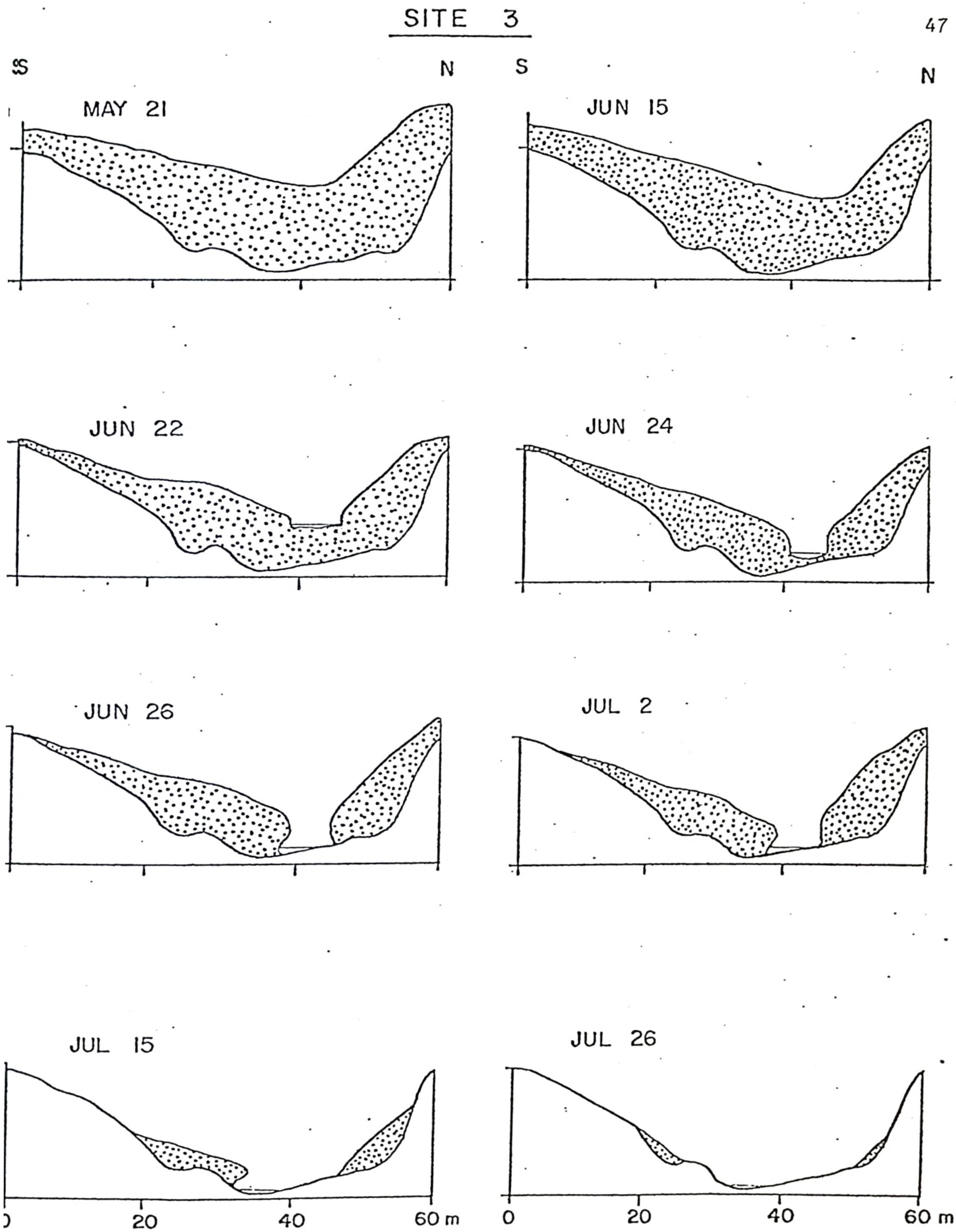


Figure 3.18 Sequence of channel development in the valley snowpack at Site 3 (near the outlet of Basin 3) showing vertical cutting and lateral shifting of the channel in the snowpack.

Figure 3.19 Streamflow diversion caused by the formation of a tunnel in the valley snowpack (indicated by an arrow). The abandoned channel on the right was left with a substantial amount of former bedload materials forming debris cones on the snow surface.

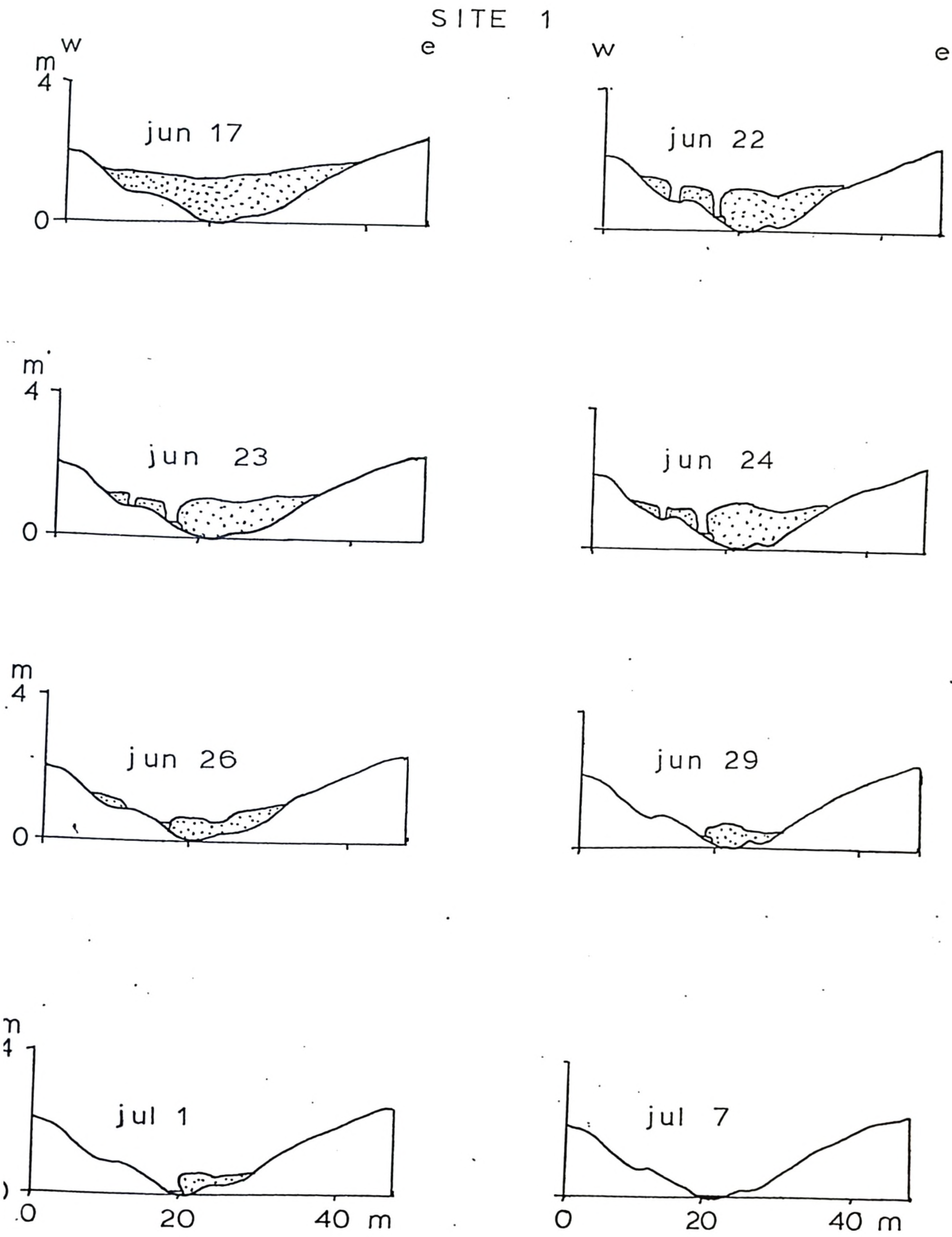


Figure 3.20 Example of flow diversion by capture, upstream of the tunnel reach at Site 1.

from one winter to another. This enables a crude prediction of snow conditions based on valley topography. It is hypothesised that given an initial snow cover condition in the valleys, the ensuing type of channel development processes can be predicted. A simple model was therefore formulated to predict qualitatively the sequence of channel development events.

The model requires an identification of the valley forms, a recognition of the valley snow distribution pattern, and an inference upon the corresponding types of channel development processes. St. Onge (1965) distinguished eight types of valley forms for Ellef Ringnes Island in the Canadian High Arctic. For the purpose of the present study, a simpler classification scheme is adopted. Based on aerial photography and field checking, four major types of valleys are identified, each possessing a distinctive topographic cross profile. These valley types include (Fig.3.21)

(1) non-incised valley with a narrow floor

(2) non-incised valley with a wide floor

(3) incised valley with a narrow floor

(4) incised valley with a wide floor

Each valley type is associated with a characteristic snow distribution due to a preferential snow drift accumulation pattern. Thus, non-incised valleys are more exposed and only a relatively thin snowpack accumulates on the floor. On the other hand, incised valleys retain a larger amount of snow. In many cases, the entire incised but narrow valley is snow-filled. For incised valleys with broad floors, snow does not accumulate uniformly across the valley and the lowest point on the

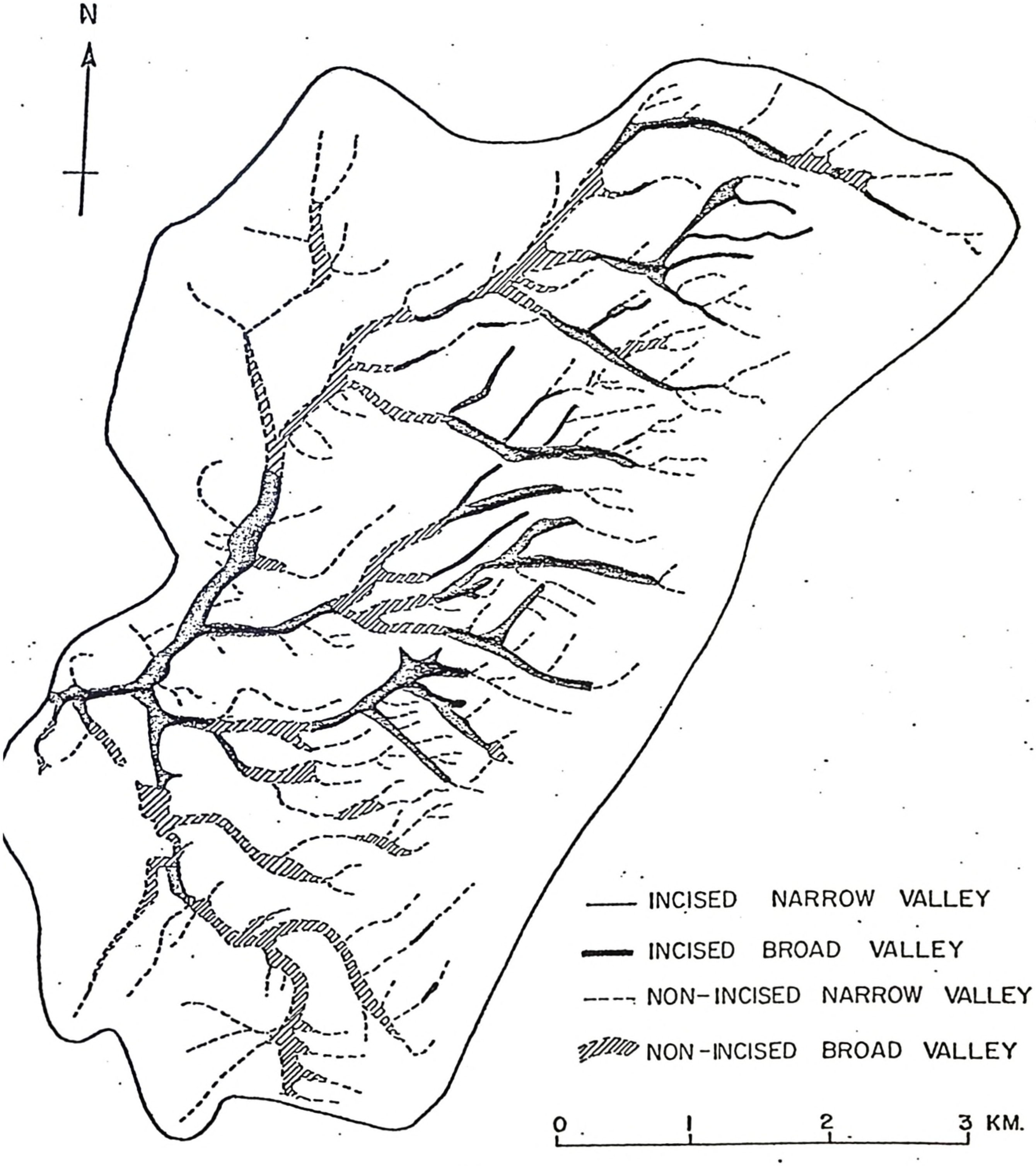


Figure 3.21 Map of the study area showing the distribution of four types of valleys each exhibiting distinctive channel

snow profile seldom corresponds with the lowest point at the valley bottom. Occasionally, transverse snow ridges span across the entire valley, enclosing troughs with a thin snowpack between the ridges.

Although it is difficult to predict the exact snow profile, the general form of the snow cover can often be deduced from topography and from prevailing winter wind directions. Together, the valley forms and the snow distribution pattern control the type of channel development sequence (Fig. 3.22).

3.2.1 Channel Development in Non-incised Valleys with Narrow Floors

These valleys usually hold a small amount of snow. Channel development begins with the cutting of miniature rills in the snowpack, followed by an enlargement to a full-size channel as snowmelt runoff increases. The channel soon reaches its clastic bed and the snowbanks are left to decay by melting.

3.2.2 Channel Development in Non-incised Valleys with Wide Floors

This category also holds a small amount of snow. However when meltwater begins to flow on the valley snowpack, local undulation in the snow surface affects the position of the incipient channel. If this incipient channel lies directly above the lowest point of the valley, vertical cutting brings the channel to the clastic bed or to a layer of bottom ice and the position of the channel is stabilised. Then the remaining valley snowpack is left to decay by melting. Sometimes, particularly at the headwater areas, there may be insufficient meltwater to enable complete vertical cutting through the snow. Then, all through summer, the channel will be flowing on a bottom ice layer or simply on

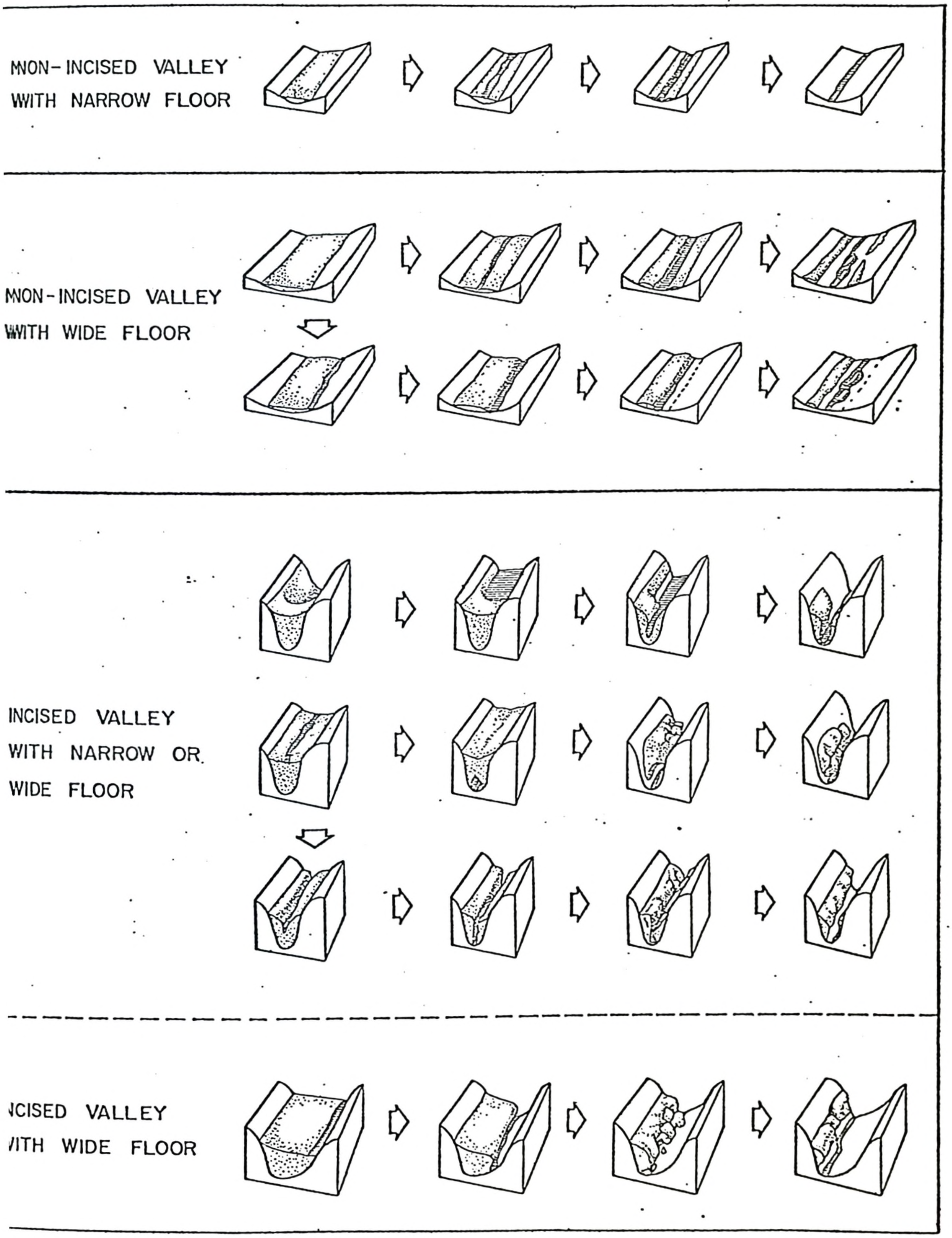


Figure 3.22 Block diagrams showing idealised sequences of channel development processes in the snowpacks of four types of valleys.

the snowpack and not reaching the clastic materials. Hence, in this case, the fact that a channel attains stabilisation is a function of the availability of water, hence basin area.

When the incipient channel lies to one side of the valley, there is a tendency of lateral shifting towards the lowest point of the valley floor. In so doing, one side of the valley floor becomes snow-free, but the other side will have an undercut snowbank which collapses as the channel migrates.

3.2.3 Channel Development in Incised Valleys with Narrow Floors

Incised valleys usually contain deep snowpacks except where local hollows occur in between transverse snow ridges. In spring these troughs are often ponded with meltwater. When the snow dams fail, the ponds drain rapidly and channels are then formed on beds of clastic materials.

Along sections where the hydrostatic water level in the snow falls beneath the surface, tunnelling results. When the tunnel roofs collapse, channels are observed winding through massive blocks of snow.

Along other sections uninterrupted downcutting into the snow soon establishes the channel on clastic materials, leaving tall snow-walls overhanging the stream. These snow-walls then decay by melting or by calving onto the channel.

3.2.4 Channel Development in Incised Valleys with Wide Floors

These valleys usually contain more snow than the other types of valley. Similar to the incised but narrow valleys, channel development can follow the ponding, tunnelling or the downcutting sequence. Added to these sequences is the possibility that the position where the channel is first established on clastic materials does not correspond with the lowest point of the valley. Lateral shifting then proceeds by undercutting the snow-walls and by the calving of large snow blocks. The extent of the ponded area is expected to be larger than that of narrow floor valleys.

3.2.5 Interruption of Channel Development Sequences

Channel development can be arrested by a diversion of flow upstream of a channel segment. Flow diversion is caused by the creation of an alternative channel which is more effective in conveying the flow. When this happens, the abandoned channel will be left to disintegrate as the snow melts.

3.3 Modelling of Channel Development Sequences

The previous section describes the various sequences in which a channel can develop in a snow-filled valley. Within a basin, snow drifting produces various snow disposition patterns as a result of spatial differences in valley shapes. Channel formation processes are therefore non-uniform across the basin. To overcome this problem, the valleys of the basin were regrouped into homogeneous reaches displaying similar snow-filled characteristics. Hence prediction of the prevalent type of channel opening in snow-filled valleys is possible through the use of a model designed for channel reaches of homogeneous snow-filled conditions.

3.3.1 Prediction

Figure 3.23 is a flow-chart summarising the channel development sequences given various terrain and snow cover characteristics. Following the flow-chart, streamflow does not begin until the snowpack is saturated. After flow commencement, the sequence of channel development differs according to the type of valley and according to the completeness of the valley snow cover. For non-incised valleys, channels stabilise quickly on beds of clastic materials if the valleys are narrow or if incipient channels are located directly above the lowest point of the valley. Otherwise, lateral shifting follows.

For incised valleys with irregular snow distribution, ponding occurs. After the ponds are drained, the channels stabilise if little snow remains in the valley. Otherwise, channel development proceeds by vertical cutting into the snow or by lateral channel shifting, depending on the position of the incipient channel formed after pond drainage.

In deep snowpacks, tunnelling may or may not take place, depending on the initial position of the hydrostatic water level. If this level is far below the snow surface, flow continues in the tunnel until the roof collapses. From then on, channel development will follow the same sequence as initiated by the downcutting of channels in snow.

The above predictive scheme does not consider flow diversion. The occurrence of flow diversion is difficult to predict, but only a few channel segments in the basin were thus affected.

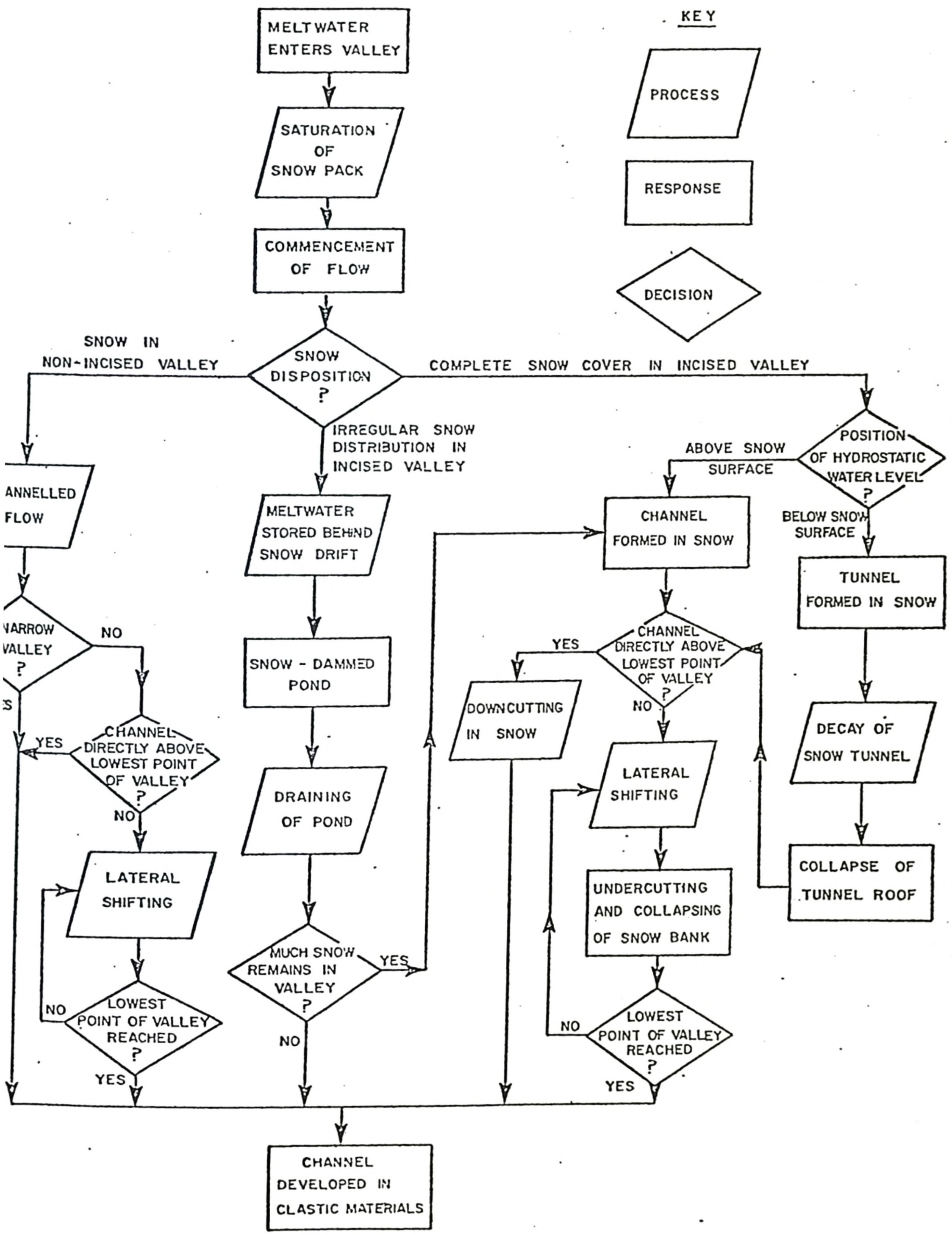


Figure 3.23 Flow-chart summarising the channel development sequences given various valley forms and snow distribution characteristics.

3.3.2. Discussion

Snow-filled valleys are not restricted to the High Arctic. In two temperate basins of Wyoming, U.S.A., Sturges (1975) observed over-snow runoff on the years that had a late spring. His basins are located in a moderately dissected plateau and vegetated by sagebrush. Strong winds blow in winter, with an average speed of 25 $km{\cdot}h^{-1}$ and exceeding 30 $km{\cdot}h^{-1}$ in December and January. The snow is wind-packed and in the valleys, snow depth reaches 6 m. From his study, it appears that oversnow flow requires deep, compacted snow which infills the winter-dry valleys. The spring should also arrive late so that uninterrupted melting is sustained while a deep valley snowpack still remains. These findings tend to complement the conclusions of our study.

