INVESTIGATIONS IN A HIGH ARCTIC

DRAINAGE BASIN

GEOMORPHOLOGICAL AND HYDROLOGICAL INVESTIGATIONS IN A HIGH ARCTIC DRAINAGE BASIN

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
October 1975

MASTER OF SCIENCE (1975)
(Geography)

McMaster University
Hamilton, Ontario.

TITLE: Geomorphological and Hydrological Investigations in a High Arctic Drainage Basin.

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SUPERVISOR: Professor S.B. McCann

NUMBER OF PAGES: xiii, 233.

SCOPE AND CONTENTS:

Four aspects of the "Schei River" basin, a 91.2 km^2 glacierized catchment at the head of Vendom Fiord, Ellesmere Island (78°N, 82°W), were investigated during the 1974 runoff season. The postglacial evolution of the basin is studied through comparison of the present river system with a reconstruction of drainage at the time of deglaciation. The three month runoff regime of the basin is analysed in terms of glacial and nival contributions, response to meteorologic inputs and ice-damming and an approximate water-balance is calculated. Measurements of concentration and load of suspended and dissolved sediment are discussed in terms of controlling factors and geomorphic implications. Finally, the fluvial and glaciofluvial landforms of the basin are described, with emphasis on the cobble-size distribution of sandur deposits.

ACKNOWLEDGEMENTS

I am greatly indebted to my supervisor, Dr. S.B. McCann, for his friendship and encouragement during the period of research for this dissertation, and for his assistance and advice in the field and during the analysis of data. His generous provision of financial and material support is also gratefully acknowledged.

In any shared field project one is inevitably heavily dependent on the help of co-workers. Of these, Dr. M-k Woo contributed much in advice and discussion, and Graham Cogley gave me the benefit of his considerable expertise on Arctic fluvial processes. Bruce Bennett's excellent organisation of the Vendom Fiord base camp, his operation of the main "Schei River" gauging station and his advice on various aspects of data reduction are all warmly appreciated. Randy Crispo proved a competent and hard-working field assistant. Thanks are also due to the economic geographers amongst whom I have worked in my two years at McMaster. They provided much advice on statistical and programming techniques, and their company was a source of helpful stimulus.

The Vendom Fiord project, of which this study is a part, was funded by the Glaciology Division of the Inland Waters Branch, Department of the Environment. Additional support was made available by Dr. McCann from a National Research Council of Canada grant, and a substantial contribution towards the funding of fieldwork by the Department of Indian and Northern Affairs through the presidential committee on Northern Studies at McMaster University is gratefully acknowledged.

My wife Barbara provided many hours of assistance in reducing much of the data to its final form and in typing the first draft. Her unflagging encouragement and enthusiasm made this task much less burdensome than it might otherwise have been.

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ABSTRACT

This dissertation reports the results of investigations into the geomorphic role of fluvial processes in a high-arctic glacierized drainage basin in the summer of 1974. A review of the relevant literature (chapter 1) is followed by a statement of objectives and a description of the study area. In chapter 3 the drainage network at the time of deglaciation is reconstructed and the subsequent modifications effected by the river are shown to involve downcutting concomitant with isostatic recovery, with the resultant formation of gorges and terracing of alluvial deposits. The three-month hydrologic regime of the river and its major tributaries is described and analysed in chapter 4. Glacial meltwater is found to constitute 55% of total discharge, snowmelt 35%. Flash floods occurring on the river are explained in terms of the temporary retardation of flow at the glacier margin. In chapter 5, figures for dissolved and suspended concentrations and load are presented for the river and its major tributaries, and sediment concentration - discharge relationships are described and discussed. Sources of sediment are investigated and results compared with those obtained elsewhere in the arctic. Chapter 6 includes a description of the

erosional landforms of the basin, and a detailed analysis of cobble-size distribution over the surface of active and relict sandar from which it is concluded that cross-sandur variations in clast size may be just as significant as downsandur fining. The major conclusions of the study are restated in the final chapter.

CHAPTER 1

INTRODUCTION AND REVIEW OF LITERATURE

1.1 INTRODUCTION

Studies of high-latitude landscapes have traditionally focussed on landforms and processes which distinguish these environments from other morphoclimatic zones. As a result, knowledge of the glacial and periglacial characteristics of arctic areas has profited, but at the expense of relative ignorance as to the effectiveness of other geomorphic processes in these areas. In the past decade this bias has been partly remedied, particularly with the inception of quantitative studies of the geomorphic role of rivers and the sea in the high arctic. One of the consequences of these studies has been an increasing awareness of the importance of fluvial activity in periglacial areas, a concept emphasized strongly in a number of recent publications, notably those of Church (1972), McCann, Howarth and Cogley (1972), and McCann and Cogley (1974).