Our study on channel development sequences adds complexity to the concept of variable source area for snowmelt runoff (Woo and Slaymaker 1975). In brief, this concept states that only parts of a basin contribute snow-melt runoff to the basin outlet. The areal extent of this source area changes with time according to the availability of meltwater. If all the meltwater input from the slopes can be conveyed downstream, the source area will include those zones with a melting snow cover. Our study shows that meltwater is temporarily stored in the valley snowpack upstream of the basin outlet. In terms of the outlet, this source of water is not available to runoff. Hence, those parts of the basin whose meltwater goes into storage are not yet contributing to streamflow. When this water is finally tapped by channels developed in the snow-filled valleys, the source area will expand abruptly. By then, however, the runoff event is no longer directly related to the snowmelt event because the flow was derived originally from a source area of the past period. Therefore, combining the channel development processes with the variable source area concept, it is possible to explain the discrepancy between snowmelt and streamflow discharge. Another water discharge characteristic associated with different process of channel development in snow-filled valleys is the manner in which meltwater flows over the snowpack. For periods ranging from hours to several weeks, very shallow sheet flow will move on the snow surface without incising it. These water discharges are of very low magnitude and it is only after an increase in discharge that stabilising processes may start.

The bursting of pond water has a more practical application since it deals with greater volumes of water. Downstream of a pond burst, a sharp and sudden increase in discharge is experienced. Although large volumes of water may be involved, "the ridges are seldom breached suddenly ... "and the" ... incision, although rapid, slows down somewhat the "Jökulhlaup effect" that the bursting of impounded water can achieve" (Pissart A. 1967, p. 221). Moreover the magnitude of the flood wave is reduced greatly, allowing the snowpack to attain its capacity level for free water storage before an initiation of flow on its surface.

The remaining discharge patterns simply follow the one of a nival regime stream (Church 1974), except for the minor stage irregularities produced by the channel undergoing lateral shifting.

From field observations and from the relation between the slope of snow-free valleys and snow disposition of the valley snowpack, the resulting channel opening sequences can be deduced.

CHAPTER FOUR

FLUVIAL PROCESSES IN SNOW-FILLED VALLEYS

Although an increasing amount of information has been obtained in the High Arctic basins pertaining to the fluvial activities, an outstanding problem remains regarding the role of the valley snowpack in influencing the geomorphic work of the streams. All researchers observed the prominence of snowmelt peak flow, but this flow often coincides with the period when the channels are still partially or totally covered with snow. McCann et al. (1972), for instance, were of the view that the duration of the valley snowpack was so short that little effect was produced to modify the fluvial activities. Pissart (1967), on the other hand, mentioned that 75 percent of the total annual flow of a small High Arctic stream took place when the channel floor was snow or ice covered. Thus, there exists conflicting thoughts on whether the valley snowpack exerts any significant effect on the fluvial processes. Both arguments were based on qualitative observations of streamflow activities, so that in order to decide on the erosional or protective roles of the valley snowpack, quantitative assessment of fluvial works during the snowmelt period is required.

With an increase in the magnitude of discharge, there is a general increase in depth, width, and velocity of flow. The force exerted by the running water on the streambed is also expected to increase, hence an increase in the potential for performing geomorphic work. If the streambed is covered by snow or ice, however, the bed roughness is reduced and the velocity increases. Those parts of the clastic bed which are protected by snow or ice will not be eroded, but the exposed segments will undergo erosion. It is therefore necessary to determine the changing hydraulic conditions as the channels develop in the snow-filled valleys.

Given a knowledge of the hydraulic variables, it is then possible to make use of some empirical equations to compute the amount of sediment that can be transported by the running water should the snow and ice barrier be absent. This computed rate of sediment transport provides the potential amount of work that the flow can perform. The actual amount of erosion work is expected to be reduced should the bed be partially or totally snow and ice covered. This latter amount represents the actual work performed by the stream. A difference between the potential and the actual amount of sediment transported will define the protective role played by the valley snowpack.

To further ascertain the role of snowpack on fluvial processes, field experiments were carried out to obtain evidence of spatial and temporal variations in erosional and depositional tendencies. Various depositional features were also examined to demonstrate the influence of the valley snowpack.

Using the several sets of information thus obtained, it is hoped that some light could be shed upon the problem of how important a role the valley snowpack plays in terms of the fluvial activities of High Arctic streams.

4.1 Hydraulics of Stabilising Channels

4.1.1 Hydraulic Considerations

"Gravity exerts a force which propels the water downslope; friction between the water and the channel boundaries tends to resist the downslope movement." (Leopold et al. 1967, p. 152).

Flow in a stream channel is confined by bed and banks which act as a boundary, this boundary generates resistance to the flow of water. A force is applied to the wetted perimeter of the channel and in uniform flow, this force can be obtained by

$$\tau = \rho\ g\ R\ S \tag{4.1}$$

where τ is the shear stress per unit area of bed $[M\ L^{-1}\ T^{-2}]$

ρ is the water density $[M\ L^{-3}]$

g is the acceleration due to gravity $[L\ T^{-2}]$

R is the hydraulic radius $\approx d$ mean flow depth $[L]$

S is the hydraulic gradient $[0]$

The boundary on which this shear stress is applied, generates a friction drag to the flow. A flow resistance equation can then be derived by balancing the resisting shear force (shear strength) at the channel boundary against the force propelling the flow (Leopold et al. 1964).

A set of theoretical equations has been developed to describe the flow resistance, one of which employes the dimensionless Darcy-Weisbach resistance coefficient ff :

$$ff = \frac{V^{*2}}{V^2} \tag{4.2}$$

where V^* is the shear velocity $= \sqrt{\tau/\rho} = \sqrt{g\ R\ S}$ $[L\ T^{-1}]$

V is the flow velocity $[L\ T^{-1}]$.

Another empirical coefficient is obtain from the Manning's equation which takes the general form of

$$V = C\ R^x\ S^y \tag{4.3}$$

where x and y are the numerical coefficients [0]

C is a flow resistance factor [0]

V, R, S are already defined.

From this relation, a coefficient of total flow resistance is defined as

$$n = R^{2/3} S^{1/2} V^{-1} \tag{4.4}$$

Manning's n provides an assessment of "the manner in which the boundary shear exerts its influence on velocity" (Leopold et al. 1969, p. 160). This measure of total flow resistance is also an indication of the relative bed roughness.

Hence, based on the three variables of R, S and V both the shear stress applied to the bed and the total resistance of the flow can be computed. The shear stress and the flow resistance indicate the hydraulic characteristics of a given flow. Variations through time and space of these hydraulics characteristics will induce changes which affect sediment entrainment and deposition (Church 1975).

4.1.2 Computation of Hydraulic Parameters

In the field, the three basic variables (i.e. R, S, V) were determined. R and V are components of the velocity-area discharge measurement and S was surveyed between two fixed points along the channel, their time trends and their relations with water discharge are then analysed over the flow season.

While channels are acquiring stability by overcoming the snow-filled conditions, the different components of the continuity equation (4.5) change frequently with time.

$$Q = W\,d\,V \tag{4.5}$$

where Q is the water discharge [$L^3\ T^{-1}$]

W is the channel width [L]

d is the mean channel depth $\approx$ R [L]

V is the mean water velocity [$L\ T^{-1}$]

Each component on the right-hand-side of equation 4.5 can be expressed as a power function of the water discharge as follows:

$$W = a\,Q^b \tag{4.6}$$

$$d = c\,Q^f \tag{4.7}$$

$$V = k\,Q^m \tag{4.8}$$

where a, c, k, b, f, m, are numerical constants.

where

$$a\,c\,k = 1.0 \tag{4.9}$$

and

$$b + f + m = 1.0 \tag{4.10}$$

These relationships are called the hydraulic geometry and they allow a computation of the hydraulic radius, the channel width and mean flow velocity given the discharge.

The at-a-station hydraulic geometry was examined using the discharge values obtained at sites 1, 2, 3 and 4 (Figs. 4.1, 2, 3, 4 respectively). Various relationships were obtained between discharge, the hydraulic radius

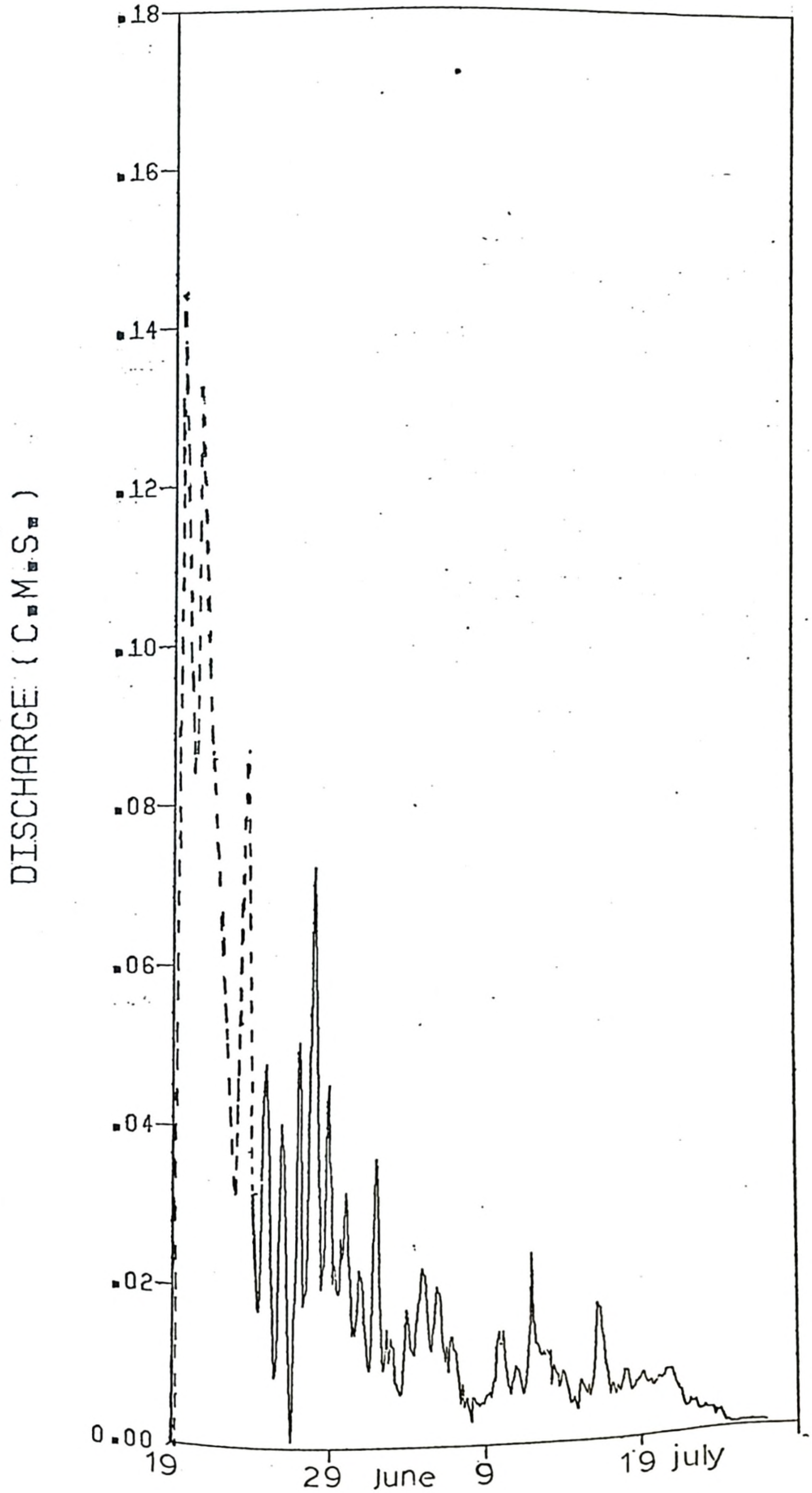


Figure 4.1 Hydrograph of Site 1.

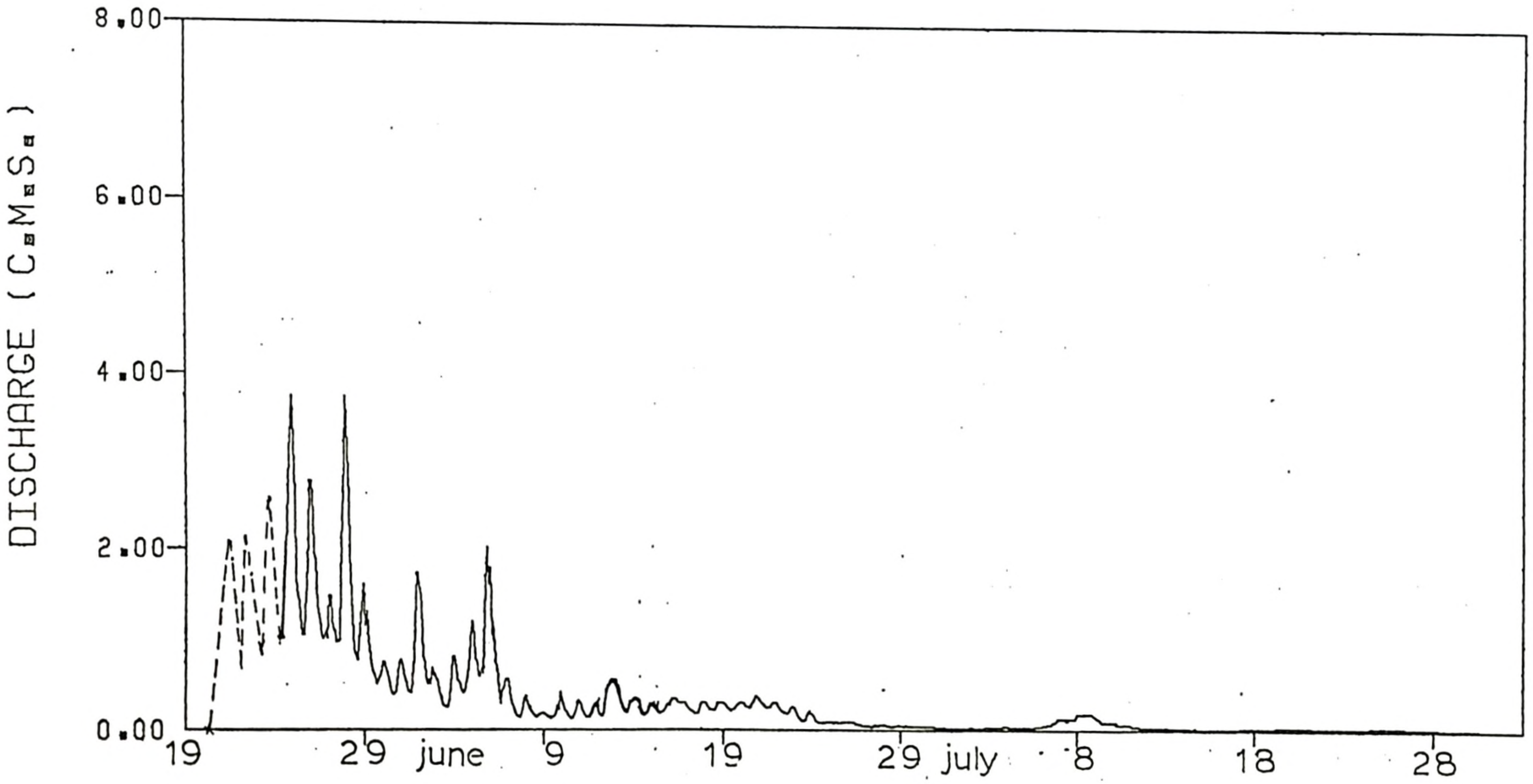


Figure 4.2 Hydrograph of Site 2.

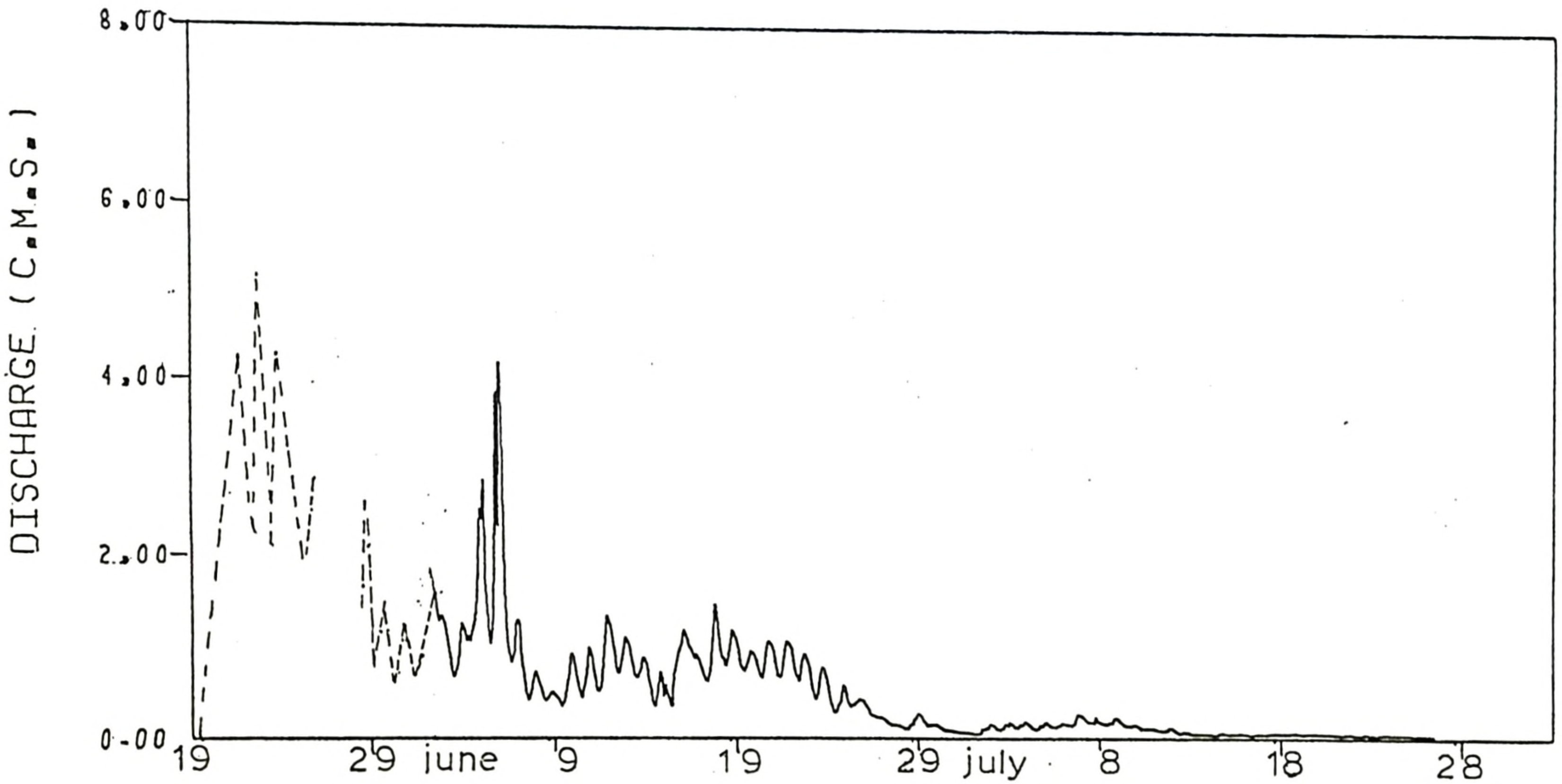


Figure 4.3 Hydrograph of Site 3.

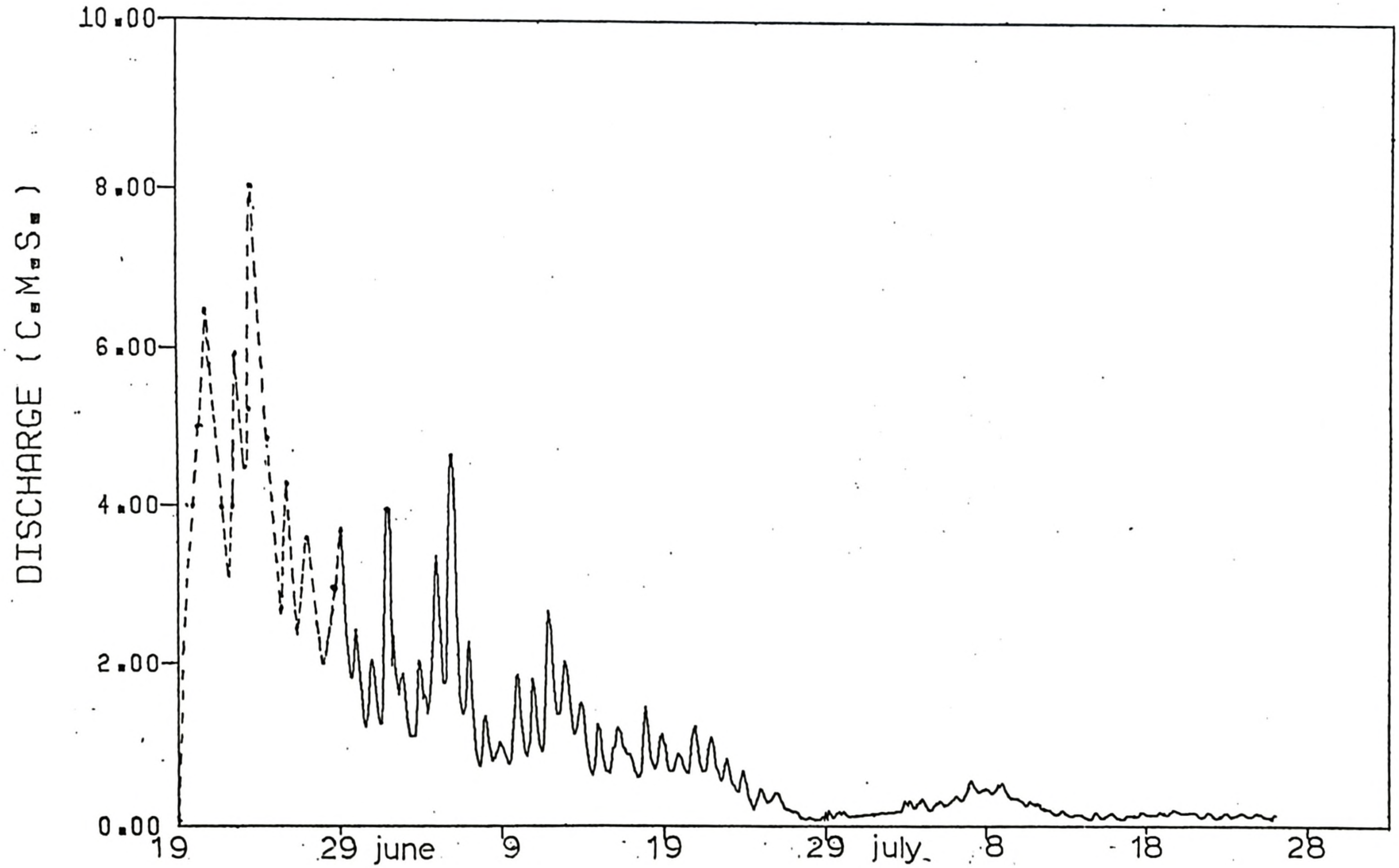


Figure 4.4 Hydrograph of Site 4.

(approximated by mean channel depth) (Figs. 4.5, 6, 7, 8), the channel width (Figs. 4.9, 10, 11, 12) and the flow velocity (Figs. 4.13, 14, 15, 16). The at-a-station hydraulic geometry graphs show considerable scattering of the data points. Such a scatter around the lines of log-log relationships is partly attributable to field measurements of channel width, depth and flow velocity taken at slightly different sections in the vicinity of gauging stations. However when the hydraulic geometry coeficients are compared with those obtained in other environments, the exponents for our data fall within the range of reported values (Table 4.1). A substantial difference from the exponent for Brandywine Creek is possibly related to its cohesive banks. The high values of "k" and "m" at Site 1 are probably due to its steeper channel slope.

It should be noted that, for the velocity-discharge relationship, empirical equations from the study basin underestimate low flow velocities during the first week of streamflow, but over estimate them during the recession flow period (Fig. 4.16). The higher flow velocity for a given low discharge is a remarkable characteristic of flow in channels with snow and ice beds. This illustrates that flow resistance is reduced when a smoother boundary was provided by the snow or the bottom ice.

A seasonal change in Manning's n value at Site 4 partially confirms an overall increase of flow resistance over the flow period (Fig. 4.17). For the other sites, the change is less apparent because the bottom ice disappeared at an early date. In the case of the hydraulic gradient, there was also an increase in the water surface slope as the amount of snow and bottom ice gradualy decreased in June at Site 4.

The hydraulic geometry relationships were used to obtain daily

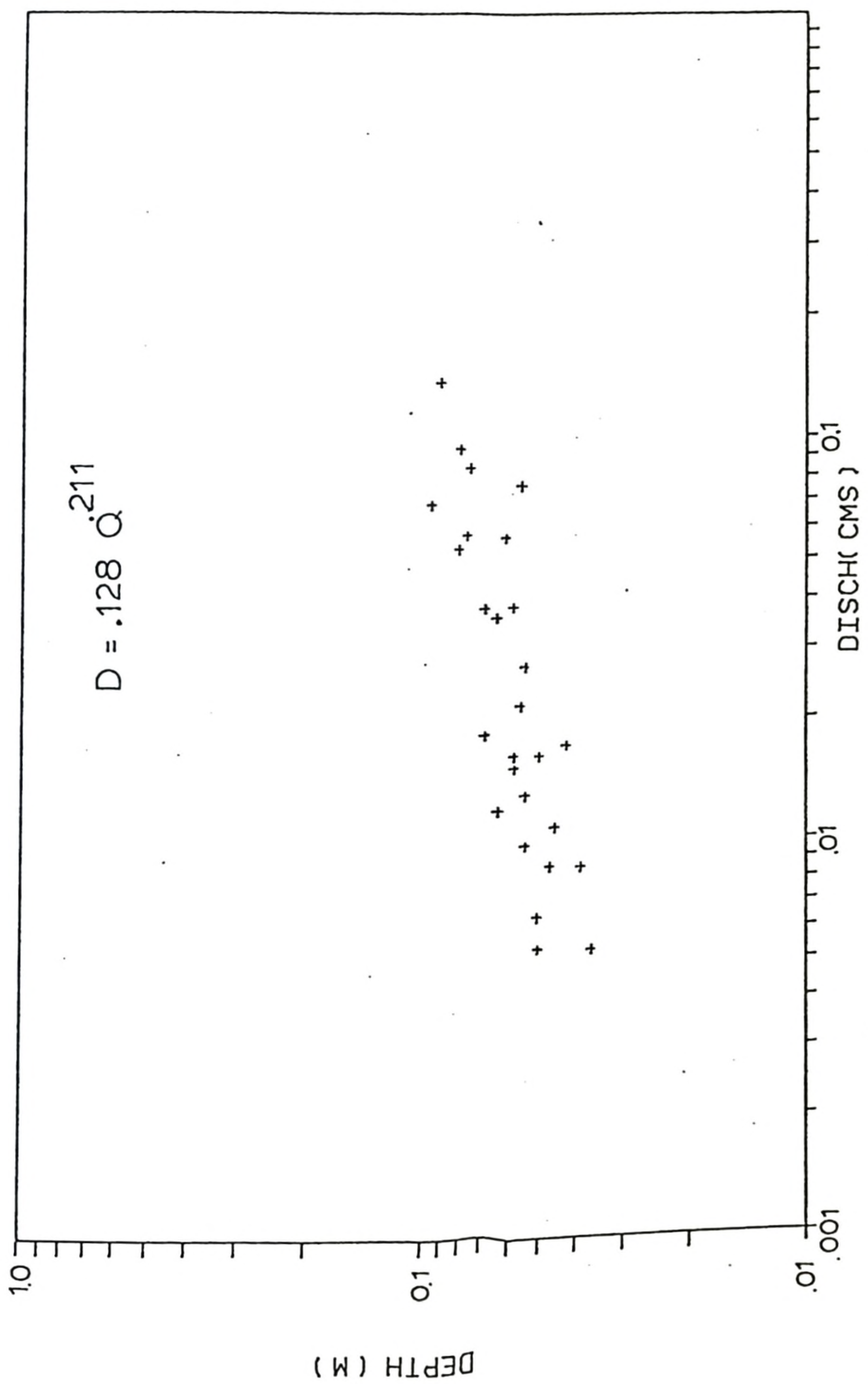


Figure 4.5 Depth-discharge relationship, Site 1.