The study of the role of Arctic rivers is still at a pioneer stage, however, and in many quarters the periglacial landscapes of the arctic are still viewed as being dominated by frost and snow action, freeze-thaw and distinctive mass-movement processes. In contrast, the evidence from fluvial studies suggests that typically "periglacial" landforms represent relatively minor though distinct landscape elements, and that the physiography of most present-day arctic areas in fact reflects most strongly the twin influences of Pleistocene glaciation and postglacial fluvial modification. If this is so, the principal geomorphic influences in the arctic are not dissimilar from those affecting mid-latitude landscapes. Such a view clearly involves considerable conceptual reorientation from the traditional viewpoint of periglacial areas dominated by cold climate processes. Before any substantial re-evaluation of the geomorphic influences can be made, however, some fundamental questions on the role of rivers must be answered and the generality of results obtained to date must be confirmed. This dissertation is directed to these ends.

In the pages that follow, four aspects of arctic fluvial geomorphology are considered through the study of a glacierized drainage basin, 91.2 km^2 in area, on Ellesmere Island, Northwest Territories, Canada (figure 1.1). The first topic concerns the postglacial evolution of the basin, a hitherto neglected theme. Consideration of the distinct hydrologic regime of a high-arctic glacierized basin forms the second aspect of the study. This is followed by a discussion of sediment load, its geomorphic significance, and the factors affecting its short-term fluctuations. Finally, some of the characteristics of high arctic fluvial and glaciofluvial landforms are described. The remainder of the present chapter is devoted to introducing these topics through a review of the relevant literature.

1.2. THE EVOLUTION OF THE ARCTIC DRAINAGE NETWORK

The literature on the development of the river systems of Arctic Canada is rich in speculative generalizations but scant in detailed evidence. The present drainage pattern undoubtedly contains preglacial

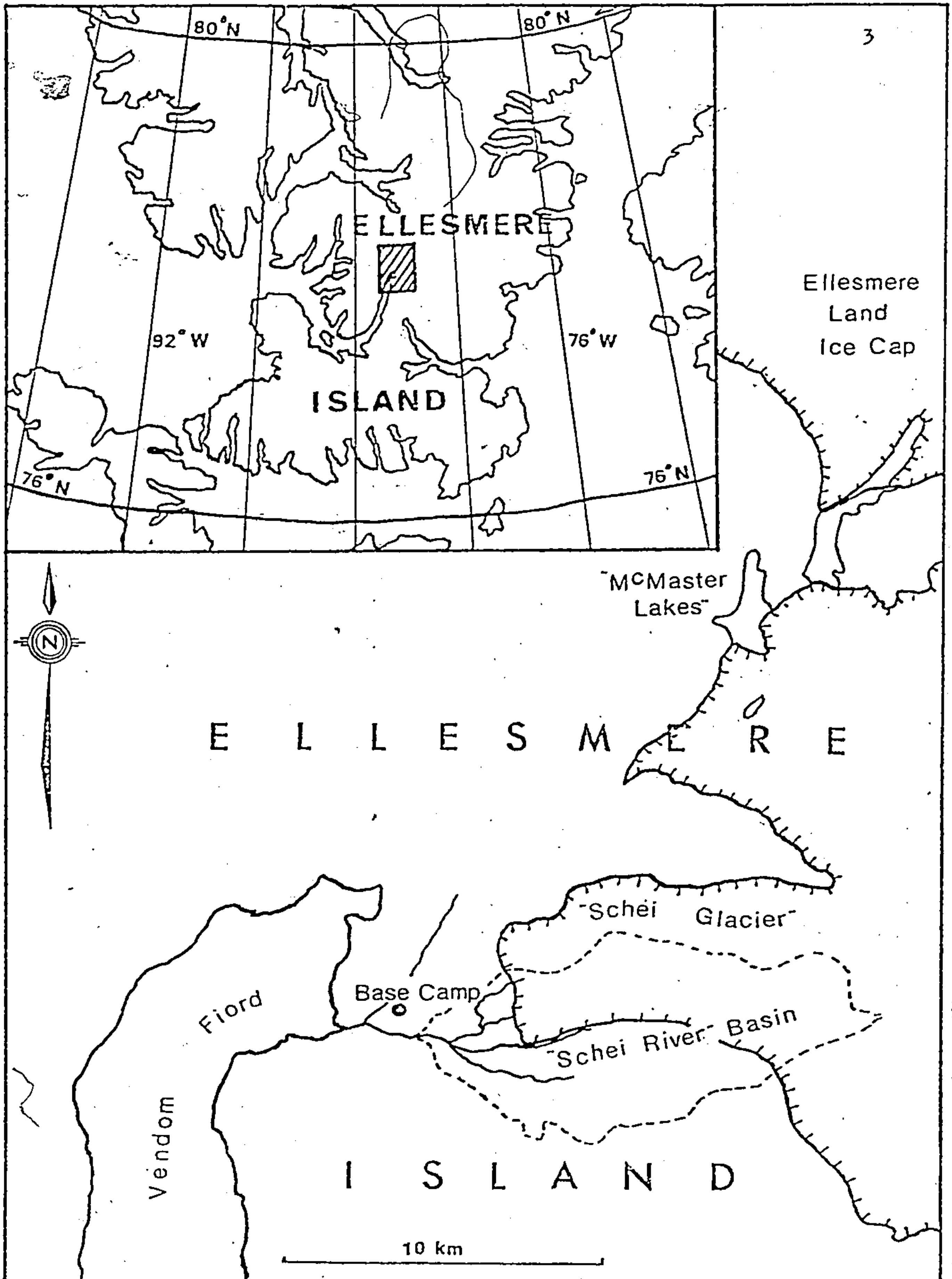


Figure 1.1: Location of the "Schei River" Basin, Ellesmere Island, 78°05' N., 85°02' W.

as well as postglacial facets, but the identification and differentiation of these elements is complicated by extensive landscape modifications effected by glaciers during the intervening Pleistocene period.

The preglacial evolution of the arctic drainage system was considered on a broad scale by Nichols (1936), who proposed that the straits separating the islands of the Arctic Archipeligo represent the present expression of a drowned glaciated drainage system. This concept was expanded by Fortier and Morley (1956), who envisaged the development, in early Tertiary time, of large consequent streams across the Cretaceous formations of the central archipelago in a north-westerly direction. Bird (1967) further elaborated this idea, suggesting that this phase was followed by the development of subsequent streams flowing in downfaulted graben such as Parry Channel. He believed that these streams were probably responsible for the formation, in the middle or late Tertiary, of the extensive peneplain known as the Barrow Surface in the central arctic archipelago, and possibly also for the deposition of the Beaufort Series further west. Bird contended that the entrenchment of major streams into the Barrow Surface and its correlatives occurred during the late Pliocene or early Pleistocene, when sea level is known to have been at a minimum. The major lines of dissection were probably enlarged and modified by the Pleistocene ice masses which covered virtually all of the Canadian high arctic except possibly areas in the extreme west (Craig and Fyles, 1960).

These ideas were adopted at a more local level by Robitaille (1959, 1960). He noted that the Barrow Surface in south-east Cornwallis Island is dissected by steep-sided valleys, and attributed the formation of these to three distinct stages of development:

> "Au Pliocène, de grandes vallées à fond plat sont inscrites dans le plateau .... à la fin du Pliocène, l'exhaussement du sol déclenche une incision rapide des thalwegs, les cours d'eau approfondissent leur vallée sans y practiquer d'erosion latérale notable. L'instauration du troisième épisode survient au cours de la phase initiale de l'emersion glacio-isostatique qüand les cours d'eau retrovent leur vallée preglaciere"
>
> (1959, p 365).

This view was not upheld by other workers in the same area. Cook (1967) described the Mecham River, Cornwallis Island, as "a small post-Pleistocene stream", and McCann and Cogley (1973) argued from evidence of recent downcutting that the Mecham River and the steep gorges of Cornwallis and Devon Islands are essentially postglacial, though possibly initiated in the Pliocene. No definite evidence of the extent of the postglacial development of high arctic streams has yet been presented, however, and the topic remains one of contention.

1.3 HIGH ARCTIC RUNOFF

1.3.1 Nival Runoff

Arctic streams which drain unglacierized catchments have a distinctive hydrologic regime. With the exception of major rivers such as the Mackenzie (Mackay, 1963) and some rivers in arctic Russia (Suslov, 1961), flow ceases completely during the winter months; in the Queen Elizabeth Islands the runoff season extends at its longest from June to September. The general pattern of flow during the high arctic runoff season involves a brief period of high discharge reflecting rapid spring snowmelt, followed by a longer period of declining discharge as the remaining snowpack attenuates. Superimposed on this pattern are diurnal discharge fluctuations resulting from systematic variations in snowmelt rate, and floods associated with rainstorms. Sparse vegetation cover and the low storage capacity of the shallow active layer cause a "flashy" response to summer rainfall. This type of regime has been classified as "arctic nival/small river" by Church (1974).

Despite the recency of studies of arctic nival runoff, a number of summary articles have already appeared. Hartman and Carlson (1970) assembled an initial bibliography of arctic water resources, later supplemented by an arctic water-balance bibliography and assessment by Dingman (1973a). For Alaska, general accounts of progress in streamflow research have been published by Childers (1971) and Kane and Carlson (1973). The sparsity of the Canadian arctic hydrological network and proposals for its expansion form the main theme of a publication by Clark and Peterson (1972), and more recently the Water Survey of Canada (1974) has provided a "historical streamflow summary" for Yukon and Northwest Territories. Significantly, the entire Arctic Archipeligo is represented in this publication by a single month's measurement on one small stream.

Of greater interest are the review papers by Church (1974) and Løken and Mackay (1974). The former classified high latitude regimes by basin size and environment (muskeg, subarctic, arctic nival, arctic proglacial) and tabulated examples of maximum runoff rates and response characteristics for each type. Løken and Mackay (1974) have presented a broad synthesis, which includes consideration of physical limnology, groundwater and water management, as well as surface runoff. In the present review a less eclectic approach is adopted, centred principally on recent studies of nival runoff in the Queen Elizabeth Islands.