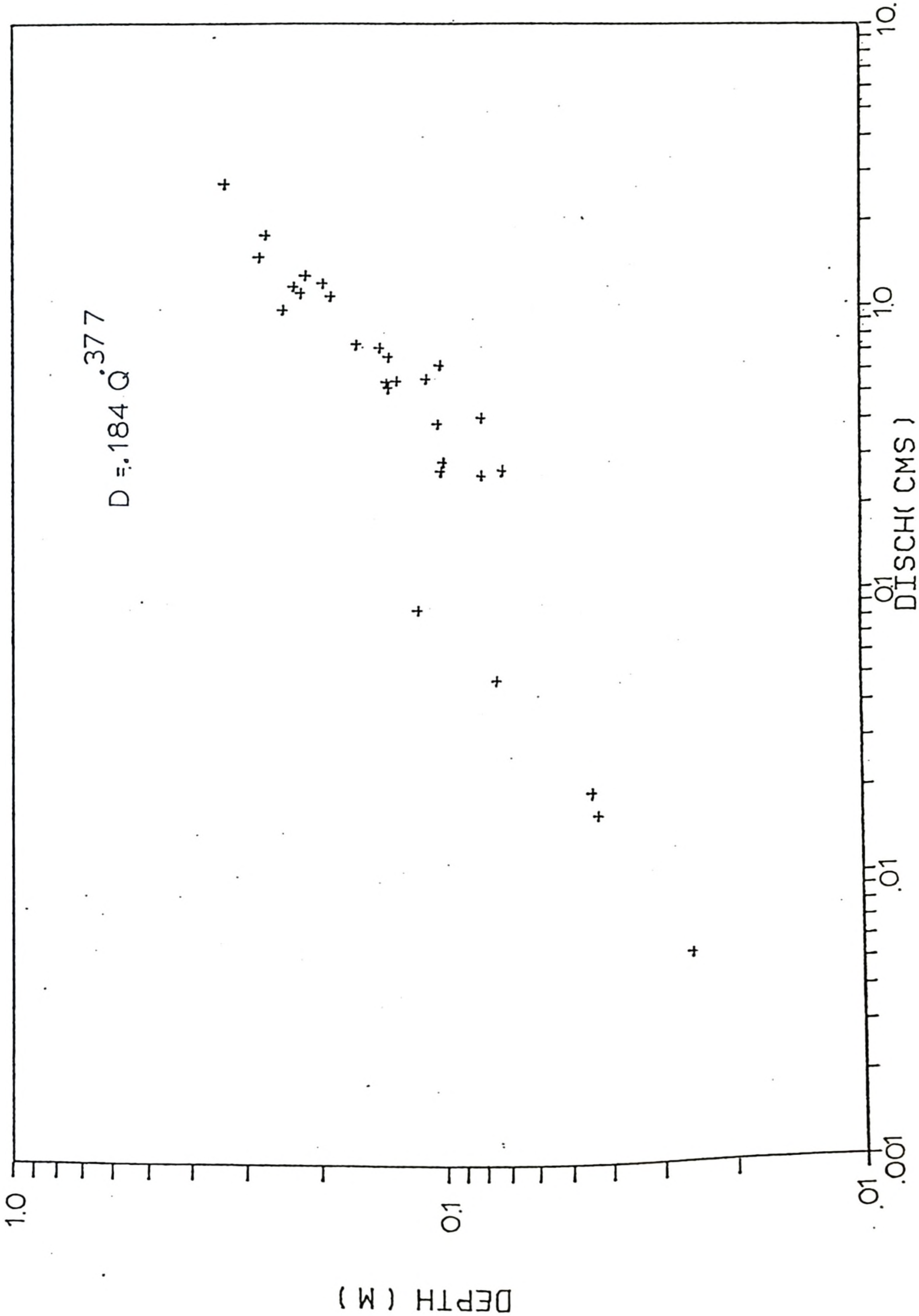


Figure 4.6 Depth-discharge relationship, Site 2.

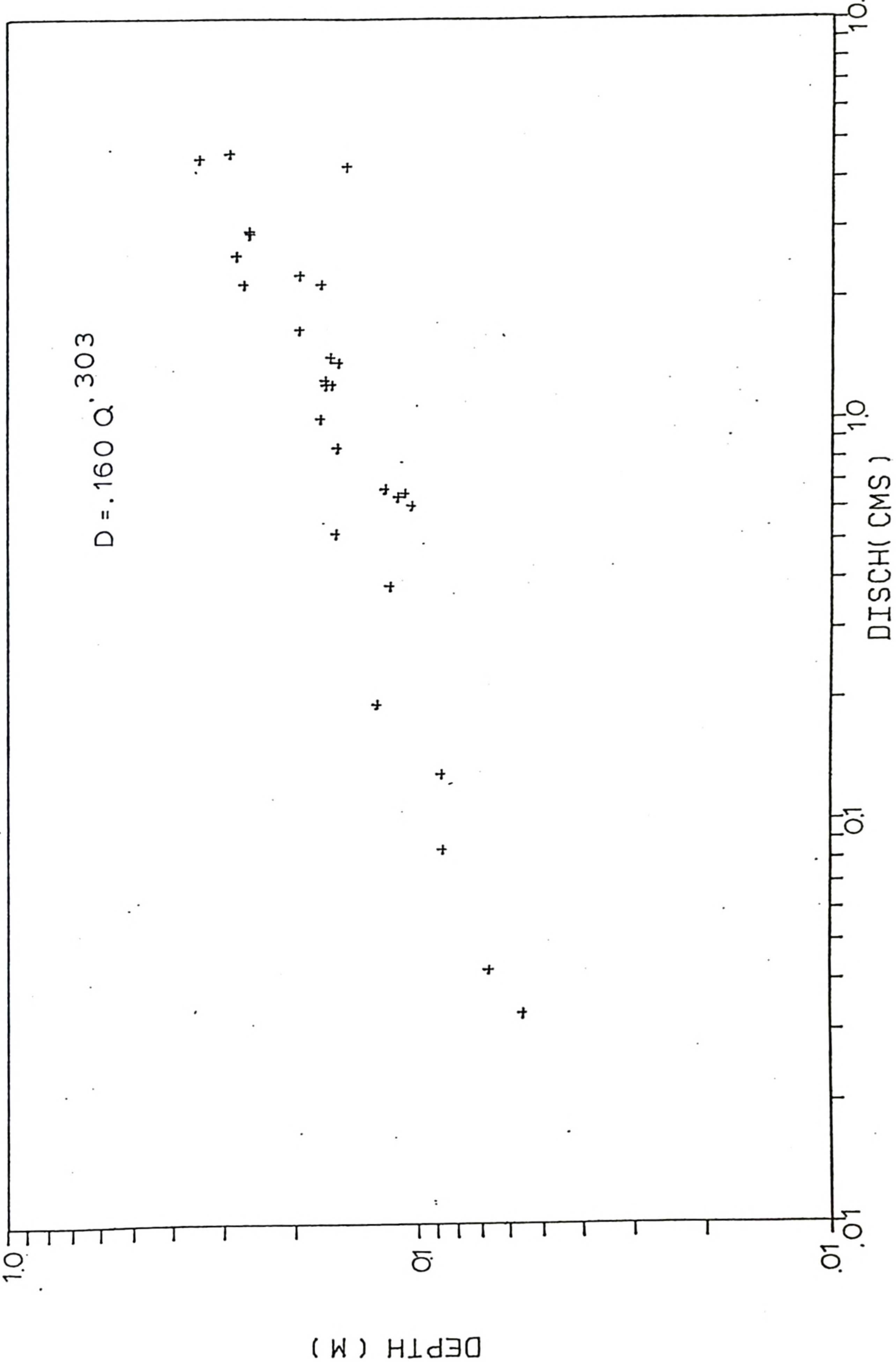


Figure 4.7 Depth-discharge relationship, Site 3.

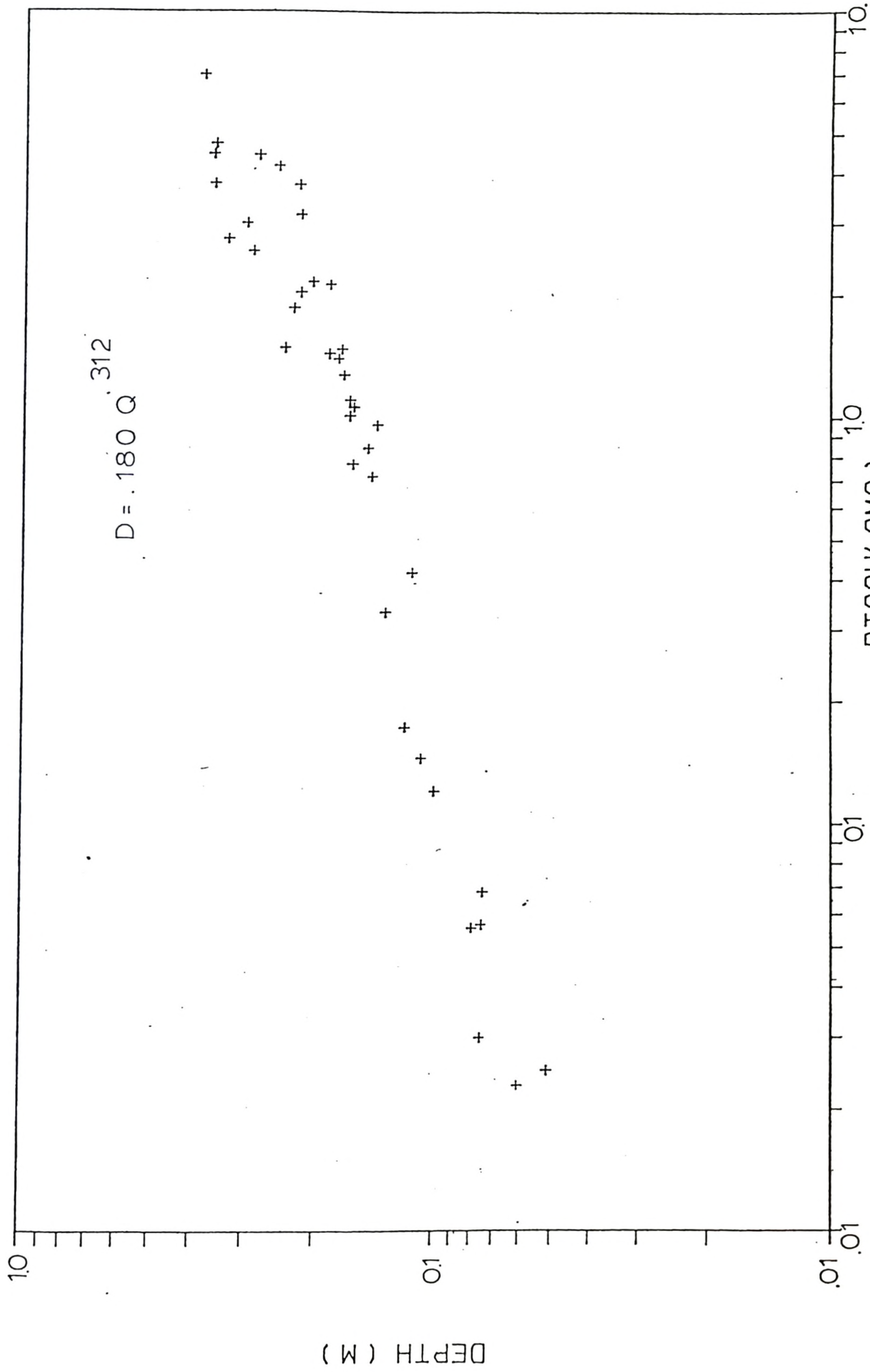


Figure 4.8   Depth-discharge relationship, Site 4.

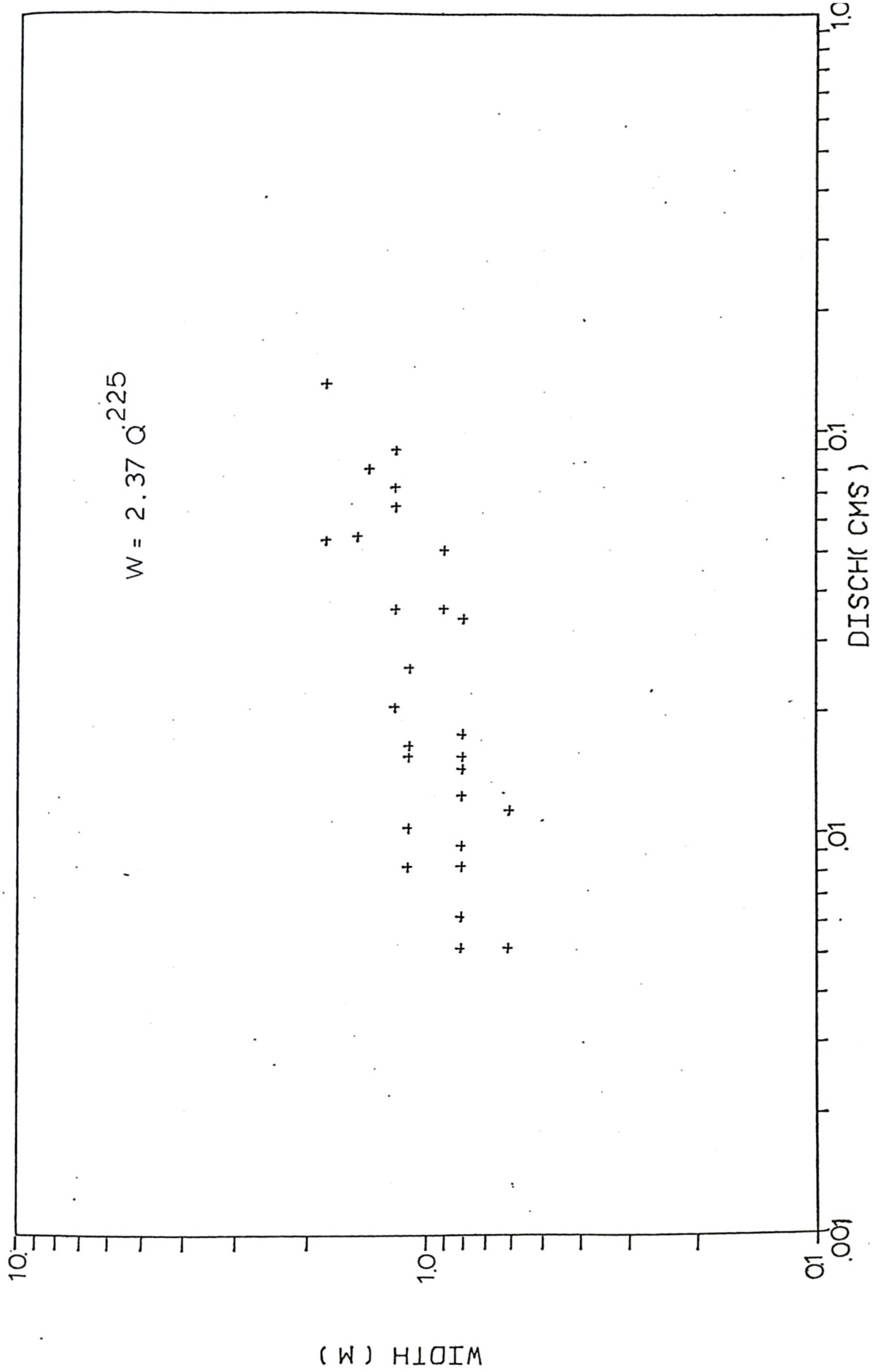


Figure 4.9 Width-dis charge relationship, Site 1.

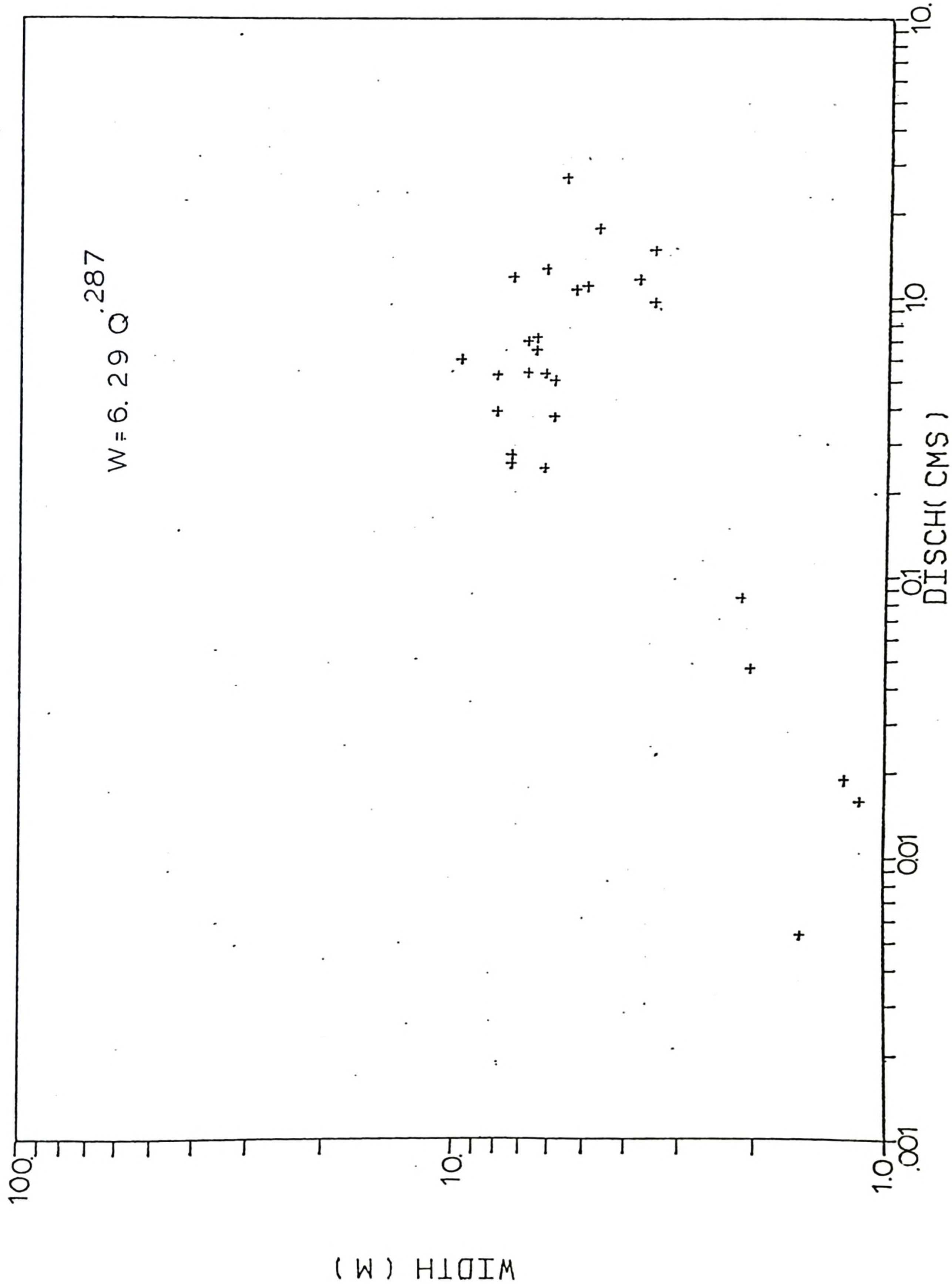


Figure 4.10   Width-discharge relationship, Site 2.

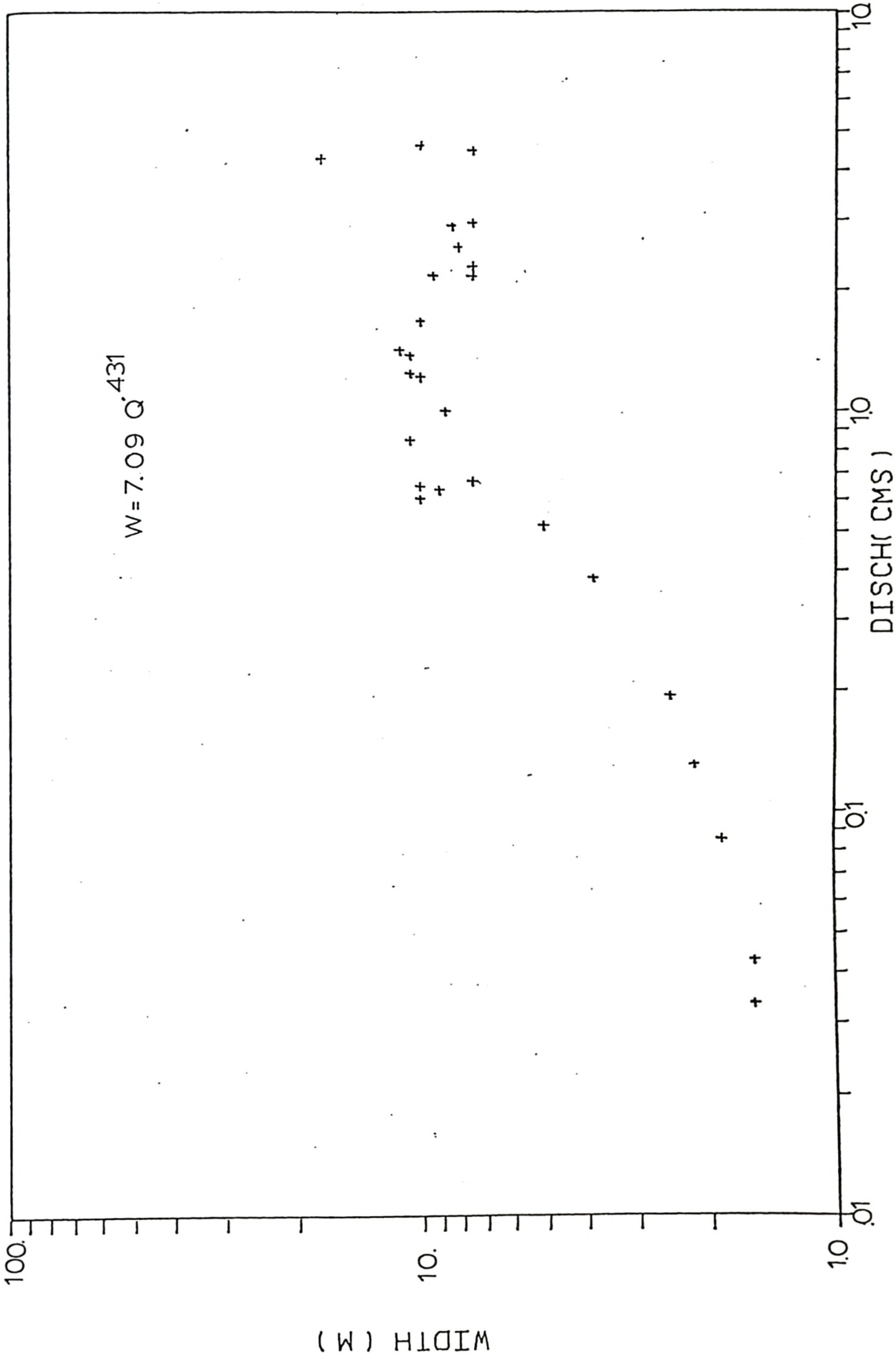


Figure 4.11 Width-discharge relationship, Site 3.