The dominant elements of the arctic nival regime were first recognised by workers in the western Canadian arctic. A description of diurnal discharge fluctuations on Ellef Rignes Island was given by St.-Onge (1965), and Pissart (1967) published a largely qualitative account of runoff on Prince Patrick Island. He noted four phases in the commencement of river discharge :

1. local snowpack melt by solar radiation while ambient air temperatures remained below freezing;
2. general melt with the percolation of rainwater to the base of the snowpack and into snow-choked stream courses; refreezing and sub-nival rillwash;
3. commencement of runoff over snow and ice in streambeds; and
4. runoff over streambeds after the melt of "anchor" ice.

Pissart also noted diurnal discharge fluctuations and measured variations of 5 - 15 m^3s^{-1}. More significantly, he realised that the sparsity of vegetation and presence of a near-surface permafrost table would lead to rapid rainfall response and consequent high (and geomorphologically effective) discharges.

In the eastern Canadian arctic some pioneer measurements were made in the summer of 1959 by Cook on the Mecham River, Cornwallis Island (Cook, 1967). He identified the sources of runoff as melting snow, rainfall and seasonal thaw of the active layer, and recorded, like Pissart, some runoff before mean daily air temperature exceeded 0°C. Following the rise in mean daily temperature above freezing point, Cook noted :

> " - a period of intense melting, lasting perhaps a week, after which daily mean temperatures rise above 32°F. During this period meltwater collects on the valley floor, held back by an ice dam at the river mouth.
>
> - a period of catastrophic flow following the breaking of the ice dam and the continued rapid melting of most of the snow in the valley. This flood lasts approximately ten days and is estimated to represent about 90% of the total yearly runoff.
>
> - a period of sharply decreased flow which continues until the river freezes to the bottom. Source of runoff during this period is from melting snowpatch remnants, rainfall and melting permafrost. The period terminates before mid-September"
>
> (p 266)

The above represented the most complete description of the high arctic runoff sequence before the present decade. Unfortunately, Cook's runoff data consisted only of maximum daily velocity readings, which are a poor surrogate for continuous discharge measurements. This precluded comparison with meteorologic variables.

Since the publication of Cook's work, there has been a small flood of studies relating to high arctic nival runoff, concerning rivers on Devon and Cornwallis Islands (Cogley, 1971, 1975 in preparation; Cogley and McCann, 1971; McCann and Cogley, 1972, 1974; McCann, Howarth and Cogley, 1972), Ellesmere Island (Ambler, 1974; Walker, Lewis and Lake, 1973) and Baffin Island (Church 1972; Vieira-Ribeiro, 1975).

A pilot study of "Jason's Creek", a small (2.3 km^2) basin draining the Barrow Surface on S.W. Devon Island by Cogley (1971; Cogley and McCann, 1971; McCann and Cogley, 1972) provided quantitative support for the conclusions of St.-Onge, Pissart and Cook, and added considerably to knowledge of the behaviour of small high-latitude catchments. The geomorphic importance of the snowmelt freshet and rainstorm floods was confirmed, and lag of discharge behind radiative and rainfall inputs was found to be brief (5h). Covariance and spectral analyses of discharge and meteorologic data showed a stronger relationship between net radiation and discharge than between temperature and discharge, indicating that radiation provides a better index for snowmelt variations than temperature.

This study was followed up by hydrologic investigations in the Mecham basin, Cornwallis Island, for which a 5 year discharge record (1970 - 1974) has been obtained. Some of the geomorphic results have already been published (McCann and Cogley, 1974; McCann, Howarth and Cogley, 1972) but the main hydrologic findings have not yet appeared. Study of the water balance (Cogley, 1975 in preparation) revealed two factors of first importance : evaporation in the Mecham basin apparently accounts for a water loss that is approximately 1.0 - 2.5

times the river discharge from the basin; and precipitation is grossly undermeasured, with the measured figure possibly as low as 0.25 of the actual figure. The latter conclusion was anticipated for all northern North America by Hare and Hay (1971) and for small catchments in the Mackenzie basin by Anderson and Mackay (1974).

Preliminary results have appeared describing the discharge of the "Weir River", which drains a 29.4 km^2 unglacerized catchment at a latitude of 81°N on Ellesmere Island, the northermost study to date (Walker, Lewis and Lake, 1973; Ambler, 1974). The hydrograph for this stream conforms to the established "arctic nival" pattern, but no in-depth analysis of data has been published. At the other latitudinal extreme Vieira-Ribeiro (1975) gauged discharge on three streams in southern Baffin Island, and attempted to derive a predictive model for runoff based on degree-day measurements. These Baffin rivers also conformed to an "arctic nival" regime.

For completeness, mention should be made of some of the principal literature on arctic and subarctic runoff under the rather different climatic regimes of Alaska and the Mackenzie Basin. The hydrology of small Alaskan catchments has been studied by Dingman (1966ab, 1971, 1973b), Brown, Dingman and Lewellen (1968) and Likes (1966). In comparison with the streams of the Canadian arctic archipelfgo, small Alaskan rivers display a less pronounced spring flood and a longer lag in response to meteorologic inputs, reflecting a denser vegetation mat and deeper active layer. The result is a damping of diurnal discharge fluctuations. This is even more marked for large Alaskan streams, such as the Colville River (Arnborg, Walker and Peippo, 1966) and for the small tributary catchments of the Mackenzie system, where research into the environmental ramifications of pipeline development has prompted a number of recent studies (Anderson and Mackay, 1973, 1974; Sellars, 1973; Anderson, 1974; Jasper, 1974; Newbury, 1974). Differences in runoff regime between these regions and the Canadian arctic archipelago result from differences in environmental conditions, and give credence to the recognition of several distinct hydrometeorological regions within the arctic, as implied by Church (1974). The definitive classification of arctic hydrologic regimes awaits further measurements.

1.3.2. Glacial Runoff

The recent development of a number of deterministic and predictive models of glacial discharge, employing both statistical approaches (for example Goodison, 1972, and Jensen and Lang, 1973) and energy balance techniques (for example Derrikx and Loijens, 1971, Wendler and Ishikawa, 1974) has raised the study of glacial hydrology to a considerable level of sophistication. The scope and progress of this field can be evaluated through reference to the three-part study of glacial runoff conducted by Stenborg (1965, 1968, 1969) in Northern Sweden, the International Association of Hydrologic Sciences Publication No.107 ("The role of snow and ice in hydrology", 1972), and the International Association of Scientific Hydrology Publication No.95 ("Symposium on the hydrology of glaciers," 1973). The growing use of discharge measurements as ablation estimates (Tangborn, 1966; Meier, Tangborn, Mayo and Post, 1971; Wendler and Ishikawa, 1973) has contributed considerably to the data available on glacial runoff.

The use of stochastic techniques has been almost entirely restricted to studies of isothermal "temperate" glaciers, and the rather small body of literature on the hydrology of cold "sub-polar" and "polar" ice masses is primarily descriptive. This bias is surprising, in view of the fact that supposedly impermeable cold glaciers constitute rather simpler hydrologic systems normally with negligible englacial or sub-glacial water movement or storage.

In the Canadian arctic archipelago, the first substantive investigation of the relationships between meteorologic inputs, ablation and meltwater discharge were made by Keeler (1964) on the Sverdrup Glacier, Devon Island. A similar but more detailed study was pursued by Adams (1966) on the White Glacier, Axel Heiberg Island. Adams noted striking similarity in the behaviour of net shortwave radiation patterns and the discharge fluctuations of a supraglacial stream, and concluded that radiation was the dominant determinant of melt rate, and hence the diurnal discharge cycles that form the main characteristic of glacial runoff. He also made detailed observations on the impact of summer rainfall : in 1960, a relatively rain-free year, the maximum discharge of the ice-marginal Ermine River hardly exceeded 5.0 m^3s^{-1}, whereas in 1961 discharges greater than 20.0 m^3s^{-1} were recorded following persistent but not torrential rainfall. The presence of snowcover on the glacier surface generally led to extensive percolation of meltwater and rain into the snowpack, markedly slowing the otherwise rapid glacial runoff. For 1960 and 1961, melting ice was found to contribute 81% and 63% respectively of the total annual runoff in a 54 km^2 basin with 75% ice-cover.

On Baffin Island, measurements of discharge made on the glacial Decade River by Østrem, Bridge and Rannie (1967) exceeded the measured contributions of rainfall and ice-melt. This difference was attributed to errors in ablation measurements and overestimation of the amount of rain which froze on the glacier surface rather than contributing to run-off. Also on Baffin Island, hydrologic measurements were made near the snout of the Lewis Glacier by members of the now-defunct Geographical Branch, Department of Energy, Mines and Resources, and reported in an anonymous paper (1967). This publication stressed the close relationship between heat supply and runoff. Cumulative runoff was plotted against cumulative degree-hours in the manner decribed in Keeler (1964), and indicated the importance of the rate at which the freezing level extended upglacier and the speed with which supraglacial channels were "slushed out", until

> "... a state is finally reached in which the entire watershed has become contributary and the old ice surface is reached. Thereafter, cumulated melt proceeds in a more or less linear relation with the heat index"
>
> (p 257)

The same Lewis Glacier discharge data, together with discharge records for the glacial rivers of Ekalugad Fiord, east-central Baffin Island, were analysed by Church (1972) in his major work on arctic fluvial processes. He investigated the diurnal discharge cycle as a periodic function

$$Q_t = \overline{Q} + c_i \sin \frac{\Theta_i t}{i} \qquad (1.1)$$

where $\overline{Q}$ is the mean daily flow and θ_i are periodic phase angles. Using hourly mean flow values, he found that the main (24 hour) harmonic usually accounted for more than 90% of the total diurnal variance. As might be expected, peak flow times exhibited much less variance during seasons with relatively fine weather. Similarly, serial correlation revealed a high degree of seasonal structure during fine-weather runoff seasons, but strong within-season effects during stormy seasons. Church also provided a detailed description of the seasonal runoff sequence (breakup, summer flow, freezeback), and discussed sources of flow, the influence of meteorologic inputs and extreme hydrologic events.