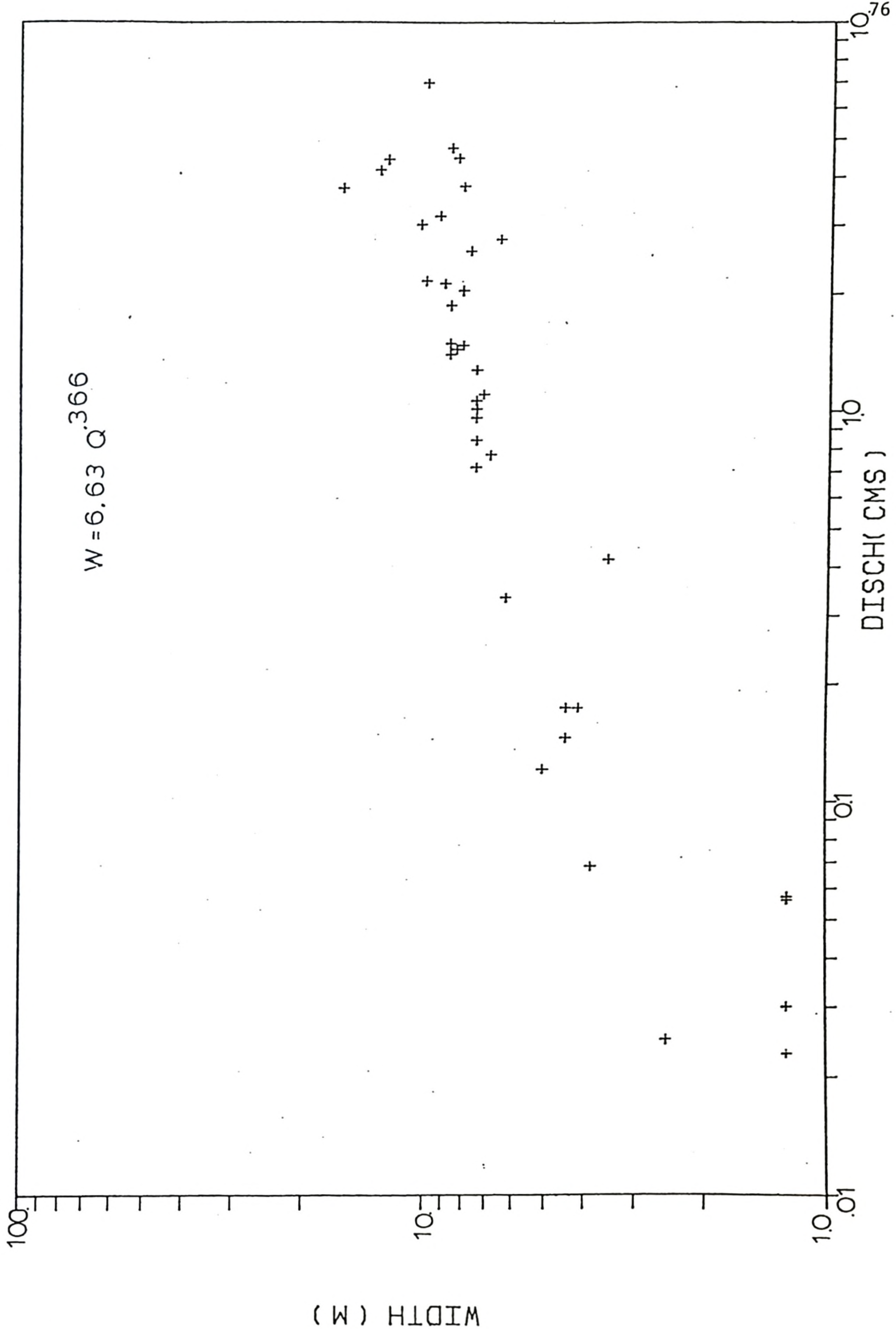


Figure 4.12   Width-discharge relationship, Site 4.

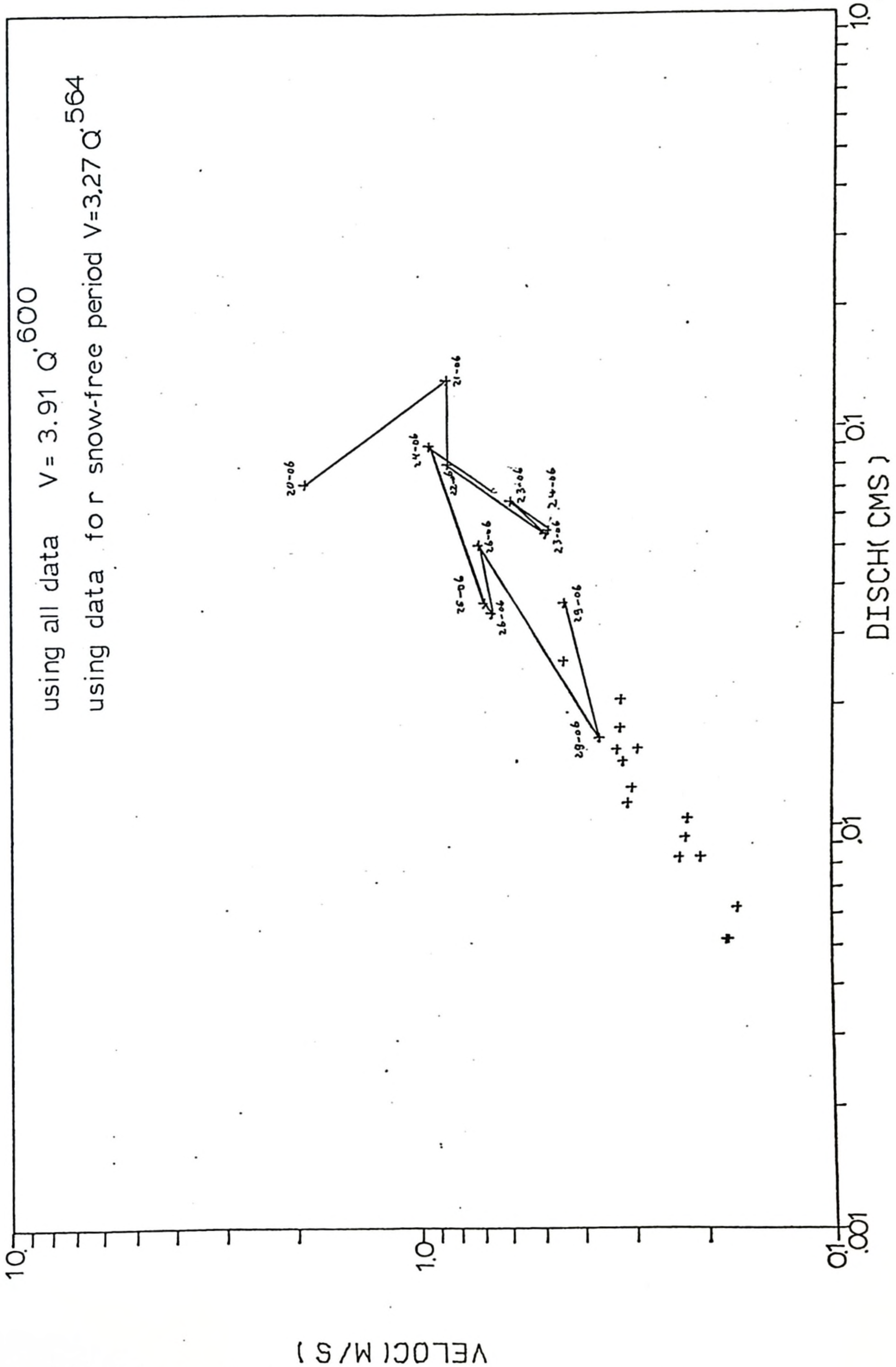


Figure 4.13 Velocity-discharge relationship, Site 1. Line segments join those points representing

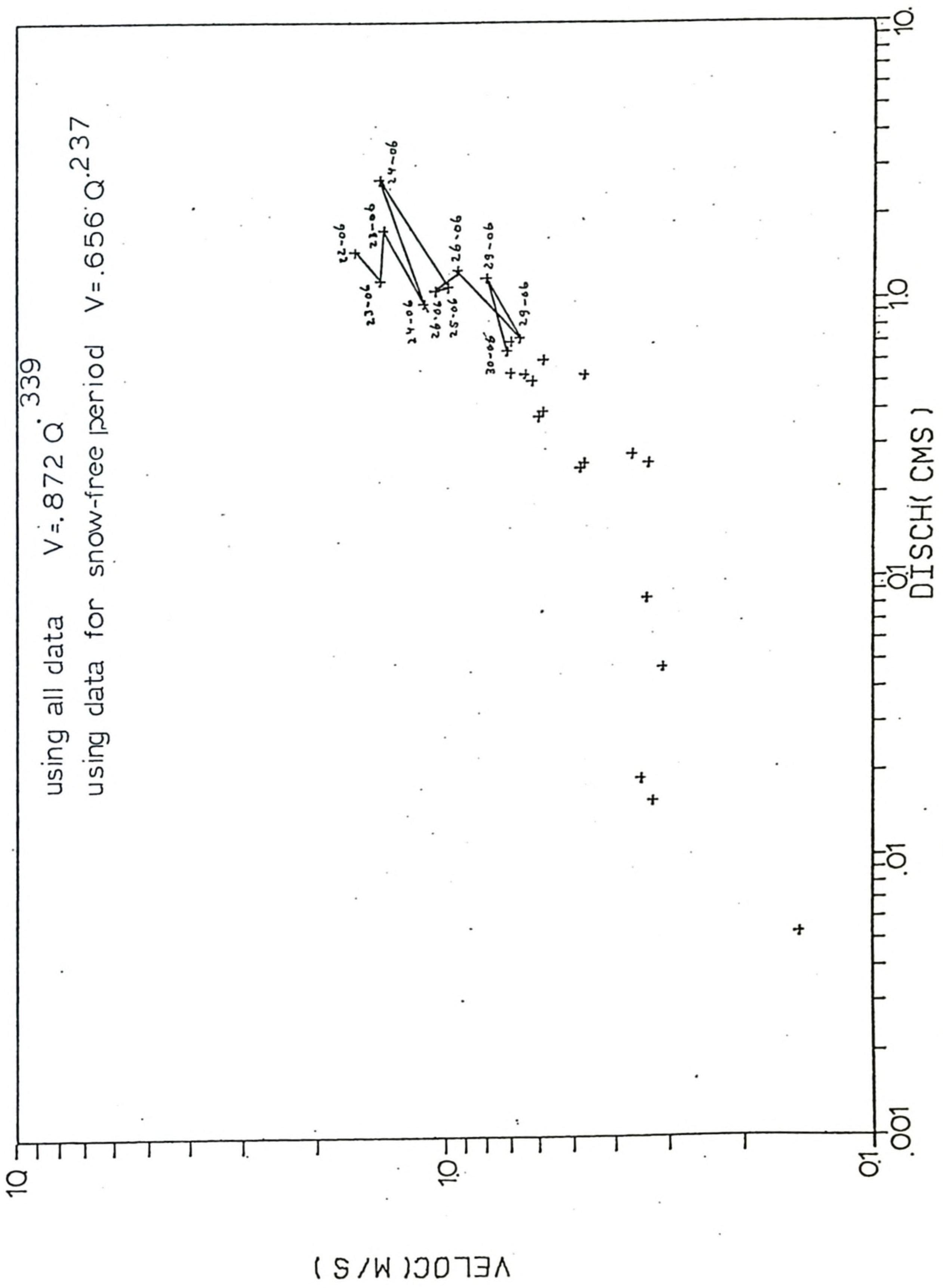


Figure 4.14 Velocity-discharge relationship, Site 2. Line segments join those points representing snow or bottom ice streambed conditions.

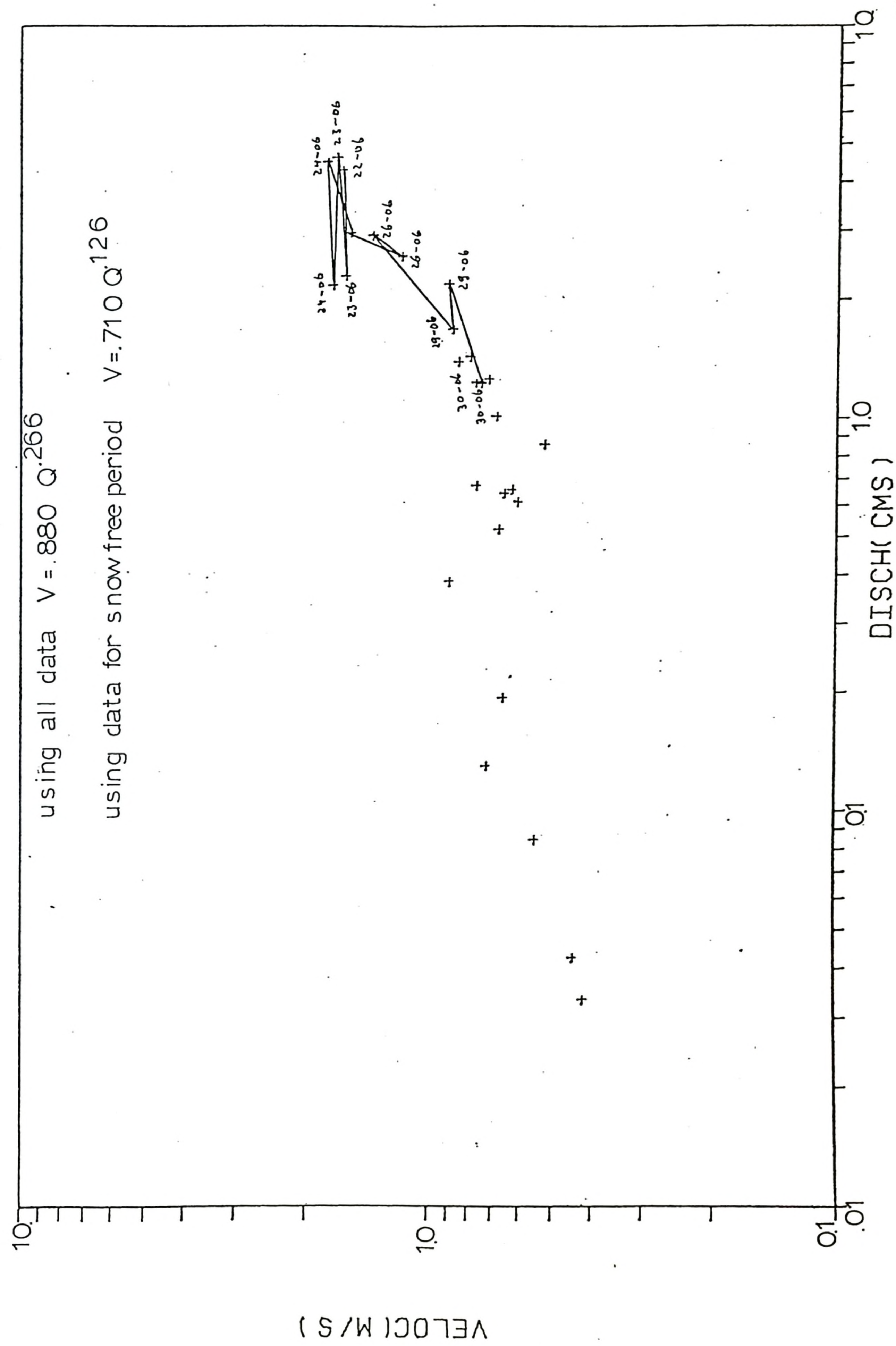


Figure 4.15 Velocity-discharge relationship, Site 3. Line segments join those points representing snow or bottom ice streambed conditions

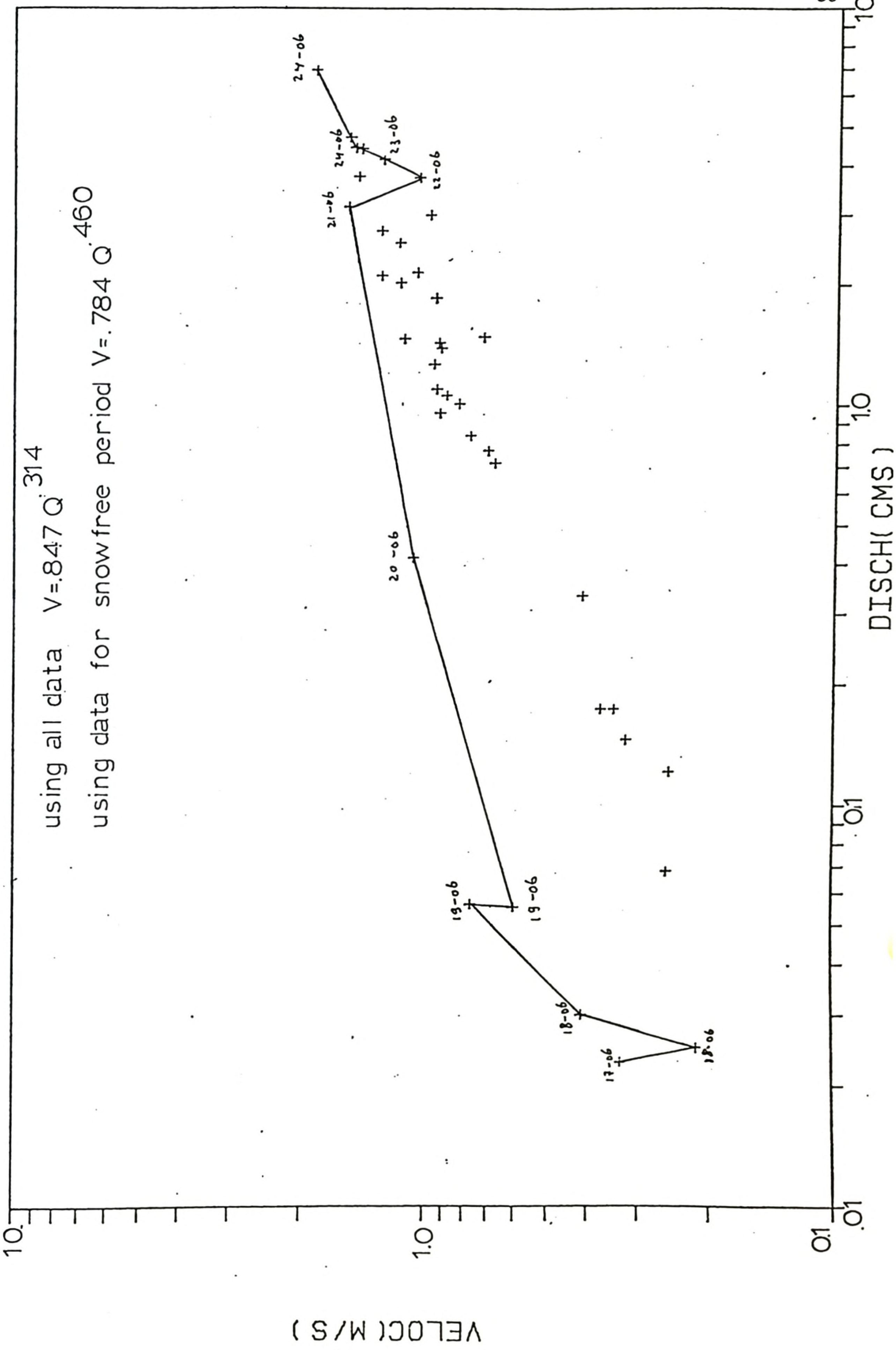


Figure 4.16 Velocity-discharge relationship, Site 4. Line segments join those points representing snow or bottom ice streambed conditions.

TABLE 4.1

Channel Parameters of Hydraulic Geometry of Unnamed Stream, Resolute at Site 1,2,3,4, Compared with Other Streams. (values "at-a-station")

Stream	Intercept			Exponents		
	a	c	k	b	f	m
White River[1]....	4.0	0.22	1.10	0.38	0.38	0.27
Southwest and Great Plains[2]						
Median....	26	0.15	0.37	0.26	0.40	0.34
Range.....	-	-	-	(0.03-0.59)	(0.06-0.63)	(0.07-0.55)
Brandywine Creek[3]						
Median....	54	0.23	0.10	0.04	0.41	0.55
Range.....	(37 - 80)	(0.12-0.52)	(0.022-0.16)	(0.00-0.08)	(0.32-0.46)	(0.48-0.69)
Ephemeral Streams						
Median[4]...	10.	0.1	1.0	0.26	0.33	0.32
Range.....	(3 - 26)	(0.03-0.2)	(0.700-1.50)	-	-	-
Flume Channels						
D_{50} 0.0022ft[5a]....	4.2	0.18	1.3	0.50	0.39	0.16
D_{50} 0.0066ft[5b]....	2.0	0.27	1.8	0.33	0.52	0.16
Unnamed Stream Resolute[6]						
Site 1....	1.07(2.37)	0.20(0.13)	1.51(3.91)	0.23(0.23)	0.21(0.21)	0.60(0.60)
Site 2....	2.27(6.29)	0.16(0.18)	0.85(0.87)	0.29(0.29)	0.38(0.38)	0.34(0.34)
Site 3....	1.52(7.09)	0.18(0.16)	1.12(0.88)	0.43(0.43)	0.30(0.30)	0.27(0.27)
Site 4....	1.80(6.63)	0.20(0.18)	0.91(0.85)	0.37(0.37)	0.31(0.31)	0.31(0.31)

where $W = a\,Q^b$; $d = C\,Q^B$; $V = k\,Q^m$.

[1]White River, Fahnestock (1963), Sandy gravel bed and banks, no vegetation Q = 0.01-400 cfs., 112 stations.
[2]Leopold and Maddock (1953): Streams of Southwestern States and Great Plains bed and bank material not noted, 200 stations.
[3]Wolman (1955): Brandywine Creek, Cohesive banks, gravel and sand bed, little snow or fill, Q = 5-2000 cfs, 7 statio
[4]Leopold and Miller (1956), ephemeral streams, streams, sandy bed, silty clay banks, Q = 0.15-800 cfs. 9 stations.
[5]Wolman and Brush (1961): flume channels, uniform sand in bed and banks, 5a = 29 channels; 5b = 13 channels.
[6]Sauriol (1977), unnamed stream, Resolute, Arctic nival regime, coarse sand and gravel bed, for Q = cfs; ( ) for Q=cm

(where 1 to 5 are found in Fahnestock 1963) mean stream depth based on discharge values. Combined with the hydraulic gradient observations, the shear stress exerted by the flow on the bed was then computed. Plotted against time (Fig. 4.17), variations in daily shear stress on the bed are shown to peak in the first half of the flow season but then taper off gradually as summer advanced. The computed shear stress was compared with discharge (Nordin and Beverage 1965), and linear relationships were obtained (Fig. 4.18).

From these daily values of the hydraulic variables, it is possible to compute daily bed material load. Before such computations are presented, the following section describes the various geomorphic features which indicate the type and the magnitude of fluvial activities which occurred in the snow-filled valleys.

4.2 Fluvial Activity in Snow-filled Valley

"The ability of flowing water to carve a channel, transport debris, and thus ultimately to degrade the landscape, depends on these forces - the gravitational impelling force, and the resistances offered to it" (Leopold et al. 1964, p. 153).

The presence of a snowpack in the valleys disturbs the normal course of water drainage hence the fluvial processes. In this section, the geomorphic evidence demonstrating the effects of valley snowpacks will be presented.

4.2.1 Morphological Evidences

Valleys infilled with a large amount of snow drift often experience the processes of stream downcutting, lateral shifting and tunnelling in the

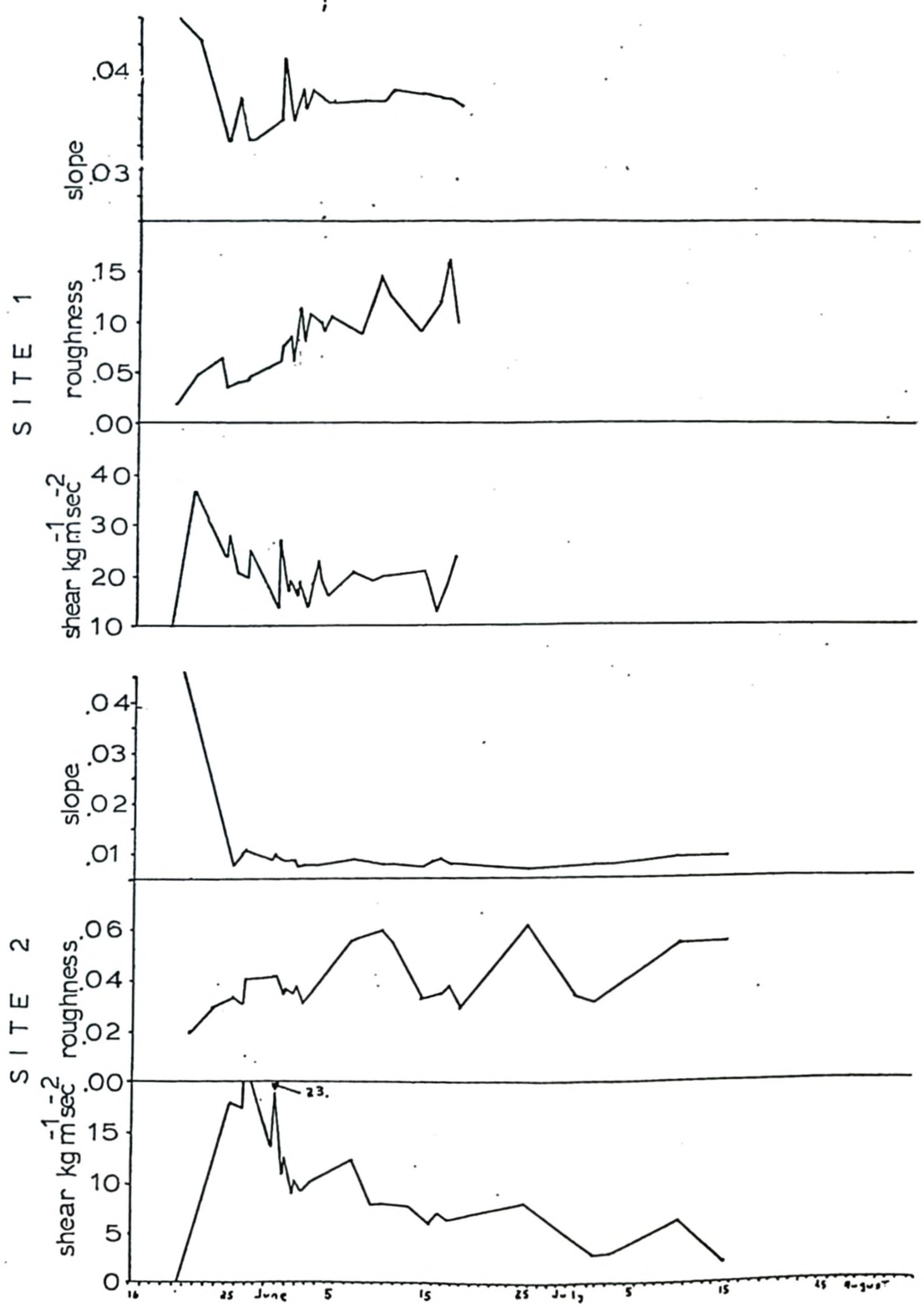
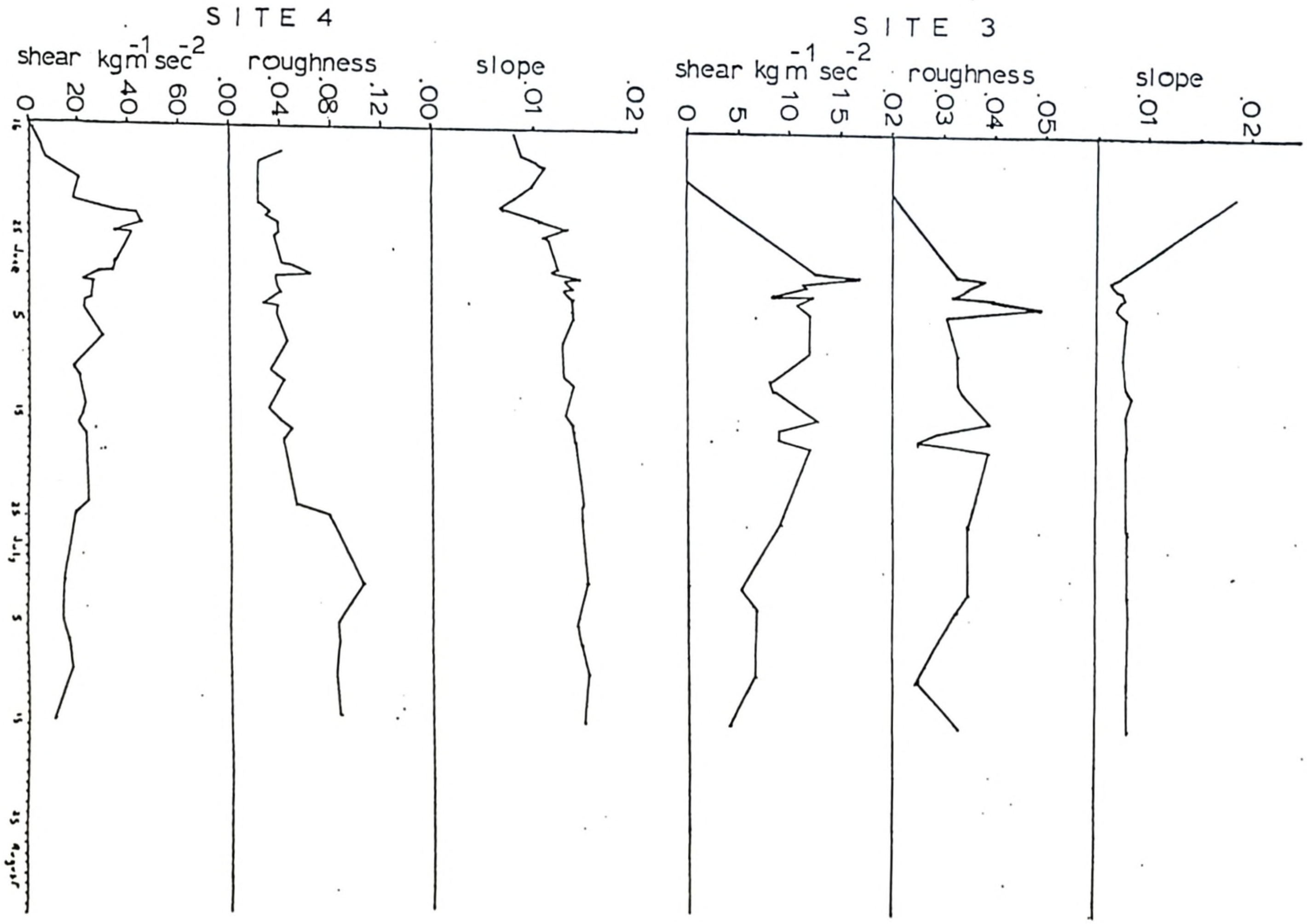


Figure 4.17 Variation of hydraulic data of the streams at Sites 1, 2, 3, and 4.