The preliminary results have appeared (McCann, James, Cogley and Taylor, 1972; McCann, Cogley, Woo and Blachut, 1974; McCann, Cogley, Blachut, Ballantyne and Bennett, 1975) of investigations into the hydrologic behaviour of two glacierized basins, one of which forms the study area for this dissertation, on Ellesmere Island. Some of these results are described in section 4.1.

The presence of an ice-dammed lake in a glacierized basin may represent a considerable hydrologic hazard. Although there is a vast literature on the behaviour of such lakes (Blachut and Ballantyne, 1975 in preparation), treatment of their drainage patterns in association with a dam of cold rather than temperate ice is restricted to a monograph by Maag (1969), and a forthcoming thesis by Blachut (1975, in preparation). More pertinent to the present study are the observations by Adams (1966), Church (1972) and Wendler, Trabant and Benson (1972) on the temporary ponding of ice-marginal streams by falling ice and slushflows, although in none of the recorded cases did the breaching of a temporary ice dam result in significant flooding.

As a conclusion to this section it should be emphasized that comparison of high arctic glacial and nival runoff has not yet been considered in detail, although many supposedly "glacial" catchments are effectively only partly glacierized and actually exhibit a "glacionival" regime. The influence of a nival component was clearly recognised by Adams (1966) but has received less satisfactory treatment from other authors (Ostrem, Bridge and Rannie, 1967; Wendler, Trabant and Benson, 1972).

1.4 FLUVIAL PROCESSES AND SEDIMENT TRANSPORT

1.4.1. Introduction

McCann, Howarth and Cogley (1972) have noted the scant treatment accorded to river action in contemporary texts on periglacial geomorphology. This deficiency is surprising, in view of the fact that a number of papers published before the present decade emphasized the importance of running water in periglacial areas. As early as 1897, Tarr discussed the erosive power of meltwater on Baffin Island and Greenland. Jenness, writing about the western Canadian Arctic in 1952, devoted several pages to the work of running water and concluded that the most significant active processes are mass wasting and fluvial erosion. Robitaille (1959, 1960) counted fluvial processes as "pas les moins importants" operating on Cornwallis Island at the present time and Czeppe (1965) emphasized the efficiency of sheetwash and rillwash as agents of sediment transport on Spitzbergen.

Rudberg (1963, 1969), working on Axel Heiberg Island,was keenly aware of the effects of river action on that island :

"The influence of running water is striking in well-developed fans, outwash trains, valley trains and in some areas where erosion is responsible for V-shaped valleys, canyons, ravines and rill-sculpture."

(1963, p 139)

The effectiveness of fluvial activity he attributed to the characteristic rapid storm runoff of arctic areas. Cook (1967) was not so certain about the effectiveness of fluvial erosion; following his investigations on the Mecham River, Cornwallis Island, he concluded that fluvial action

"... is essentially a cathartic, transporting the weathered products of other processes by not, in itself, producing great amounts of erosion"

(p 267)

Maag's study of aspects of glacial hydrology on Axel Heiberg Island, carried out between 1960 and 1963 but published in 1969, represents in part an early attempt to measure the effectiveness of fluvial processes, in this case in the proglacial zone. He noted that the relatively sediment-free waters draining an ice-dammed lake eroded considerable quantities of debris from a proglacial moraine, and provided estimates of river load. His figures must be treated with caution, however, as his sampling methods were crude, and he did not state the means whereby the bedload figure was derived.

In more recent studies, measurement of stream load has replaced intuitive assessment as the dominant approach to the study of fluvial processes in arctic areas. The rationale behind such measurements is that the sediment load transported by a stream over a given time period (usually one year) provides an approximate index of denudation rate. Although such an index has little intrinsic meaning, useful comparisons can be made with similar indices obtained for other environments (Douglas, 1973). The following three sections summarise the results of such measurements.

1.4.2 Solute Concentration and Load

In 1959 J. Corbel suggested that solutional removal from carbonate terrain is more intensive in cold than in warm climates. This idea was based on the fact that the level of CO_2 in solution is greatest at low temperatures. High CO_2 levels measured in and at the bases of snow-patches by Kauko and Laitinen (1935) and Williams (1949) support this reasoning, but more recent studies have shown that the CO_2 content of surface waters is determined by a complex interaction of climatic, lithological and biological variables (Douglas, 1968), and that Corbel's ideas are subject to oversimplifying assumptions.

The results of recent studies of water hardness in the Canadian arctic illustrate some of the deficiencies of Corbel's conclusions. Smith (1969, 1972) measured water hardness at sites on Somerset Island and obtained a rather low (57 p.p.m.) mean concentration for major rivers. The mean value obtained for eleven "marsh and meadow" sites was 106 p.p.m., indicating the importance of vegetation in increasing the acidity of water, and hence its capacity to maintain larger quantities of solutes. The relatively low hardnesses obtained by Smith prompted him to conclude that solution has not been a significant factor in postglacial erosion.