SITE 4
shear kg m-1 sec-2
0 20 40 60
roughness
.00 .04 .08 .12
slope
.00 .01 .02
25 June
5
15
25 July
5
15
25 August
SITE 3
shear kg m-1 sec-2
0 5 10 15
roughness
.02 .03 .04 .05
slope
.01 .02

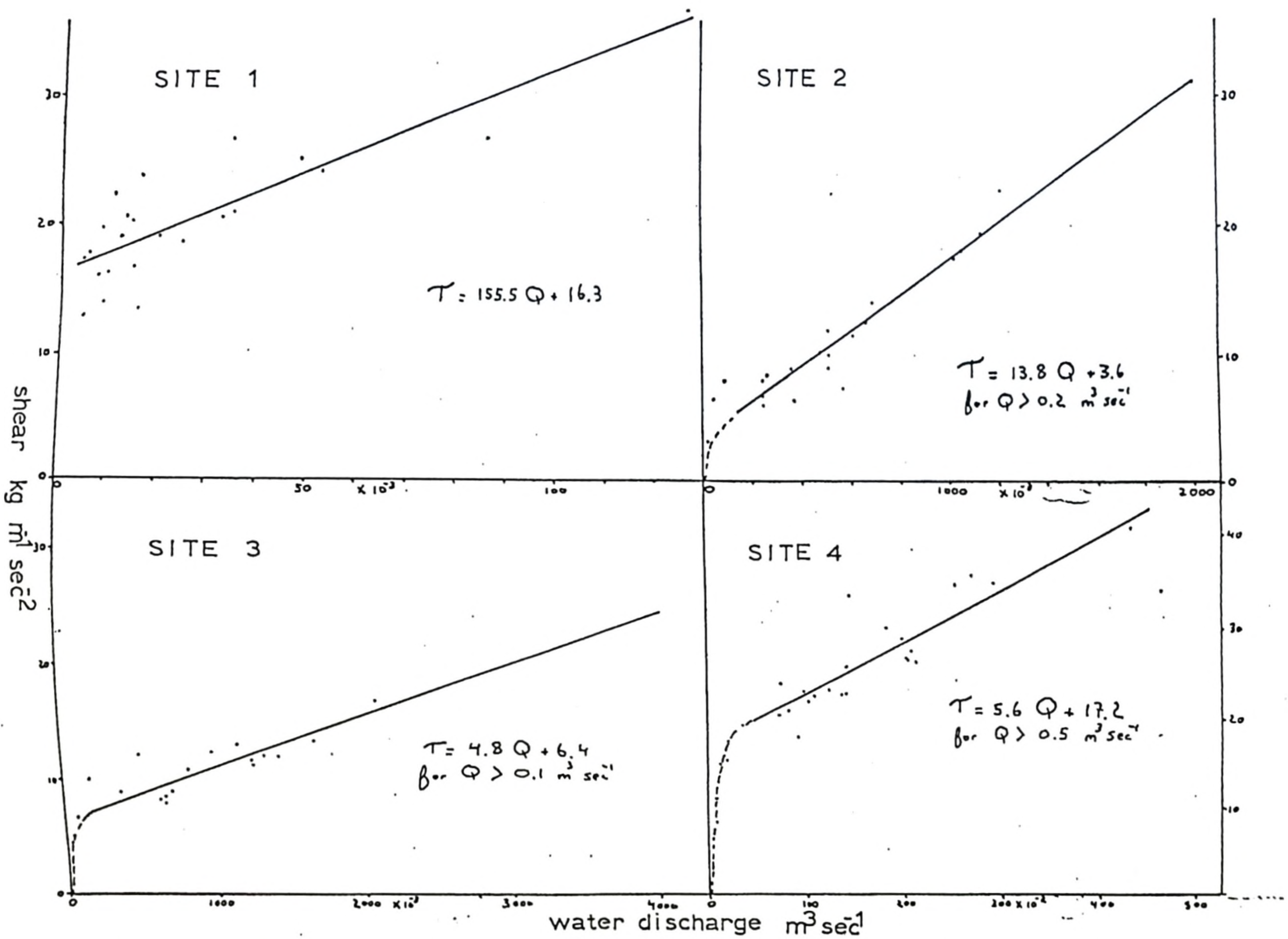


Figure 4.18 Relationship between shear stress and water discharge at Sites 1, 2, 3, and 4.

snow. In certain areas such as the tributary valleys at the upper parts of the basin, discharge is low and there is insufficient energy to incise the snowpack. In this case, channel stabilisation will be mainly accomplished by the melting of the pack. On the other hand, where discharge is considerable, the streams erode the snowpack to reach the bed. This produces various morphological features in the valley which reflect the role played by the snow in buffering the streambed against the flow.

4.2.1.1 Sources of Material

When the streambed is protected by the snowpack, no geomorphic work occurs at the site. Along this reach, any material carried by the running water is material in transit. The source of the material in transit is mainly derived from upstream reaches where the bed is cleared of snow. Another source of sediment is from the banks where the snow disappears earlier than at the valley bottom. A third possible source of sediment, mainly fines, comes from the slopes and the sediment is carried to the stream channel via a miniature drainage network developed on the slopes.

Although not measured because of their instantaneous occurrence, dense clouds of suspended sediment were observed during the first few weeks of the flow season. This phenomenum is possibly related to sudden releases of water temporarily ponded behind collapsed snowblocks associated with channel shifting at some upstream reaches (Fig. 4.19).

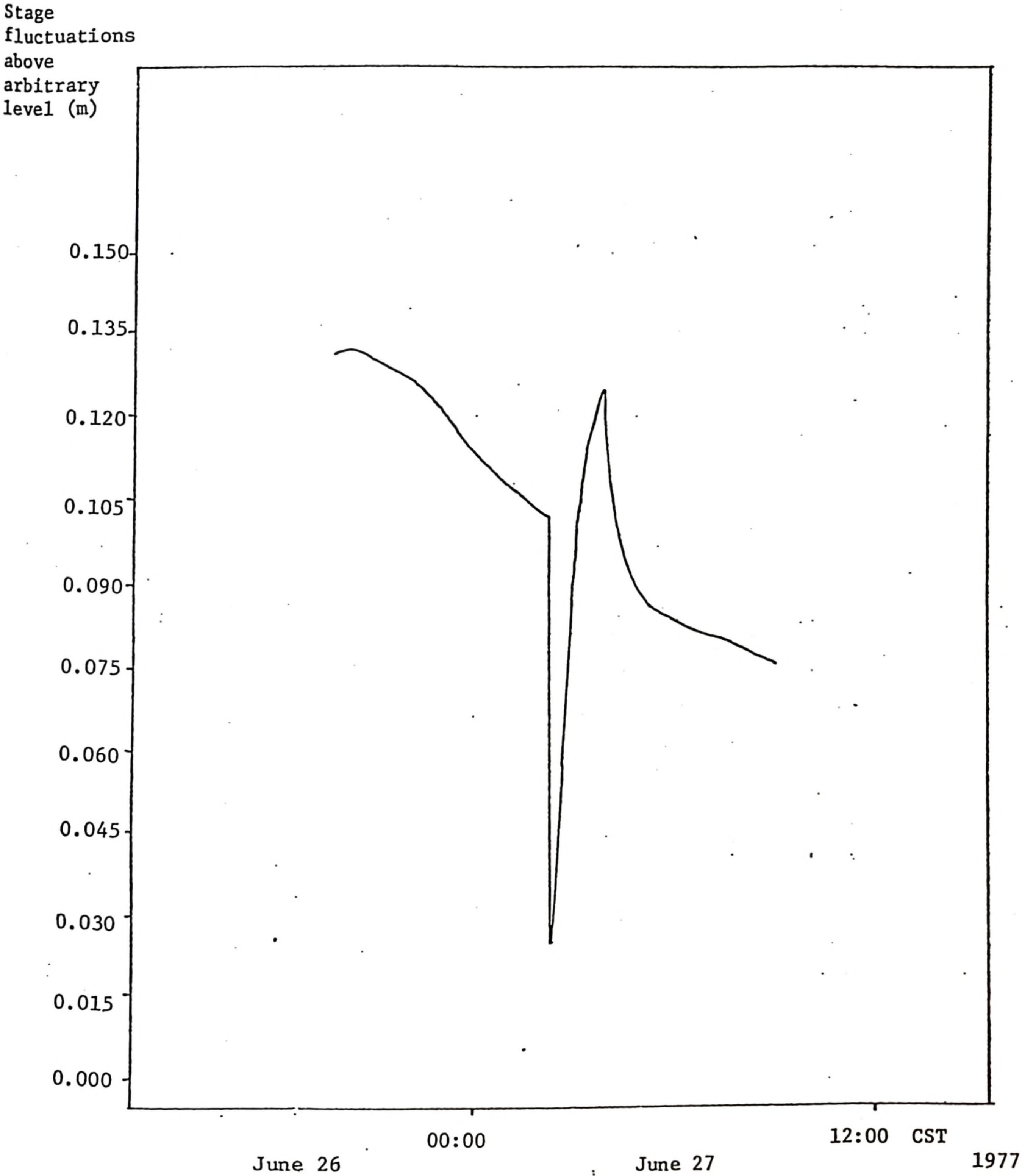


Figure 4.19 Stage fluctuation, as recorded on the hydrograph Site 2, due to temporary ponding behind collapsed snowwall during channel shifting.

4.2.1.2 Features

Genetically, two types of features are distinguished, namely depositional and erosional. Within these categories, the features are subdivided according to size scale (Table 4.2).

4.2.1.2.1 Depositional Features

On a macro scale, elongated debris ridges from casts on the snow surface mark the location of an abandoned channel before its flow was captured by an alternate channel created by lateral shifting or tunnelling. At a small tributary in the studied basin, such an elongated form was observed (Fig. 4.20). This ridge, that was deposited in a previous summer since there was snow accumulation above and underneath the gravel cast, covered an area of 93 m x 2 m. Sample plots from this ridge yielded an average of 123.3 kg[1] of debris per square metre area. The entire cast therefore included 23 metric tons of material, most of which is gravel or pebbles. The fact that such coarse material was overlying the snowpack indicates that the flow carried considerable load which had to be eroded from a snowfree area upstream from the study reach. Minor casts were also found in similar situations at other upper basin tributaries (Fig. 4.21). Grain size analysis was performed on the material sampled. The particles exceeding 2 mm diameter constituted 95 percent of total sample weight (sample #7, Fig. 4.22).

At the lower reach of tributary No. 1, a cast of bed material (20 m x 7 m in area) was left on the snow surface after the stream was captured by tunnelling. Six cross sectional samples were taken, which

[1]density of material = 2.60 g cm^{-3}

TABLE 4.2

Morphological Features Characterising
Fluvial Processes in Snow-filled Valleys

	Scale	
	Micro	Macro
Depositional	Elongated snow contact debris ridge deposited on clastic bed while channel shifting. e.g. Site 2.	Elongated debris ridge cast on snow. e.g. Site 1. Fresh gravel deposits. e.g. upper basin tributaries
Erosional	Sequence of small carved channel. e.g. confluence of tributary to No.2.	Subparallel carved channels. e.g. Site 1.

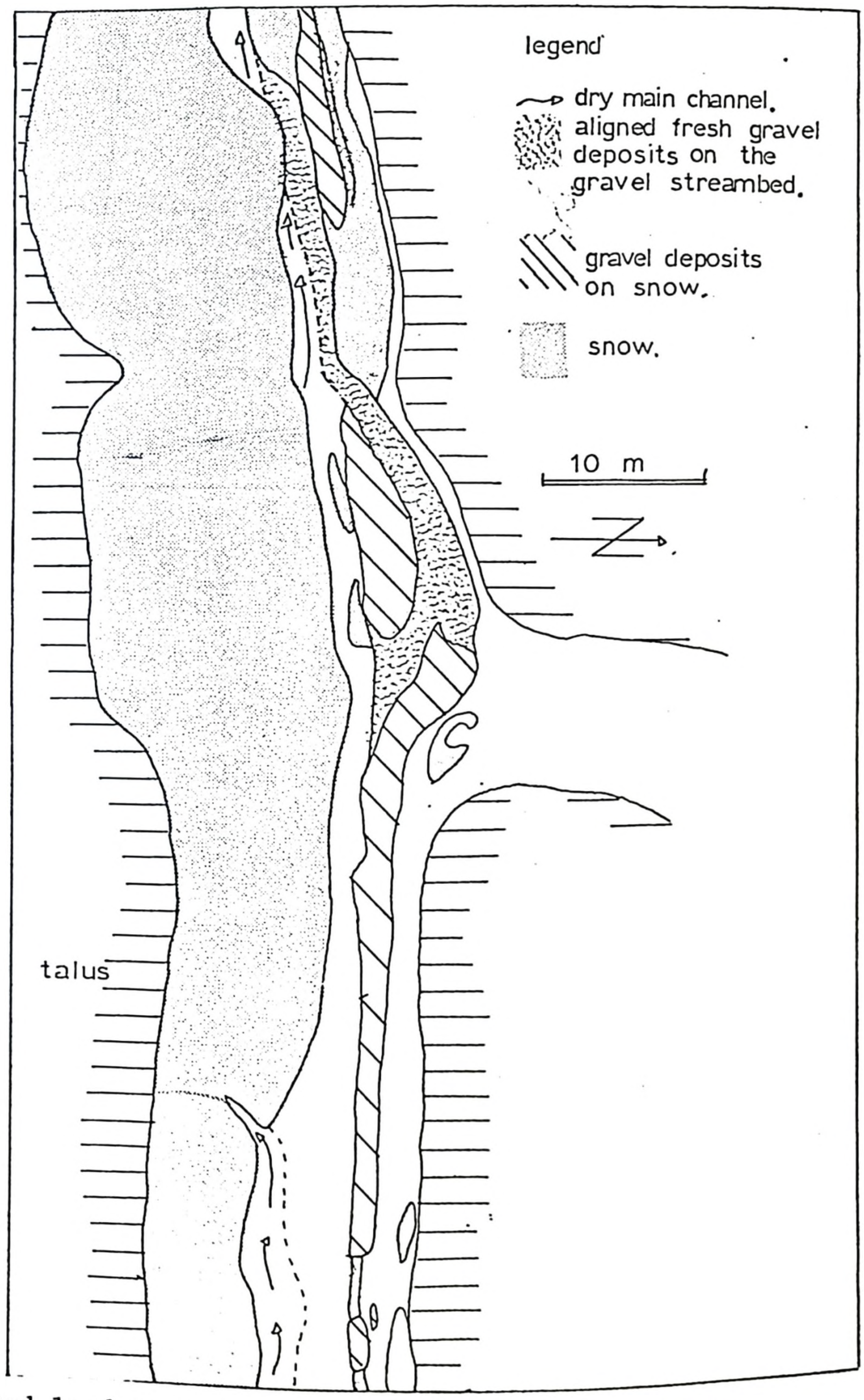


Figure 4.20 Bed load cast resting on snow after the stream was captured by channel shifting, upper basin tributary.

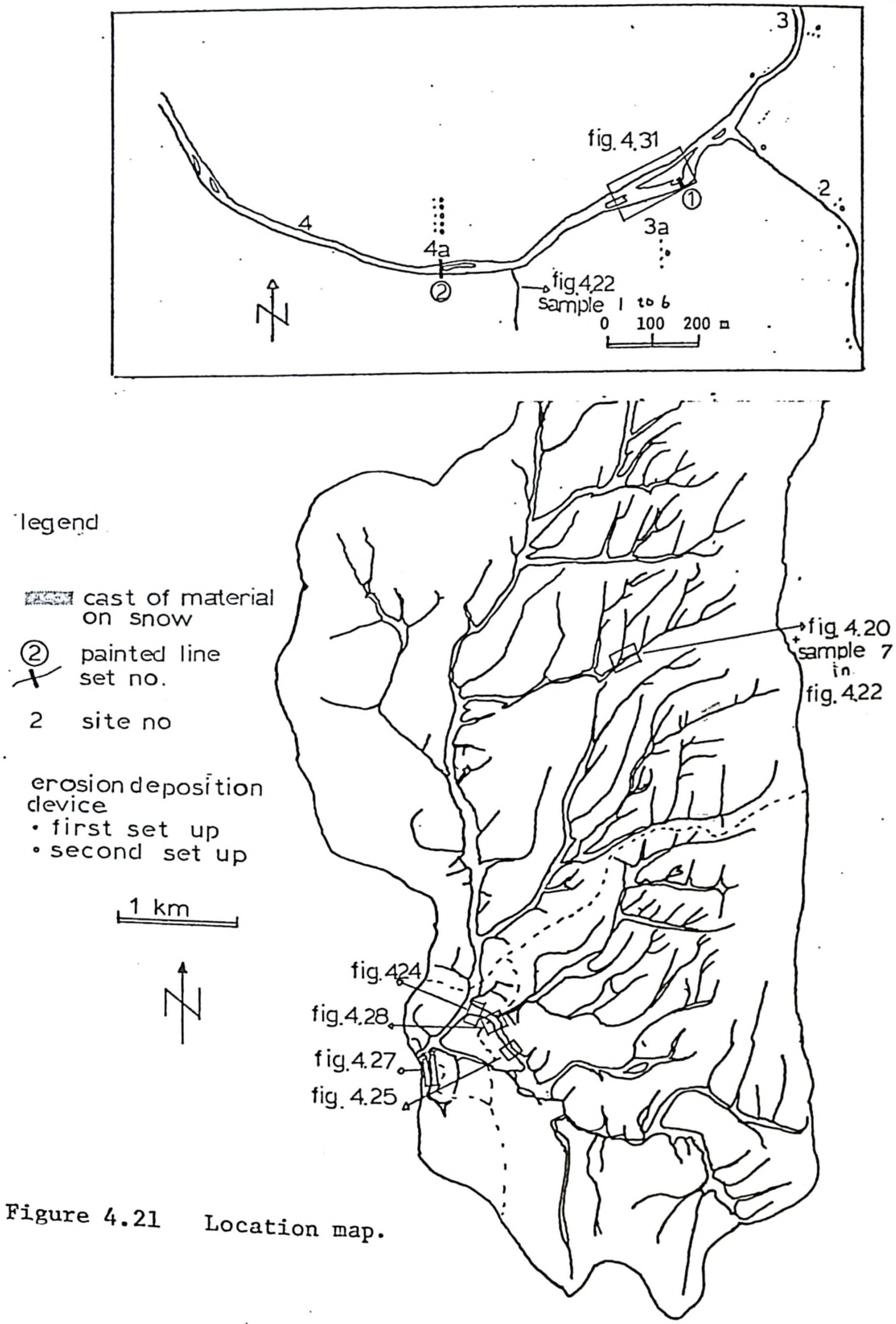


Figure 4.21 Location map.

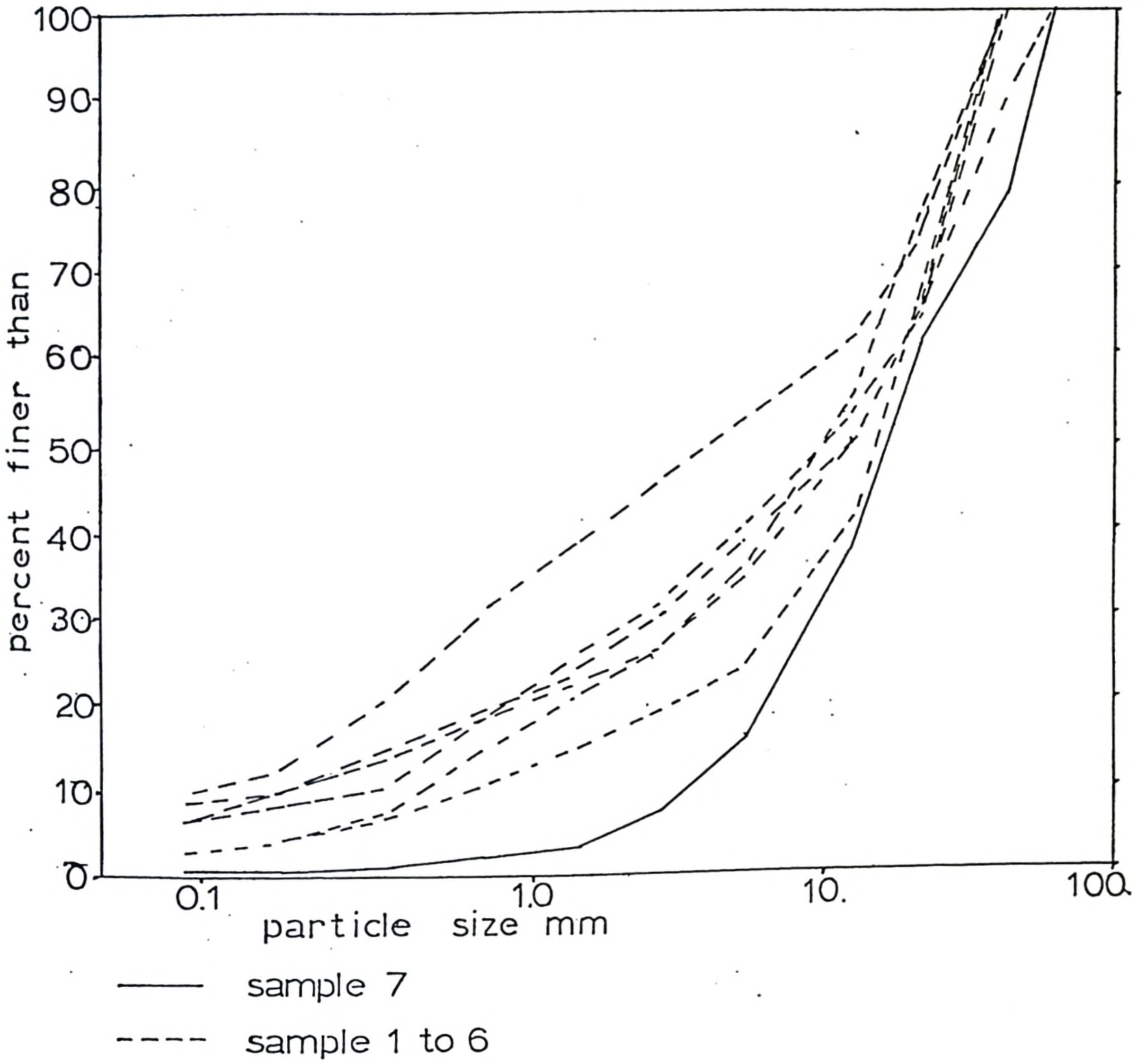


Figure 4.22 Grain size analysis. (see Fig. 4.21 for location).

yielded a mean weight of 6 kg of debris for each square metre area. The total load deposited was therefore approximately 0.9 metric ton. Overall, less than 30 percent of the material is smaller than 2 mm in diameter (sample #1 to #6, Fig. 4.22).

When the snow underlying the elongated debris ridges begins to melt, the ridge deteriorates into cones of debris which finally produces hummocky micro relief of fresh gravel (Fig. 4.20).

At another scale, micro features associated with channel shifting were observed near Site 2 and Site 3, where the main mechanism of channel opening was by lateral shifting. Figure 4.23 presents a typical cross-section of such a depositional feature. A top view of these features shows sinuous ridges with a length of about 1 m. They always occur as a group along an aligned trend (Fig. 4.24). Their formation is related to the different stages of lateral shifting as the channel encroaches upon a freshly collapsed snow-wall. At that time, the snow was impermeable enough to act as a channel bank. The flow at the channel edge, being subjected to greater resistance and because of lower velocity, deposited some material at the contact with the snowbank. With continuous under-cutting into the snow, the flow becomes diverted at short distances up-stream of the reach concerned. As a new channel is activated, the former channel becomes dry. The stranded snowblock is left to compact the underlying gravels and it protects them from being disturbed by fluvial processes (Fig. 4.25).

Where lateral shifting does not take place, the snowpack does not interfere with streamflow and no such features occurs. In these cases, the slope of the snowbank is usually much more gentle (Fig. 4.26).

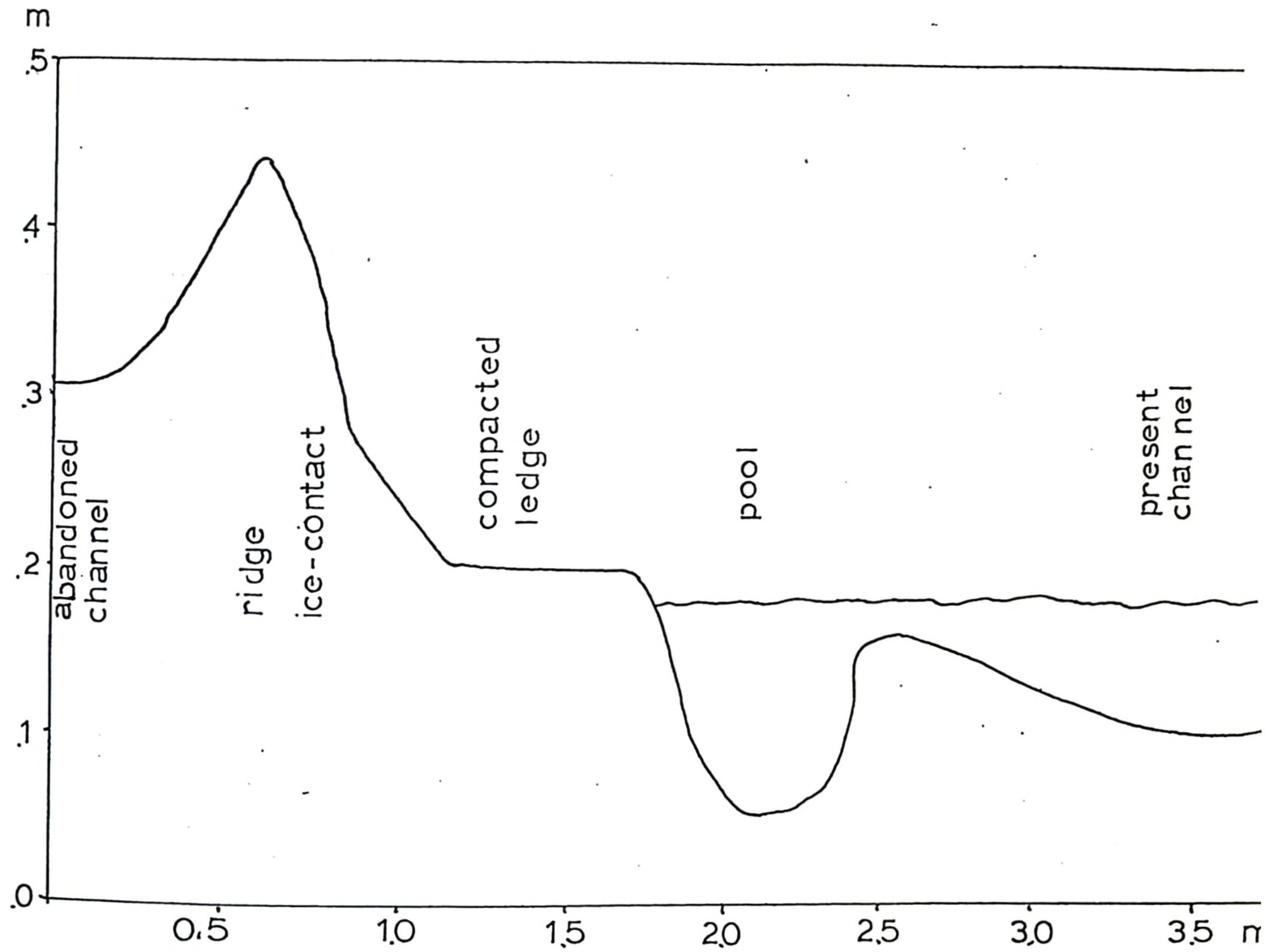


Figure 4.23 Magnified cross section of a depositional feature associated with channel shifting.

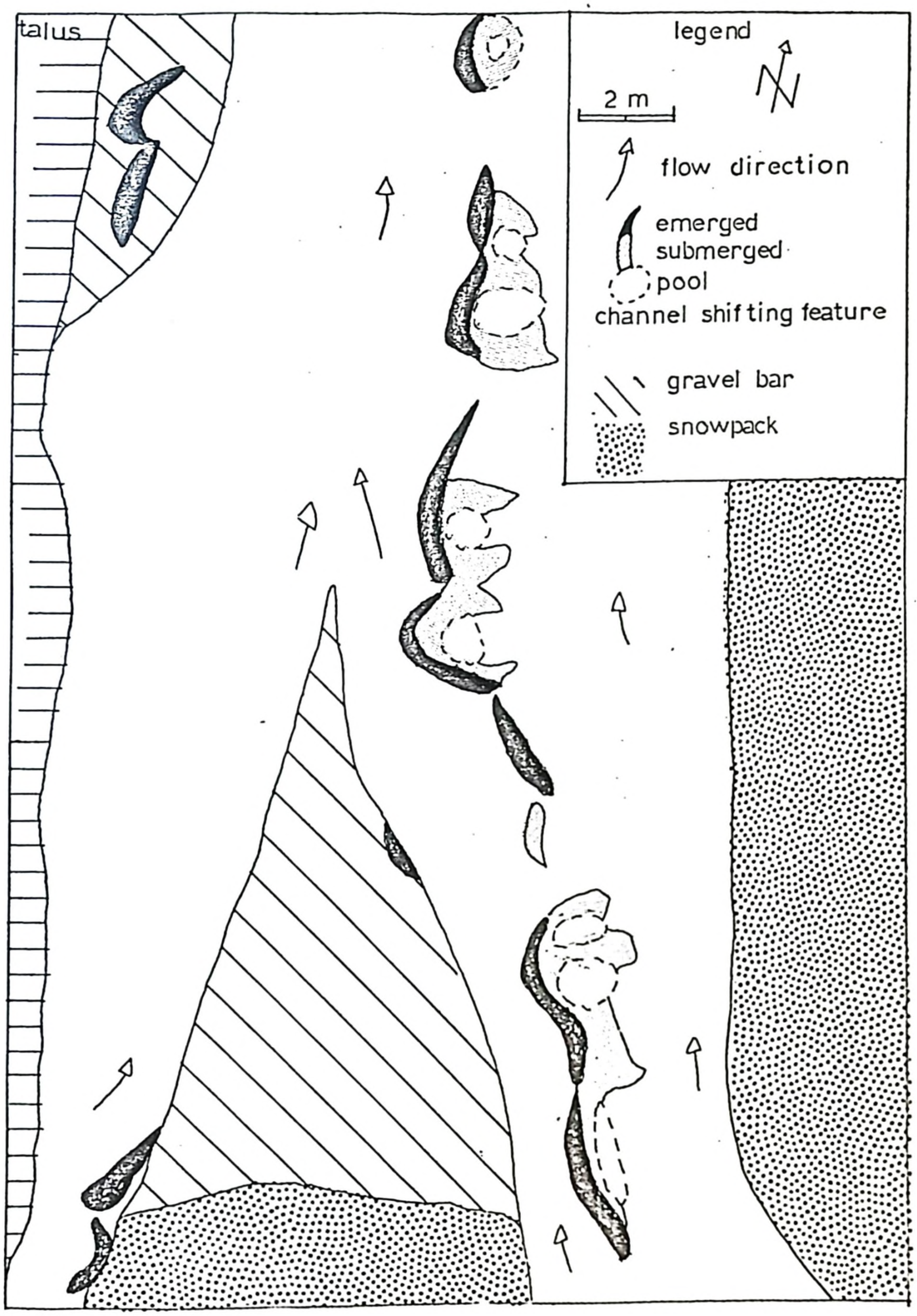


Figure 4.24 Plan view of channel shifting features.

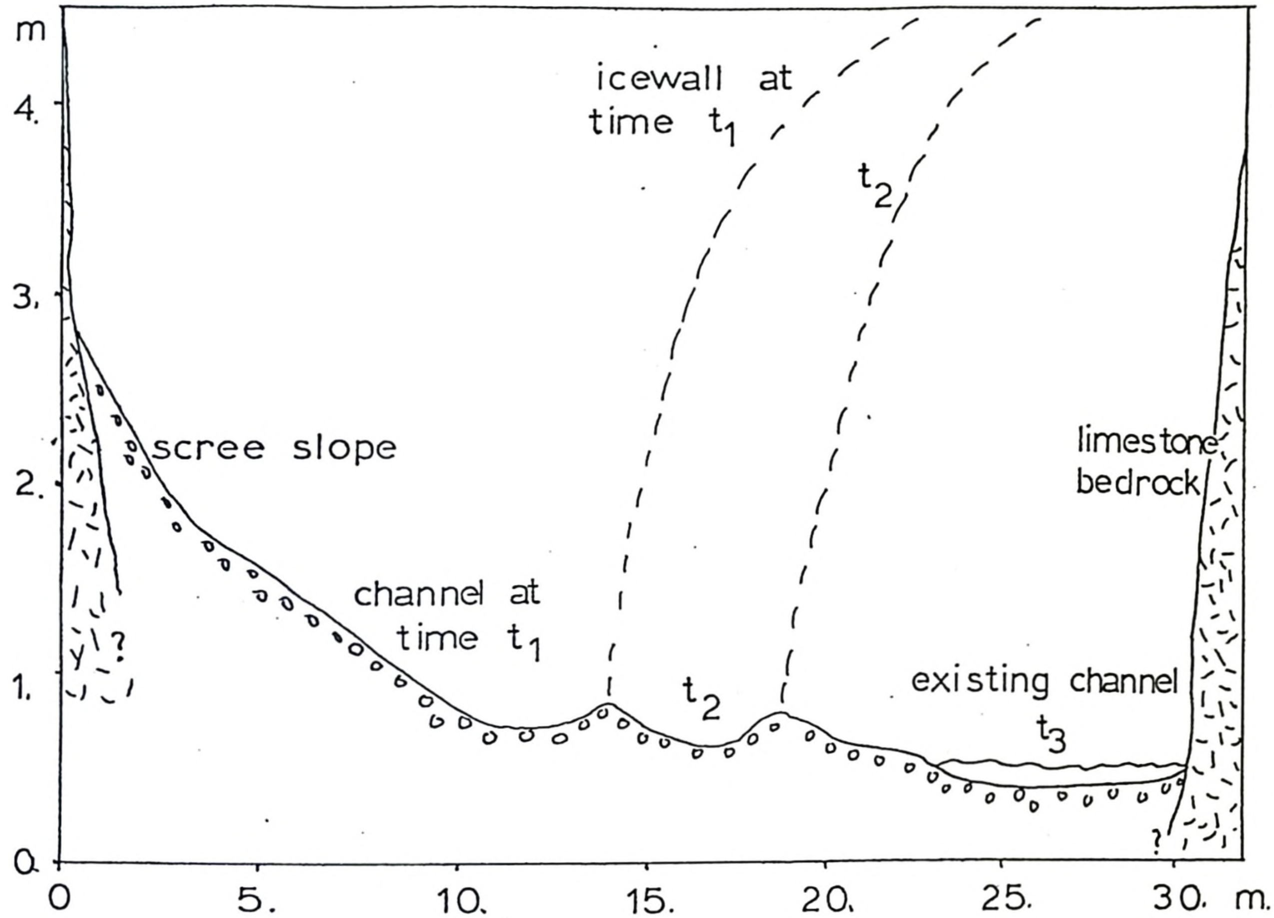


Figure 4.25 Cross section of an incised wide flat floored valley showing a chronological sequence of three positions of the snowbank related to channel shifting, upstream of the gauging Site 2.

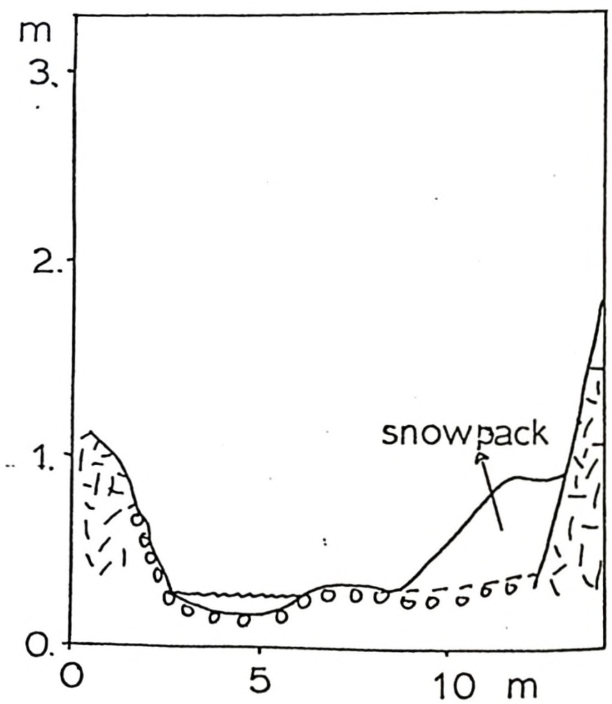


Figure 4.26 Gentle sloping snow surface section where there is no interference of the snowpack with the lowest point of the valley.

4.2.1.2.2 Erosional Features

Lateral shifting of channels developed in snow can produce erosional forms. At tributary No. 1, this process has carved out a channel on the gravel material running subparallel to the main thalweg. As shown in Figure 4.27, this channel length exceeds 35 m. Here, the presence of such degradational feature is probably induced by a relatively steep slope (c.0.045), thus providing a greater competence and capacity of the flowing water. A topographic section across the channel illustrates the effect of the snow diverting the flow to a higher position on the valley side (Fig. 4.27).

At the confluence of a tributary with Stream 2 and on a smaller scale, a sequence of small subparallel channels were carved out in the gravel bed to a depth not exceeding 0.15 m (Fig. 4.28). The length is about 6 m and they are located on a gentle (almost horizontal) slope. These channels were formed underneath a snowpack (possibly under hydrostatic pressure) as water seeped out from beneath the snowbank which impeded a normal drainage of water from the tributary. These close conduit hydraulic conditions will dictate degradation processes.

4.2.2 Summary

As a summary, features described in this section are admittedly inconspicuous. However they provide morphological evidence indicating the effects of different types of channel opening, hence the effect of the valley snowpack on fluvial processes. A quantitative assessment of the sediment transport rate of the Arctic streams will be presented in the section to follow.

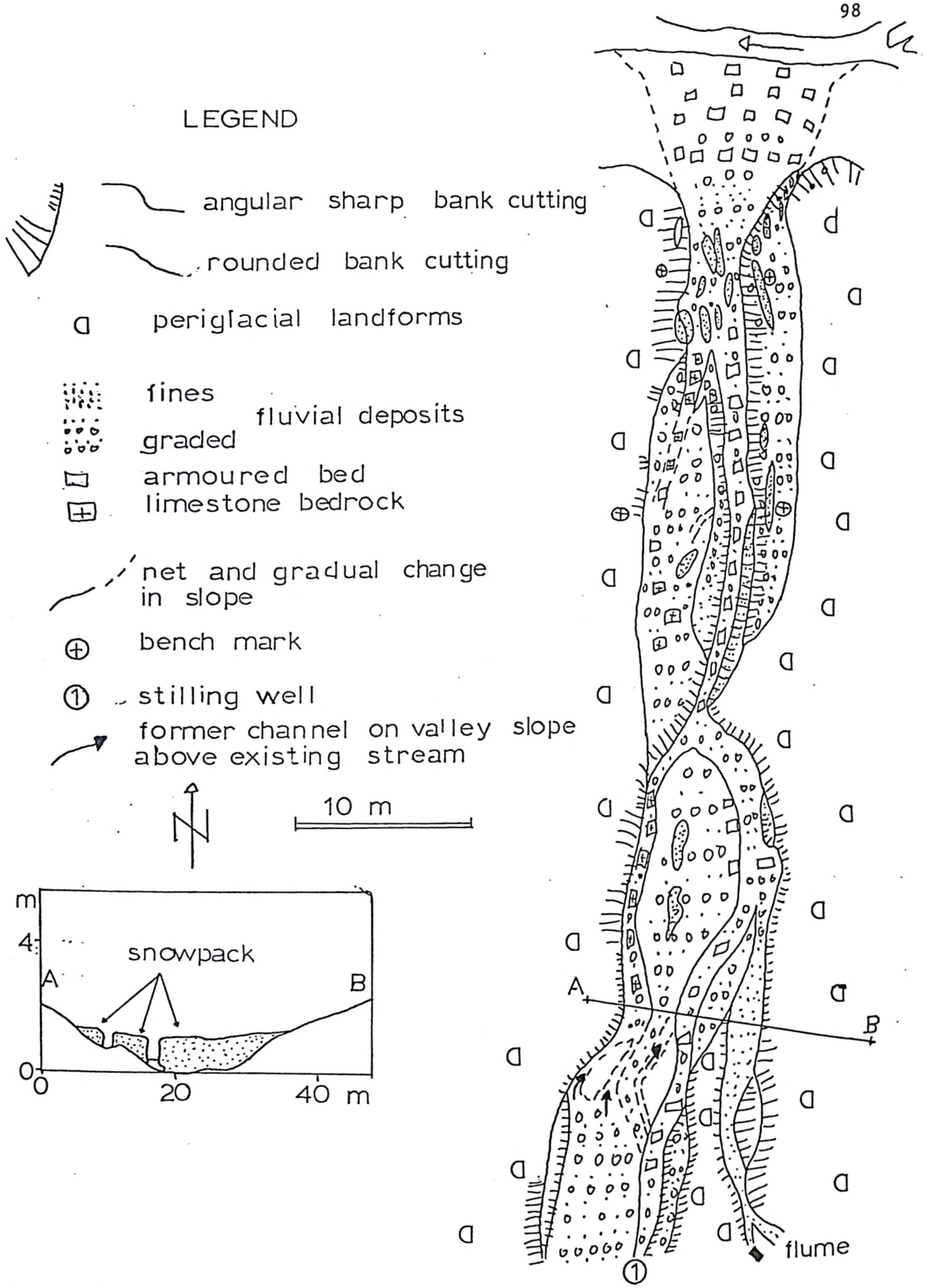


Figure 4.27 Distribution of surficial material at the lower reach at Site 1. Inset cross section shows the position of the carved channel in the valley snowpack of June 23, 1977.

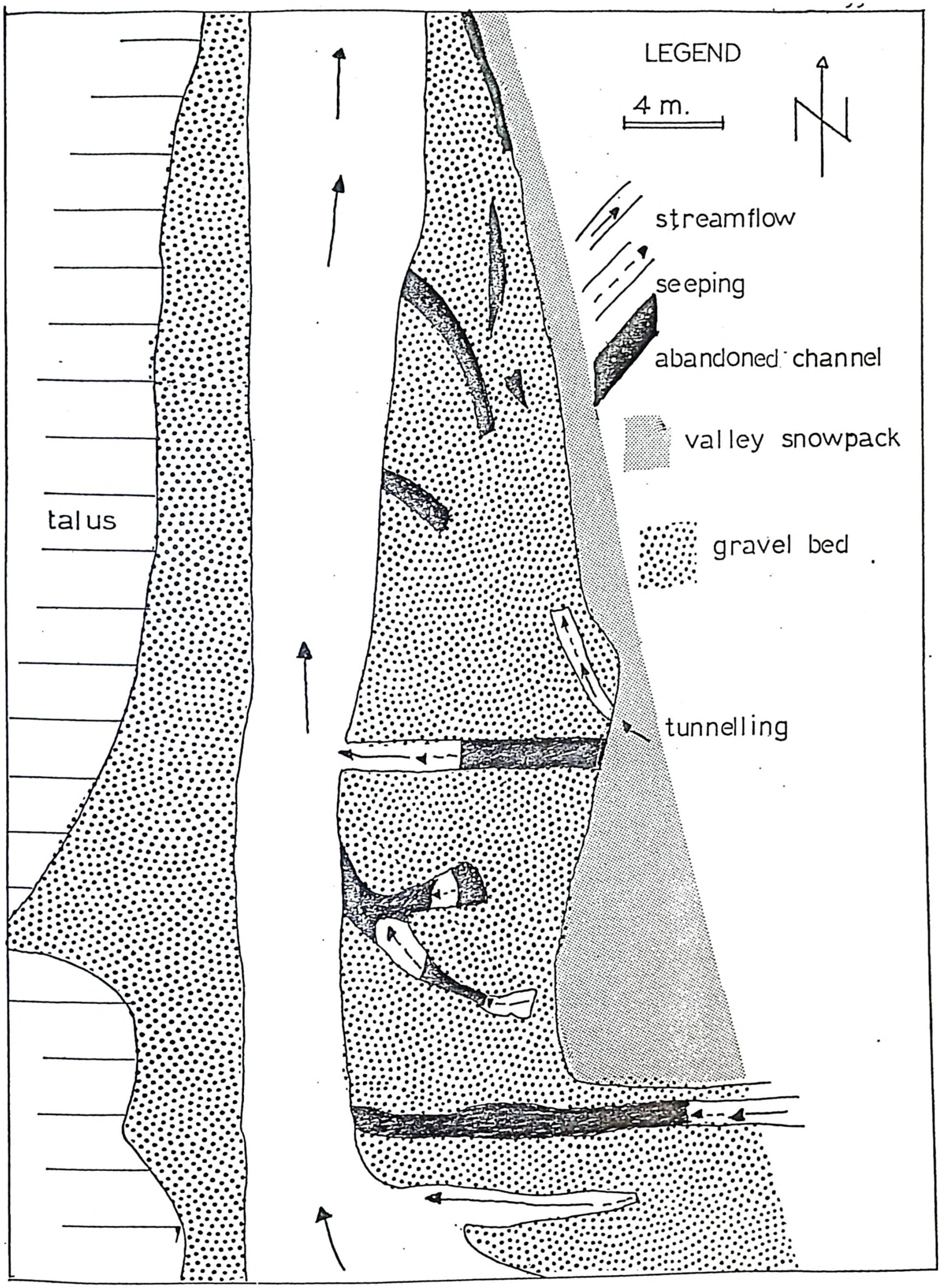


Figure 4.28 Erosional channel shifting features.

4.3 Effect of the Valley Snowpack on Fluvial Processes

During the peak flow period and along many reaches of the valley, the presence of snow in the valley floor prevents a direct contact between the bed material and the flowing water. As observed in the upper parts of the basin tributaries, perennial snowpack mitigates significantly the amount of geomorphic work, hence the fluvial processes of the streams along these reaches. On the other hand, a differential disposition of the snow in the valleys leads to water pondage behind snow ridges, the break up of which produces flash floods. This flood water has a higher potential to erode and hence to perform geomorphic work. Moreover, the snowpack may bias the stream channel outside its stable bed and subsequently enhance bank erosion.

Both destructive and protective effects can therefore result from an uneven distribution of snow in the valleys. An assesment of the snow-induced effects on the geomorphic processes can be made by comparing the actual geomorphic work observed in the study area with the potential geomorphic work deduced from the sediment transport capacities of the measured discharge. Their difference will provide information on the magnitude of the effect emanating from this snow drift accumulation in valleys.

4.3.1 Potential Geomorphic Work

The potential geomorphic work is defined as the mass of bed material that could be potentially transported over unit length of the channel per unit time if the flow was in direct contact with the bed material.

4.3.1.1 Types of Stream Loads

Following Church (1975), sediment transported by streams is partitioned into solute, wash and bed-material loads. The solute load transport will be considered in Section 4.3.2.1. Wash load is often limited by the supply of material rather than by the transport capability of the stream because the stream is adequately competent to transport materials of this size throughout the channel system. In this case, wash load cannot be assessed by streamflow considerations alone. On the other hand, in alluvial streams, the supply of bed material is seldom a limiting factor for a given flow. The amount of material that can be moved as bed material load therefore depends on the force exerted by the stream on the bed.

The bed material in the studied basin, shown in Figure 4.22, indicates lack of very fine materials. The finer material is carried as wash load while bed material load is transported at or near the bed. In the following section, the potential geomorphic work of streams will be devoted exclusively to bed load transport.

4.3.1.2 Computation of Bed Load Transport

"The theory of sediment transport is the least satisfactorily developed area in open channel hydraulics" (Church 1975, p. 35).

Comprehensive reviews of bed load transport processes can be found in such works as Rouse (1950), Leopold et al. (1964), Bogardi (1974), and Church (1975). In brief, incipient motion of a cohesionless particle occurs when the lift force of the fluid overcomes the drag force exerted by the submerged weight of the particle. Hence there exists a threshold value of flow before any movement can occur. At and above this value, bed load discharge will be the resultant of the balance of the sediment characteristics (i.e. fall velocity and the grain size for bed roughness determination) and the flow characteristics (i.e. shear velocity) (Rouse 1950).

Morris and Wiggert (1972) provided an elaboration of du Boys' sediment transport equation. In this approach the shear stress per unit area of the bed is given by equation (4.1).

The resisting force per unit area of the exposed grain is the product of its submerged weight and its coeficient of friction.

$$\text{resisting force} = [g(\rho_s - \rho)\delta]Cf \qquad (4.11)$$

where ρ_s is the density of sediment [M L^{-3}]

δ is the thickness of moving bed [L]

Cf is the coefficient of friction [0]

Equating (4.1) and (4.11), a critical shear stress τ_c is defined as the force per unit area that generates incipent motion of the exposed grains (Fig. 4.17).

Then, a transport function for bed load volume can be obtained as

$$g_s = C_s \, \tau(\tau - \tau_c) \qquad (4.12)$$

where g_s is the rate of transport in volume of bed material per unit time, per unit channel width [L^3 T^{-1} L^{-1}]

C_s is the sediment parameter [L^6 M^{-2} T^{-1}]

τ is the shear stress [expressed here as M L^{-2}]

The volume of C_s, expressing the relative susceptibility of a given sediment to movement, is the key factor on which depends a successful determination of g_s. Many subsequent workers developed other bed load equations based on du Boys' relationship, each of them aiming for a better assessment of this sediment parameter. These workers include Straub, Shields, Kalinske, Bagnold, Schoklitsch to name a few (Rouse 1950). The only assumptions to be satisfied when applying equation (4.12) are that alluvial stream should be shallow and with a wide cross-section where the the channel bank shear stress is negligible. However, even though these assumptions are fulfilled in the studied basin, du Boys' basin equation does not provide any theoretical consideration for armoured bed which prevails in the channel beds under consideration. For this reason, alternative semi-empirical equations were derived from flume studies undertaken in paved bed condition.