Investigations of the solute contents of Arctic streams have also been carried out on Devon and Cornwallis Islands (Cogley, 1971, 1972;

McCann and Cogley, 1971). In the former area Cogley found an inverse relationship between solute concentration and water discharge for a small creek with a 2.3 km^2 drainage basin on carbonate rocks. This relationship he attributed to a "dilution" effect; during periods of high discharge, a smaller percentage of the total volume of water is in contact with the channel perimeter. The "high discharge" solute concentration for the snowmelt flood was 30% - 40% less than that measured during rainstorm floods. This is probably explained by CO_2 depletion in snowpacks, and is reflected in pH values : snowmelt pH 7.85, rainwater 6.0 and 6.6. The total (Ca + Mg) hardness values measured by Cogley in the Devon Island creek ranged from 29 - 102 p.p.m., and total seasonal dissolved load was roughly equal to, or slightly greater than, total seasonal suspended load. Cogley concluded that salts in solution formed an important component of total load; the generality of this conclusion for high arctic limestone terrain remains to be tested.

The only reported measurements for non-carbonate areas are those of Church (1972), who sampled glacial rivers on Baffin Island. Solute concentration was generally low, a function of the insolubility of the Baffin rocks (metasediments and granite gneisses). Total concentrations were generally less than 25 p.p.m., except during floods. During the 1963 snowmelt flood on Lewis River, a value of 93.4 p.p.m. was recorded, of which Cl^- and NO_3^- ions accounted for 47 p.p.m.; jokulhlaup floodwaters in the same area contained concentrations as high as 328.6 p.p.m. ($Cl^- + NO_3^-$ = 179 p.p.m.). Total seasonal solute load however constituted a very minor part of total load, and Church rightly concluded that salts in solution play only a small role in the denudation of this non-carbonate area.

1.4.3. Suspended Sediment Concentration and Load

Although a wealth of data exists on the concentration and load of suspended sediment carried by rivers in temperate regions, relatively few measurements have been reported for the Arctic. The earliest significant study was made by Arnborg, Walker and Peippo (1967) on the Colville River, which drains a large (50,000 km^2) catchment in Alaska. They reported on the size, concentration and load of material in suspension, and noted a general covariance between sediment concentration and discharge, a relationship long established for lower latitudes (Hjulstrom, 1935). The early spring floodwaters of the Colville River had low concentrations of suspended inorganics, as melting snow and ice contributed very little silt-size material and the tundra had not yet started to thaw. During the ice-breakup flood, however, concentrations of 1,500 mg l^{-1} were recorded. Flow velocities of up to 2.0 ms^{-1} and high (14°C) water temperatures caused extensive bank collapse at the time of the breakup flood (Walker and Arnborg, 1966) with the consequent supply of abundant fine material to the river. The highest recorded concentration (1,650 mg l^{-1}) occurred immediately downstream from an area of bank collapse. 62% of the total annual load was carried downriver in the 13 days of the breakup period.

The concentration and load of suspended sediment in streams draining much smaller non-glacierized arctic basins has been studied on Devon and Cornwallis Islands (Cogley, 1971, 1975 in preparation; McCann, Howarth and Cogley, 1972; McCann and Cogley, 1974). As a result of investigations on "Jason's Creek", which drains a 2.3 km^2 basin on S.W. Devon Island, Cogley (1971) concluded that suspended sediment concentration levels in the high arctic are of the same magnitude as those of lower latitudes. He also found covariance of water discharge and suspended sediment concentration, but noted that this relationship was subject to the vagaries of sediment supply; a mudflow moving into the stream had the effect of raising the concentration from 2 mg l^{-1} (above) to 158 mg l^{-1} (below).

For the Mecham River, which drains a 95 km^2 catchment near Resolute, Cornwallis Island, McCann, Howarth and Cogley (1972) reported high suspended sediment concentrations at the onset of the snowmelt flood. This was problematical as the snowpack source appeared "clean". They concluded that at some stage meltwaters must flow beneath the base of the snowpack, entraining sediment en route to the river. In general, concentrations of suspended sediment measured in the Mecham were low; the 1970 measured maximum was less than 600 mg l^{-1}.

Østrem, Bridge and Rannie (1967) have reported the results of suspended sediment observations made on the glacial Decade River on Baffin Island. Measured concentrations during the study period rarely exceeded 1,000 mg l^{-1}, although 60% of the total silt discharge during the observation period occurred on a single day of high discharge following rain. This paper also contributes a useful discussion of the significance of sampling time: little change in concentration was observed during the first half of any day, but after 1200 h large fluctuations occurred. From the results of intensive sampling over 24 h periods it was calculated that the mean concentration value for the second half of the day was 60% of the observed value.

Church (1972), also working in the proglacial zone on Baffin Island, recorded "normal" high flow values of 500 - 1000 mg l^{-1}, with a maximum of 4,800 mg l^{-1} recorded during storm runoff at Ekalugad Fiord. Even this maximum value is considerably smaller than values obtained from low-latitude glacial streams. Fahnestock (1963) measured concentrations up to 17,200 mg l^{-1} at Mt. Rainier, Washington State, and Mathews (1964) obtained 8,000 - 10,000 mg l^{-1} at the snout of the Athabasca Glacier. Church considered the hardness of the Baffin rocks to be partly responsible for the relatively low values he obtained, noting that the discharge of wash load is limited by supply rather than by stream competence or capacity.