The Meyer Peter and Muller equation has been widely applied to gravel bed streams. The equation can also be adapted to suit armoured bed condition. The present study employs a version of the equation given in Church (1975).

$$g_s = 8.57 \frac{SR}{(n\,k_r)^{1.5}} - 0.67\,D_{50}^{1.5} \qquad (4.13)$$

where g_s is the bed load discharge in mass per unit channel width $[M\,L^{-1}\,T^{-1}]$

k_r is the particle friction obtained as $\dfrac{26}{D_{90}^{0.17}}$

D_{90} is the size below which 90 percent of the bed load armouring material are contained [L]

n is the total bed resistance (Manning's n) [0]

D_{50} is the mean grain size [L]

From the study reach chosen near Site 4, $D_{50} = 0.10$ m, $D_{90} = 0.20$ m. The latter yields a value of 34.0 for kr. After substitution

$$g_s = 8.57 \frac{SR}{(34.n)^{1.5}} - 0.00067 \tag{4.14}$$

Hydraulic variables including S, R and n are obtained as described in Section 4.1. Their daily means for Site 1, 2, 3, and 4 are applied to equation (4.14) to compute the daily rates of bed load transport at these sites.

Total bed load discharge was obtained from the product of g_s and the effective width, i.e. the width over which bed load transport occurs.

$$G_s = g_s W_{eff} \tag{4.15}$$

where

$$W_{eff} = W K_1 \tag{4.16}$$

W is the total width of stream cross-section as measured or computed by hydraulic geometry [L]

K_1 is the decimal fraction of total channel width where bed boad transport occurs.

From Church (1972), a mean value of $K_1 \approx 0.6$ was assumed. Equation (4.15) then allows a computation of the potential geomorphic work performed during the field season of 1977 at Site 1, 2, 3, and 4 (Fig. 4.29).

4.3.1.3 <u>Discussion</u>

Applying the bed load computation to Site 4 (Fig. 4-29), the maximum daily potential load was close to 20.metric tons day^{-1} and this

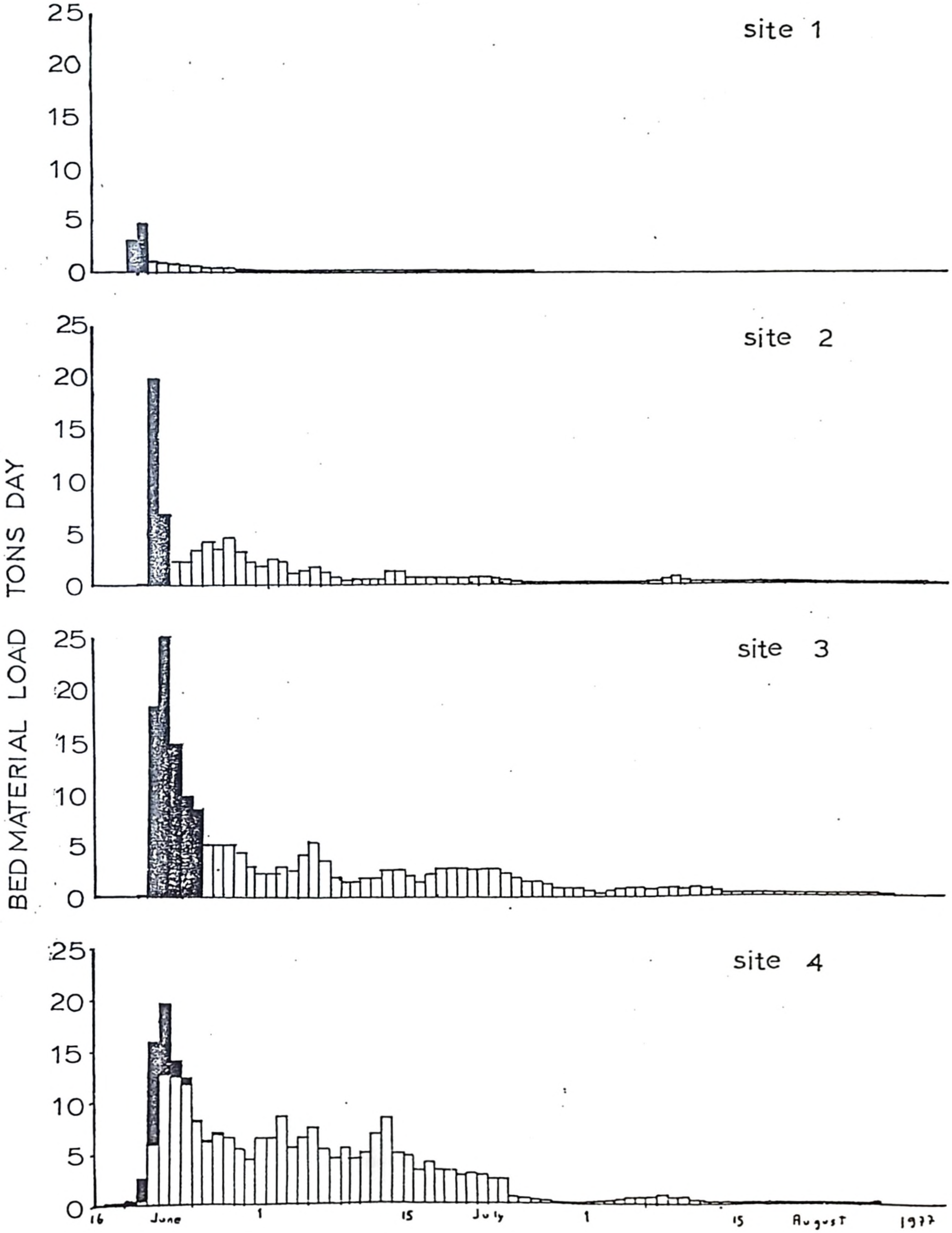


Figure 4.29 Bed material load discharges at Sites 1, 2, 3, 4, 1977. The shaded area indicates the computed protective effect of the valley snowpack on fluvial processes as obtained by equation 4.19.

was calculated for June 22. Total annual load can be obtained as the integral of daily loads for the duration of the flow season, and the value is computed to be 241.5 metric tons for the year 1977. Standardising with respect to basin area and to volume of water discharged, the basin being studied yielded 7.32 tons yr^{-1} per km^2 of basin area or 0.5413×10^{-4} tons yr^{-1} per m^3 of water flow. These values can be compared with results from other Arctic streams (Table 4.3). Both the Lewis and the Schei Rivers yielded values one or two orders of magnitude larger than our stream at Site 4. A closer agreement with the Mecham River was encountered because of their environmental similarity. Large differences with the other rivers are due to a pro-glacial disposition of the Lewis and the Schei Rivers which occurs in high energy and high material supply environments.

The same considerations apply to Sites 2 and 3 where the bed load transport rates ranged in the same order of magnitude as Site 4 (Table 4.3). On the other hand, Site 1 presents higher values that are probably due to a steeper channel bed gradient.

It is to be noted that annual total loads of 15.03, 85.93, 152.99 and 241.55 tons yr^{-1} for Sites 1, 2, 3 and 4 respectively are conservative estimate since only mean daily values of the hydraulic data were used in the computation. Hollingshead (1971) suggested instantaneous variations in shear stress on the bed to be nearly twice the mean value. Because of this inherent characteristic of bed load discharge, a large proportion of total bed material transport occurs within short periods of time, i.e. during low frequency but high magnitude events (Church 1972, McDonald and Lewis 1973). Despite these considerations, our under-estimation of daily bed load may have inadvertently take into account

TABLE 4.3

Comparison of bed load for four Arctic basins

(I) Standardise per unit area

Stream		Year	Potential bed load yield	Method	Reference
Unnamed	Site 1	1977	30.06 tons yr^{-1} km^{-2}	Meyer Peter and Muller	Sauriol 1977
(near Resolute)	Site 2	1977	8.59 tons yr^{-1} km^{-2}	(as in Church 1975)	"
	Site 3	1977	7.29 tons yr^{-1} km^{-2}	"	"
	Site 4	1977	7.32 tons yr^{-1} km^{-2}	"	"
Lewis (Baffin Is.)		1963	1049.00 tons yr^{-1} km^{-2}	Meyer Peter and Muller 1948	Church 1972
		1964	236.00 tons yr^{-1} km^{-2}	"	"
		1965	482.00 tons yr^{-1} km^{-2}	"	"
Schei (Ellesmere Is.)		1974	164.60 tons yr^{-1} km^{-2}	"	Bennett 1975
Mecham		1971	1.06 tons yr^{-1} km^{-2}	"	Cogley 1975
(Cornwallis Is.)		1971	4.70 tons yr^{-1} km^{-2}	Kalinske	"

(II) Standardise per unit volume of water

Stream		Year	Potential bed load yield	Method	Reference
Unnamed	Site 1	1977	4.0668×10^{-4} ton yr^{-1} m^{-3}	Meyer Peter and Muller	Sauriol 1977
(near Resolute)	Site 2	1977	0.5180×10^{-4} ton yr^{-1} m^{-3}	(as in Church 1975)	"
	Site 3	1977	0.6729×10^{-4} ton yr^{-1} m^{-3}	"	"
	Site 4	1977	0.5413×10^{-4} ton yr^{-1} m^{-3}	"	"
Lewis (Baffin Is.)		1963	12.8299×10^{-4} ton yr^{-1} m^{-3}	Meyer Peter and Muller 1948	Church 1972
		1964	8.0236×10^{-4} ton yr^{-1} m^{-3}	"	"
		1965	9.9492×10^{-4} ton yr^{-1} m^{-3}	"	"
Schei (Ellesmere Is.)		1974	7.8988×10^{-4} ton yr^{-1} m^{-3}	"	Bennet 1972
Mecham		1971	0.0355×10^{-4} ton yr^{-1} m^{-3}	"	Cogley 1975
(Cornwallis Is.)		1971	0.1582×10^{-4} ton yr^{-1} m^{-3}	Kalinske	"

the imbricated nature of the bed material. To overcome the imbricated conditions, higher instantaneous energy is needed to generate incipient notion of an imbricated particle (Church 1975). This invariably reduces the transport power exerted by instanteneous fluxes in the streamflow energy.

4.3.2 Actual Geomorphic Work

The actual geomorphic work is defined as the mass of bed material that is actually transported as bed load over a unit length of the channel per unit time, by the streamflow in direct contact with the clastic bed.

4.3.2.1 Field Measurements

In the field, transported sediment is subdivided into dissolved, suspended and bed loads. For short distances along a reach, it is impossible to distinguish between wash load and bed material load being transported in suspension.

Supended and dissolved sediment concentrations, expressed as milligram per litre are shown in Figure 4.30. These values correspond closely to these of Jason's Creek, southwestern Deveon Island, and to dissolve sediment concentration of Mecham River, Cornwallis Island (Cogley 1971) during the summer of 1970. Suspended sediment concentration of the latter stream is about half an order of magnitude larger than the one of our basin. Owing to the time scale of sampling (twice daily) the graph of suspended sediment concentration for the study basin does not show the sporadic, yet rather frequent waves of turbid water during the

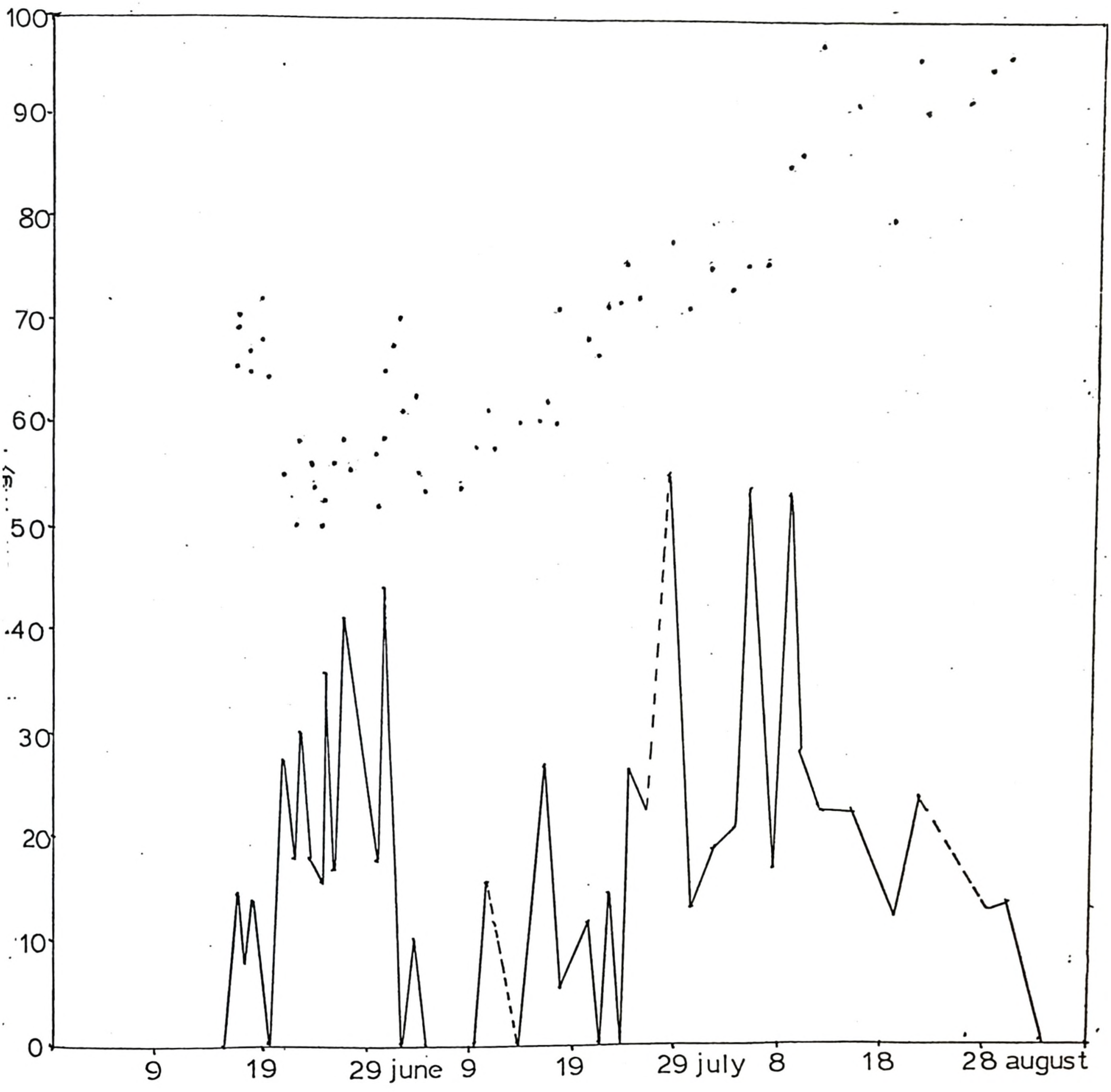


Figure 4.30 Suspended(———) and dissolved for calcium and magnesium (. . .) concentrations, Site 4.

early part of the flow season. Such high concentrations of suspended sediment seems to relate with the temporary storage and release of water due to collapsing snow blocks (Section 3.1.5).

Bed material load was estimated by indirect methods because no sediment trapping device can be readily employed. A bed load sampling device à la Østrem (essentially a fence with regular mesh size opening through out its length) was considered (Østrem 1975); but the abundance of bed materials below the size range of pebbles prevented the sampling device to be successful.

One of the indirect methods consisted of painting a line across the stream channel on imbricated undisturbed bed material. However, it was extremely difficult to make adequate observations during the high flow period, after which it was impossible to conclude whether the painted materials were washed away or whether the stream aggraded along that reach.

Lines of overloose material were set up across the channel at Site 4A. These lines were lay out in two sequences: one to cover the snow-melt peak flow period and a second one for the seasonal recessional flow period. One hundred and fifty subrounded pebbles, with their a-axis smaller than 55 mm and a b-axis ranging between 25 and 40 mm were placed on the streambed along a 15 m line. Only 40 pebbles (located along the main channel portion of the section) moved, of which 55 percent were lost (probably transported for substantial distances). The distribution of those pebbles recovered (Fig. 4.31) shows clearly that the lost pebbles were from the main channel, this represents the bed load movement at that site for a 35 day-period. No daily rates were observed.

After the high flow period, the rate of movement of overloose

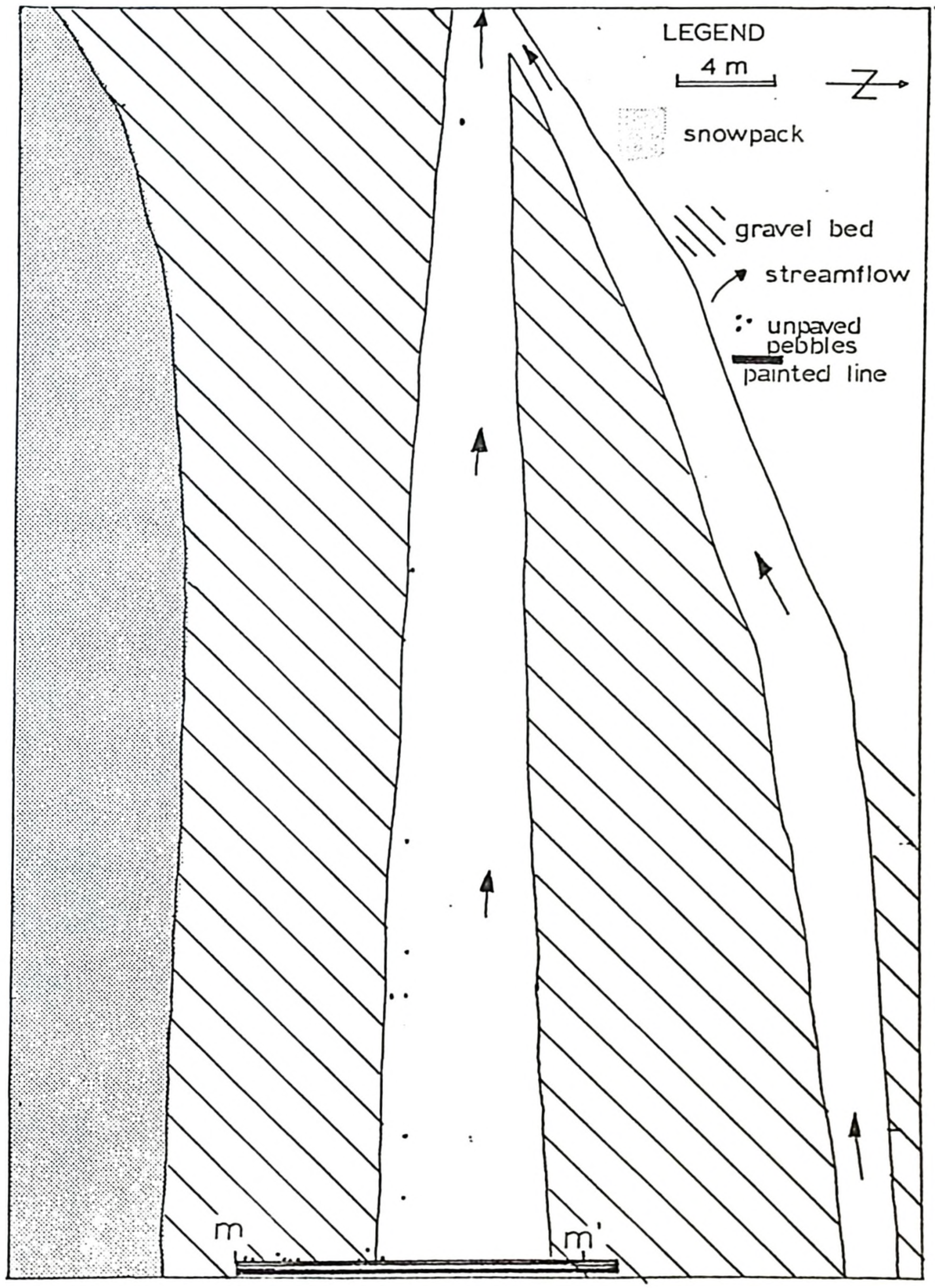


Figure 4.31 Distribution of the recovered loose pebbles, Site 3A, July 21, 1977. M-M' shows the maximum channel width between the period June 16 and July 21.

pebbles was observed more closely. However, as shown in Figure 4.32, little movement of bed material of that size occurred, except for a settlement of some individual pebbles onto a more stable surface.

A second indirect method was utilised to estimate the rate of bed load transport using spikes and washers to show the amount of degradation and aggradation at individual points along a reach. After the snowmelt peakflow the erosion and desposition characteristics at these points are shown in figure 4.33 with the measurements at 18 markers positioned according to their location along the longitudinal profile of the stream. Thirteen other similar devices were reset in mid-July. They demonstrated an absence of significant geomorphic work during the recessional flow period, except at Site 3 where the stream bed aggraded another 0.05 m. A detailed network of spikes and washers was set up at Site 4A alongside the second non-paved painted line. This site occupies a constricted channel which should provide an optimum condition for sediment transport (due to a higher velocity and a lower pressure according to Bernoulli's equation in Church 1975). Neither deposition nor erosion occurred at that site during the second half of the flow season.

Other evidence of bed material load occurrence includes a cast of substantial bed material dropped by a former stream prior to its flow being captured. As shown in Section 4.2.1.2.1, approximately 0.9 metric tons of coarse material was deposited on the lower reach of tributary No. 1, during a 5 day snowmelt high flow period and before it was captured by flow diversion. As a conservative estimate (because some sediments must have been carried out of the reach during that period) a mean of 0.18 metric ton of bed material was removed from the basin each day,

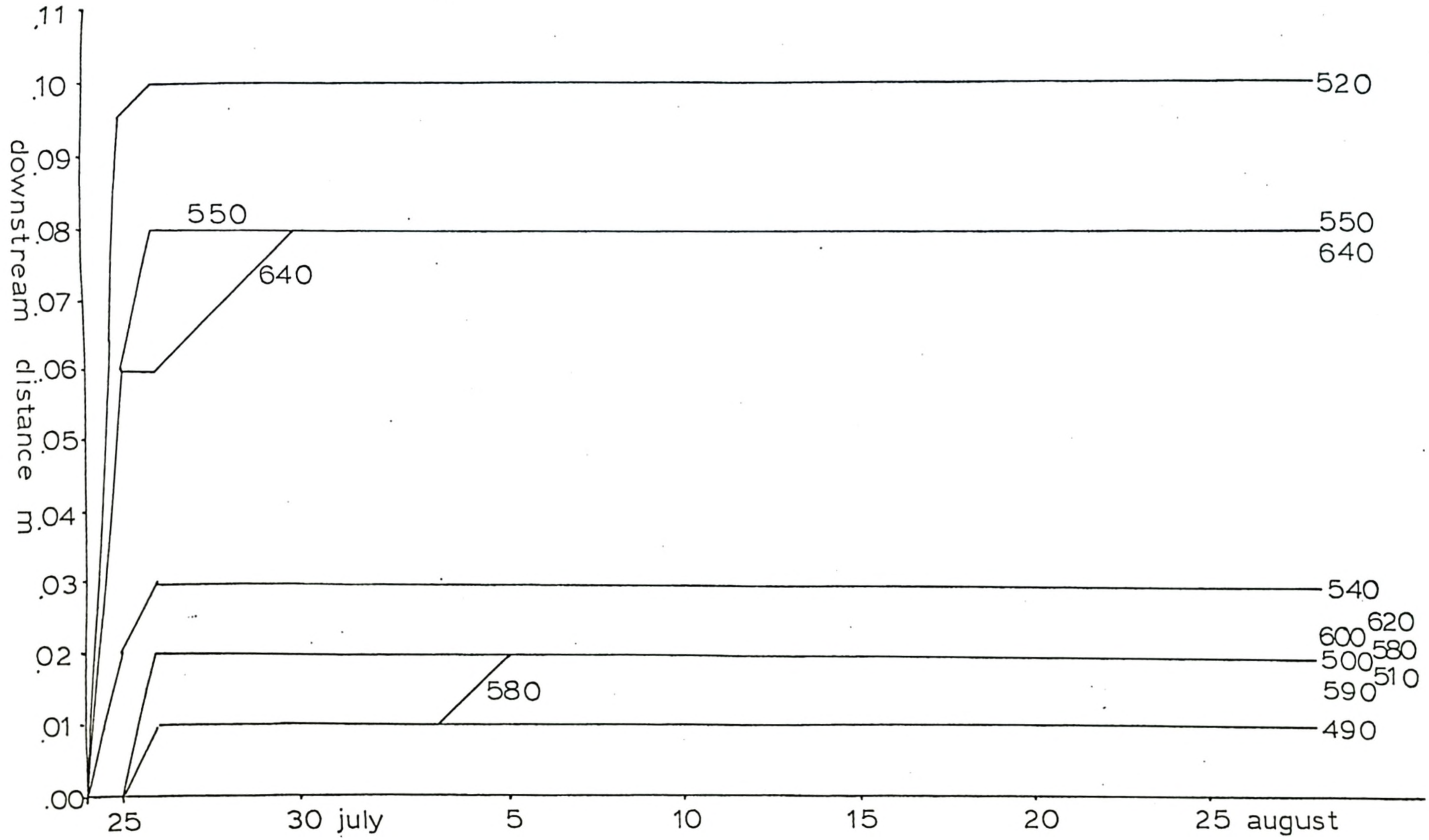


Figure 4.32 Rates of movement of loose pebbles, July 24 - August 28, 1977, Site 4A. (pebble no. refers to their respective distance in cm from the south bank).

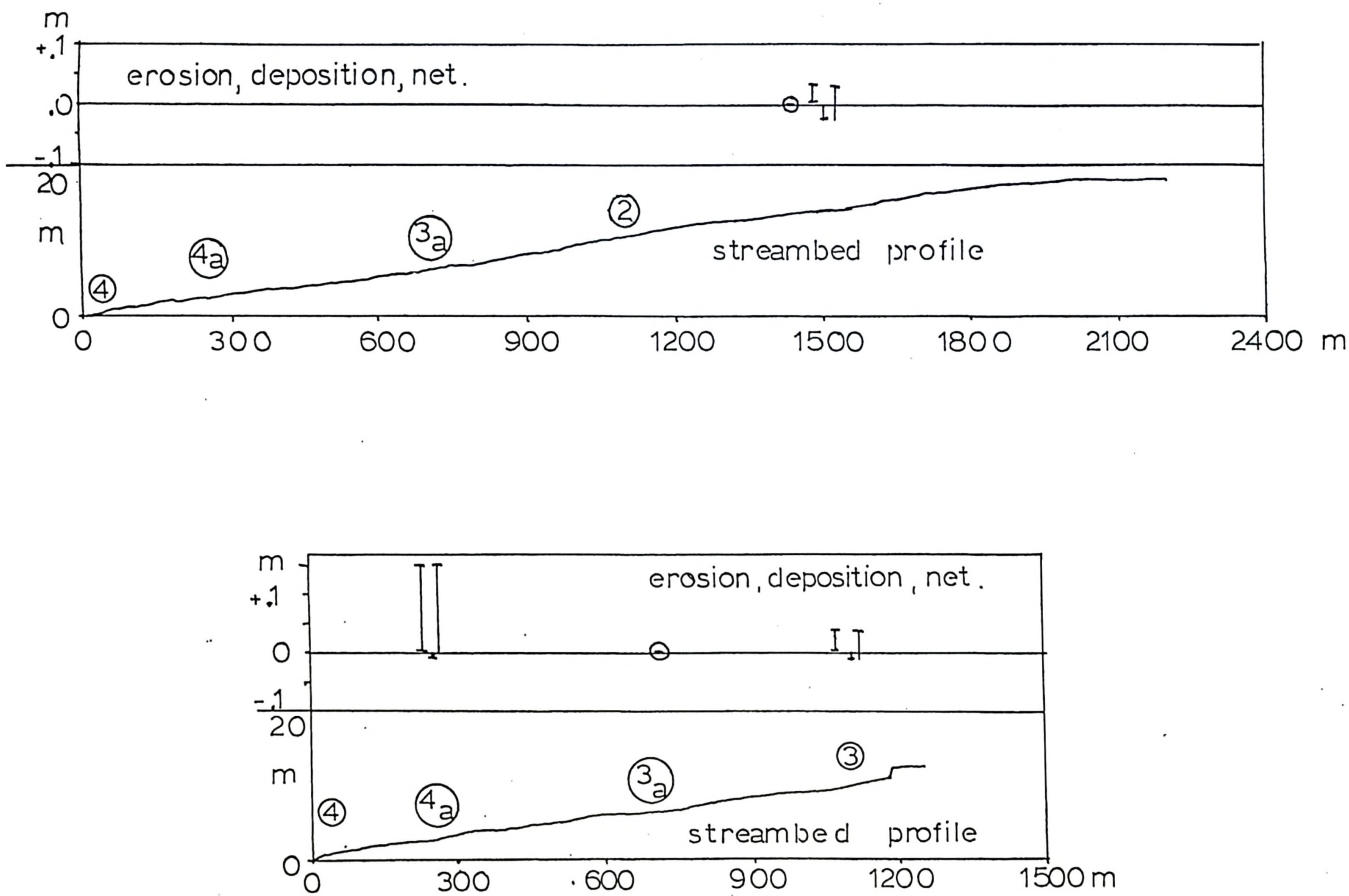


Figure 4.33 Net erosion and deposition at-points along the channel, mid-May to mid-July, 1977. Positive values indicate deposition and negative values, erosion.

representing a figure of 0.36 ton day^{-1} per km^2 or 0.04549 x 10^{-4} ton day^{-1} per m^3 of streamflow for this 5 day-period. The bed material load of Basin 1 can therefore contribute substantially to the total load of the main stream. Although such casts of deposits need not be representative of the rates of bed material movement, they are indicative of the magnitude of the sediment transport processes involved during the high flow events.

A convincing but non-measurable evidence is the bruises that moving pebbles left on the legs of the wader while measuring stream discharge. Sizeable gravels are observed to be transported over a bottom ice layer which constituted the channel floor at a shallow clear reach near Site 3. The rate of movement of the bed materials was sufficiently large to damage a current meter.

Bed material load remains the most difficult component of the total load to be measured. In view of inadequate information on observed bed material load, an alternative method has to be employed to show the effect of snow and bottom ice in channel upon the geomorphic work of stream flow.

4.3.2.2 Bed Load Transport and the Channel Snow and Ice Cover.

Based on observations pertaining to the behaviour of channel opening in snow-filled valleys, part of the specific energy of the running water is consumed in flushing the snow and bottom ice out of the valley. Until stream water is contact with the bed material, no geomorphic work can be performed on the bed (Fig. 4.34).

Using the observation of the snow and bottom ice duration along a typical channel cross-section, the actual effective width can be

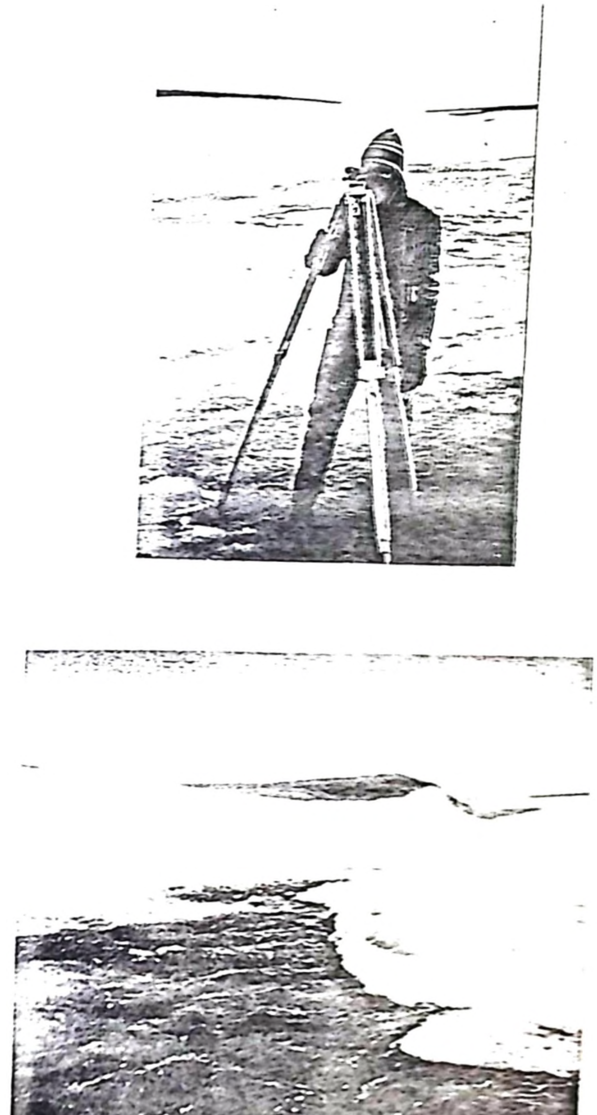


Figure 4.34 Examples of no-geomorphic work occurrence until there is a direct contact between the stream flow and the bed.

determined; this width is obtained as

$$\text{actual } W_{eff} = W_{eff} K_2 \tag{4.17}$$

where K_2 is the decimal fraction of W_{eff} where the bed is free of snow and ice.

Therefore the adjusted bed material transport ($G_{s\ adj}$) can be obtained (Fig. 4.29), and its value at a site is

$$G_{s\ adj} = \int_{W_{e1}}^{W_{e2}} g_s\, d_w - \int_{W_{E2}}^{W_{E3}} g_s\, d_w \tag{4.18}$$

where the limits of the integrals are defined in figure 4.35.

At Site 4, the daily change of K_2 was obtained from the observations of the channel bed characteristics (Fig. 4.36). For the computation of the bed material load based upon mean daily values of shear stress, it was assumed that any protected sections of the width was an integral part of the effective width. At Site 4, using equation 4.18, the actual load for the flow season is found to be 218.01 tons yr^{-1} corresponding with a yield of 6.6064 tons yr^{-1} km^{-2} and 0.4885 x 10^{-4} ton yr^{-1} m^{-3}. The actual daily bed material load is summarised in Figure 4.29, and it shows a maximum daily value of 12.83 tons day^{-1} at Site 4.

4.3.3 Effect of Valley Snowpack on Fluvial Processes

In this section, the protective effect of substantial snowpack in valleys on fluvial processes will be examined quantitatively followed by a discussion on the temporal and spatial variations of this effect on

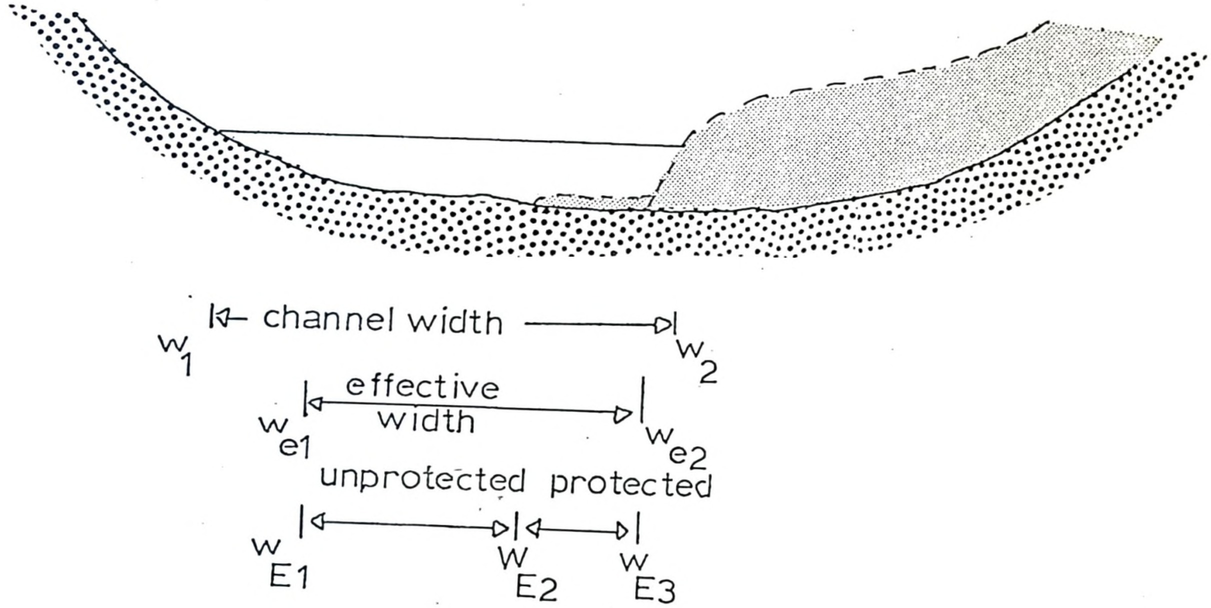


Figure 4.35 Schematic diagram showing an example of total channel width (W) the effective width (W_e) and the adjusted width (W_E).

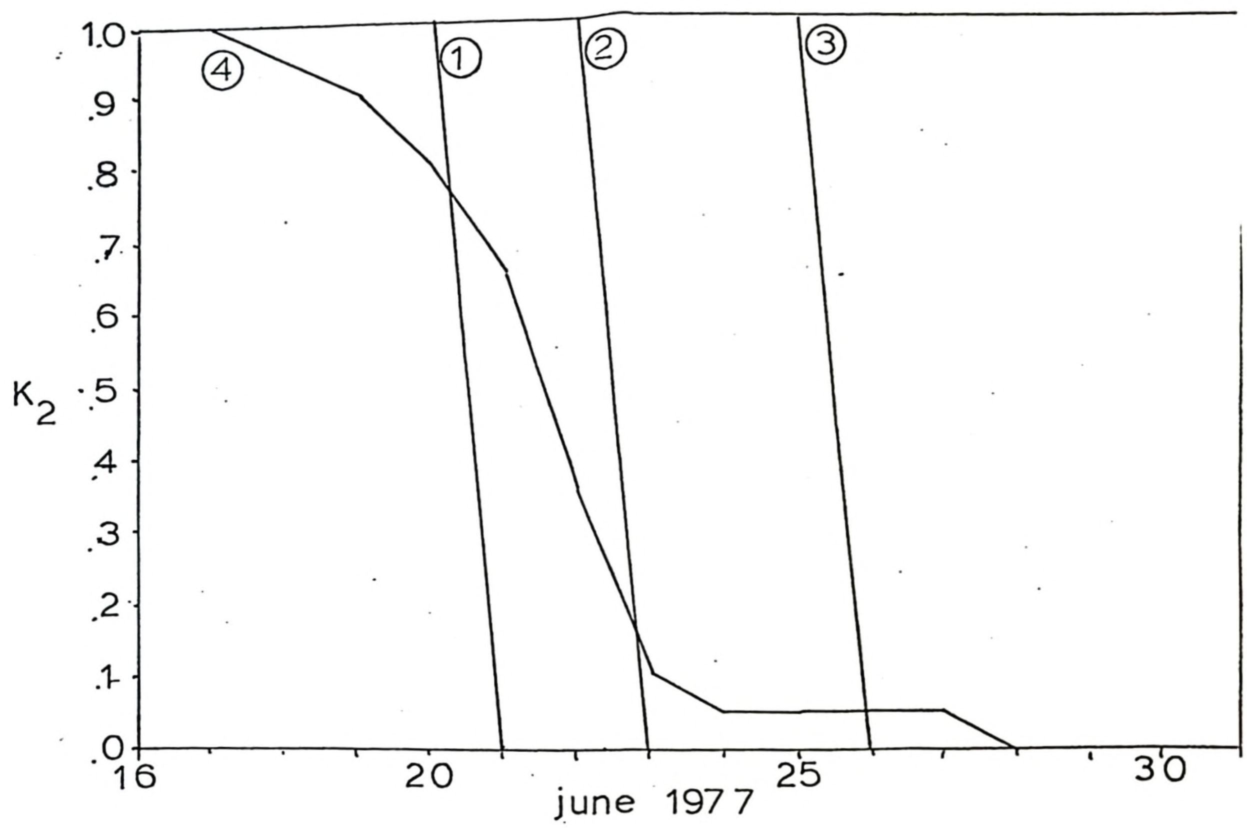


Figure 4.36 Rate of change of the proportion of the channel bed protected by snow and bottom ice (K_2).

fluvial processes.

4.3.3.1 Protective Effect

The protective effect of the snowpack on fluvial processes is defined as the amount of material that could have been removed and transported, should snow and ice be absent on the bed. This effect is expressed in units of mass per time, and, for a flow season, it can be defined as:

$$\text{Protective effect} = G_p - G_a \tag{4.19}$$

where

$$G_p = \int_0^T G_s \, dt$$

$$G_a = \int_0^T G_{s\ adj} \, dt$$

where G_s is obtained from equation (4.15)

$G_{s\ adj}$ is obtained from equation (4.18)

T is the duration of the flow season

dt is the time interval.

At the mouth of the studied basin (Site 4), the protective effect was computed on a daily basis and integrated for the flow season. For the summer of 1977, G_a = 218.01 tons yr^{-1}, out of a potential of G_p=241.55 tons yr^{-1}. This leaves 23.53 tons yr^{-1} (or 9.75 percent) of the bed material load shielded against transport losses because of the presence of snow and ice on the streambed (Fig. 4.29).

4.3.3.2 Temporal and Spatial Variations of the Protective Effect

In High Arctic basins, the bulk of the flow occurs when valleys are still choked with a substantial snowpack. The rate of clearance of this snow and bottom ice depends on the magnitude of the discharge , the water temperature, and on the intensity of solar radiation. An example of the snow and ice clearance rate is shown in Figure 4.36. Over the total flow season, this rate is admittedly fast. Consequently the screen effect of snow and ice is of short duration once stream flow begins. This protective effect, however, is of considerable significance because its presence coincides with the period when the bulk of the annual flow occurs, thus, the presence of this valley snowpack substantially reduces the amount of geomorphic work on the clastic bed.

Along the valleys, spatial variations in snow distribution cause the protective influence to vary from site to site. It is evident that, in a reach with neither snow nor bottom ice cover, the actual geomorphic activity will equal its potential. On the other hand, as in the upper basin tributaries with an important snowpack and a low water discharge, little geomorphic work is performed because of the snow remaining in the valleys for most parts of summer (Fig. 4.37).

4.3.3.3 Destructive Effect

An uneven distribution of snow in the valleys can also increase the bed material transport potential of the running water. Such destructive effect of the snowpack is induced either by ponding or by channel shifting processes (Fig. 4.38). The destructive effect due to ponding is produced

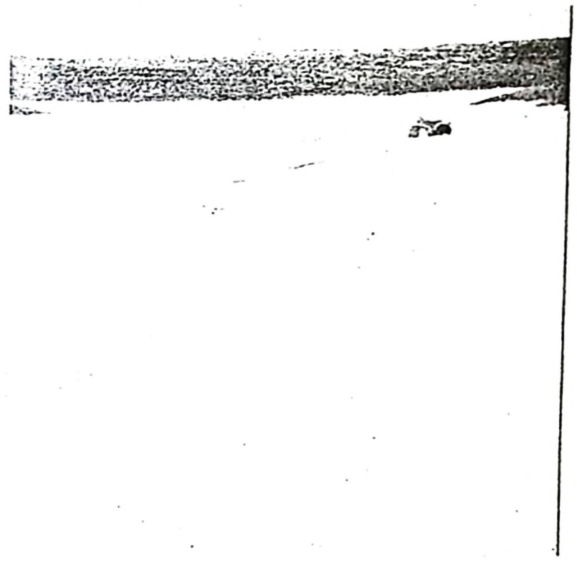


Figure 4.37 Protective effect of snowpack in upper basin tributaries, August 1977.

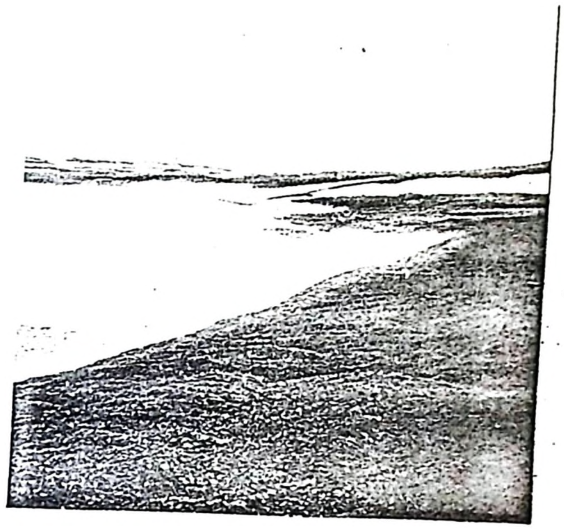


Figure 4.38 Illustration of the destructive effect created by channel shifting, where the snowpack forces the stream to flow on a bank instead of the armoured bed.

when snow dam failure releases a large volume of water at a large hydraulic gradient thus seriously affecting the channel downstream. Where a lateral shifting of the channel occurs, the stream can temporarily occupy gently sloping banks whose materials are more erodible than the armoured streambed. The stream may then be provided with a great availability of fine sediment and the flow will not have to overcome the armouring conditions prior to an initiation of movement. However these destructive effects are difficult to asses since the bed conditions charges continually as shifting progresses. Thus, the magnitude of the additional destructive effect attribuable to the presence of snow and ice can only be intuitively surmised.

4.3.3.4 Summary

The effect of valley snowpack on fluvial processes is summarised in Figure 4.39. It shows the conditions under which the snowpack provides destructive or protective effects on the streambed. Accordingly, one may determine the enhancement or hindrance of fluvial activities by the presence of snow.

The actual amount of fluvial actions performed on the bed along a given reach of snowfilled valley varies both spatially and temporally. The spatial considerations include (1) the initial conditions of the bed (i.e. whether it was frozen under wet or dry conditions in the past winter) and the imbrication of the bed material, and (2) the initial snow distribution characteristics in the reach. The temporal variability depends on the relative rates of snowmelt peak flow and the rates of valley snow cover depletion. The earlier the occurrence of the bulk of annual flow,

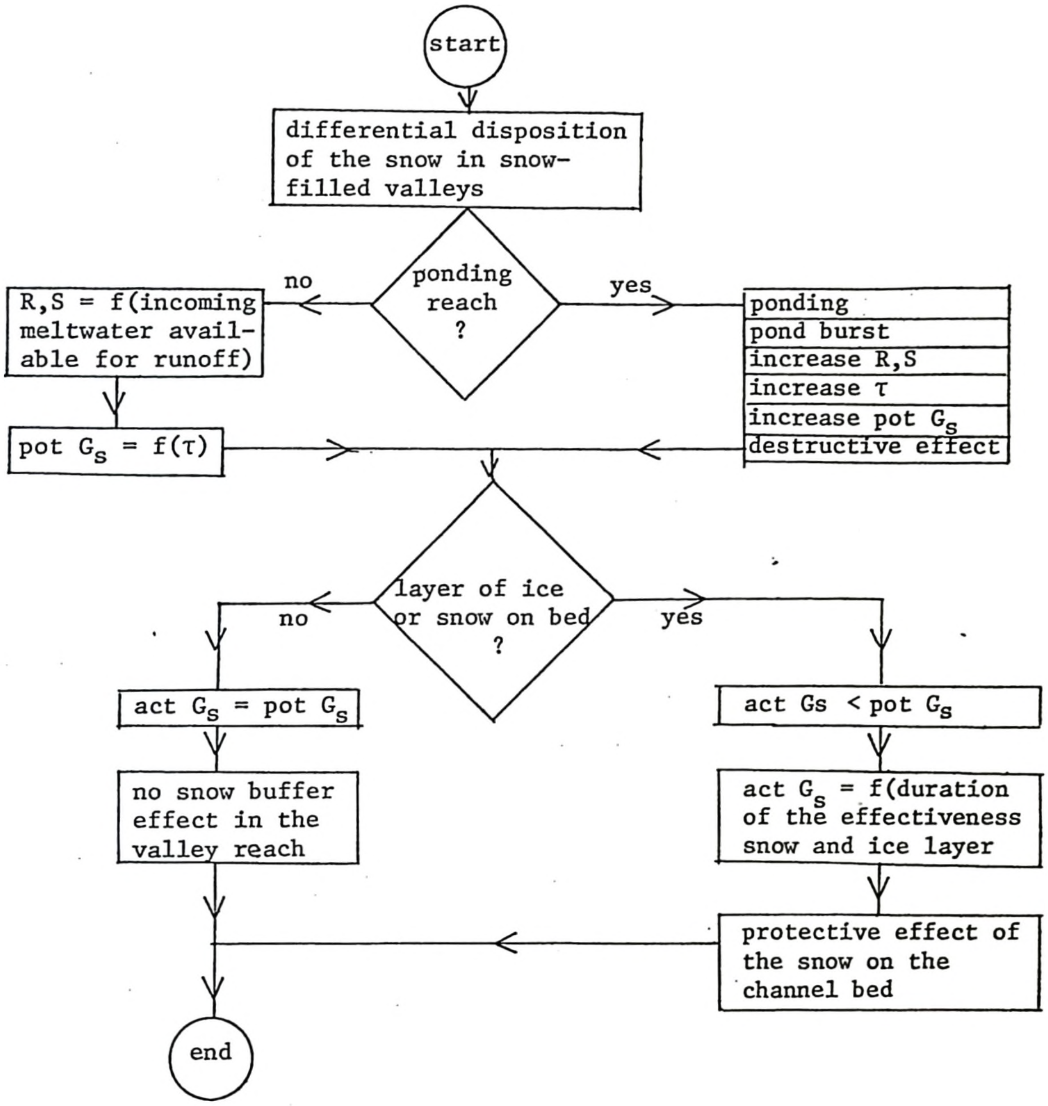


Figure 4.39 Effect of snowpack in valleys on fluvial processes.

the more snow there will be on the streambed to resist the flow and this hinders geomorphic work. In the case of a longer duration, smaller magnitude snowmelt discharge, the bulk of the flow will be extended over a longer period, and the slower will the valley snowpack depletion be by melt-erosion of the stream. If these conditions are associated with a channel shifting reach, more geomorphic work can be performed on a sensitive bank. However in the case of a thick snowpack in an incised valley, geomorphic work will be hindered for a longer period allowing a smaller yearly transport.

CHAPTER FIVE

CONCLUSION

In recent years, the significance of fluvial activities in a High Arctic environment has been acknowledged. Recent literature (St-Onge 1965, Cook 1967, McCann et al. 1972, Church 1972) provided estimates of the rates of erosion and sediment transport in high latitudes streams.

Related to an High Arctic environment, Pissart (1967) and McCann et al. (1972) reported the presence of important snow accumulation in valleys. Different views were expressed regarding a qualitative assessment of valley snowpack effects on the fluvial processes. Pissart emphasised a protective role of the snow because during a ten-day observation period, 75 percent of total annual discharge merely flowed over ice layer without eroding the bed. On the other hand, McCann et al. stressed the ephemeral status of these snow accumulations in valleys; they were stating that the snowpack depletes rapidly as flow begins.

Nowhere in the literature, however, was any quantitative measurement of the role of valley snow and ice on fluvial processes. Moreover, despite Pissart's (1967) description of the manner in which streams break-up, no comprehensive discussion was reported on the processes involved in the channel development sequences in snow-filled valleys.

It was therefore the objective of this study to determine channel development sequences in snow-filled valleys and to quantify the role of valley snowpack on the fluvial processes of High Arctic streams.

The major findings of this study are as follows:

(1) Sequences of events were established to explain channel development in terms of several processes including:

the ponding and subsequent release of water behind snow-dams formed by drifts,

the formation and collapse of snow tunnels due to subsurface flow in snow,

the vertical incision and lateral shifting of channels in the snowpack,

the abandonment of channels due to flow diversion elsewhere in the snowpack.

An attempt was also made to predict the channel sequences given the established relationship between valley shape and snow disposition characteristics.

(2) This study assessed the role of snowpack on fluvial processes. Where runoff is substantial, the rate of snowpack depletion is fast. Otherwise, as in the tributary valleys at the upper parts of the basin, the valley snowpack stayed on through out the flow season. Nevertheless, since the bulk of annual water discharge occurs while the snowpack is interposed between the running water and the bed material, little geomorphic work is performed during that time period. A protective effect in the order of 52, 31, 50, and 10 percent of the annual potential bed material load were computed for Sites 1, 2, 3 and 4 respectively. Hence this material was protected by the snowpack.

It is concluded that fluvial activity is responsible for the removal of substantial amount of materials in the High Arctic as does elsewhere. The outstanding difference is that these streams have to overcome their valley snowpack before sediment transport becomes effective.

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