Multivariate analyses of suspended sediment data revealed that suspended load essentially followed the relationship

$$C_{sus} = k\ (Q,\ P) \tag{1.2}$$

where C_{sus} is suspended sediment concentration, Q is water discharge, P is number of days since last precipitation plus one, and k is a constant. Suspended sediment discharge (Q_{sus}) was found to vary approximately with the square of discharge

$$Q_{sus} \propto Q^2 \tag{1.3}$$

implying near linearity in the response of suspended sediment concentration to changes in discharge.

Although the availability of fine sediment has been acknowledged in all of the abovementioned papers as a factor of prime importance, the nature of sediment sources in the arctic has received little attention. Malaurie (1954) observed in Greenland the importance of mass-wastage in supply fine material to rivers, and Czeppe (1965) emphasized the role of rill-wash in transporting fines to the rivers of Spitzbergen. Wilkinson (1972), working in the Canadian Arctic, demonstrated that cryoturbation released fine materials to the rill system; even on low-angle slopes high (up to 700 mg l^{-1}) silt concentrations were obtained from rills, indicating their importance in supplying fine sediment to arctic rivers.

Church (1972) emphasized the importance of glacial debris as a sediment source :

> "... whilst Baffin watersheds appear, from measurement of present rates of sediment transport, to behave like remarkably high energy environments, it is not clear that they would continue to exhibit such behaviour after substantial completion of postglacial redistribution of materials..."
>
> (pp 63, 64)

This condition is not necessarily true for all of the arctic archipelago. A comparative lack of glacial debris on S. Cornwallis Island may explain the relatively low suspended sediment concentrations measured by McCann, Howarth and Cogley (1972) for the Mecham River.

1.4.4 Bedload

Again for the Mecham River, McCann, Howarth and Cogley (1972) considered that bedload was the most important component of total stream load. Cogley (1971) reached a similar conclusion for "Jason's Creek", S.W. Devon Island, and the ubiquity of coarse clastic valley fills throughout the arctic archipelago bespeaks the dominance of bedload transport. Unfortunately, the difficulty of assessing bedload (Hubbell, 1964; Yalin, 1973) has deterred its accurate evaluation. Only Church (1972) has attempted bedload measurement in an arctic area. This he did in three ways : by placing size-graded pebbles, cobbles and boulders in the stream and inspecting for removal at intervals; by the use of a bedload sampler; and by a computational formula. Comparison of the results of the first two methods emphasized strongly the difference in stream competence when material is set in exposed positions on top of the bed as opposed to being incorporated in the stream deposits. Church eventually resorted to the use of the Meyer-Peter formula to measure bedload. Again, however, availability of material for transport proved an important factor. Church acknowledged that the Baffin streams frequently flow underloaded, so that the computed bedload values represented potential transport; actual bedload transport must have been considerably less.

Even so, Church's results indicated the dominance of the bedload component. He did not suggest that this is typical for arctic rivers, although it seems to typify sandur streams.

1.5. FLUVIAL AND GLACIOFLUVIAL LANDFORMS

1.5.1. Erosional Forms

The most striking legacy of fluvial erosion in high arctic areas has been the formation of deep, steep-sided gorges. Such features were first described by Jenness (1952) in the western Canadian Arctic, and on

Cornwallis Island Robitaille (1959, 1960) noted the abundance of "vallées, courtes et rectilignes ... du type canon" and attempted to explain their existence in terms of a Pliocene drainage network. Gorges and ravines have also been described on Ellef Rignes Island and Axel Heiberg Island by St-Onge (1965) and Rudberg (1963, 1969) respectively. McCann and Cogley (1973) ascribed the development of the "V-shaped gorges" of Cornwallis and Devon Islands to fluvial entrenchment during isostatic recovery, but presented inconclusive evidence. It is surprising, in view of the ubiquity and impressive appearance of such gorges, that they have not been subject to more detailed research.

Probably the only small-scale erosional feature unique to permafrost areas is the thermo-erosional niche (Walker and Arnborg, 1966; Cooper and Hollingshead, 1973). This form develops when river waters melt a lateral slot, sometimes of considerable depth, into the permafrost or ground ice underlying a bank of unconsolidated deposits. Eventually, large blocks of frozen material slump into the river. Maag (1969) made a detailed study of this form on Axel Heiberg Island. He found niche undercuts of up to 9 m, and observed that niche development could be halted by the sealing of an undercut through deposition during periods of low flow. Bank retreat through the collapse of overhangs averaged 1.2 m per season, even though water temperatures did not exceed 5.0°C. This appears to indicate that rapid bank erosion is to be expected in areas of rapid thermal undercutting.

1.5.2 Depositional Forms

Unquestionably, the most significant landform of fluvial/glaciofluvial deposition in the high arctic, from the point of view of both ubiquity and extent, is the sandar or outwash plain. Such features have been described by Church (1972) as

> "... gently sloping surfaces (usually less than 5° gradient) of coarse clastic sediment ... usually characterised by rapidly-shifting, generally aggrading, braided streams, with high competence and high bed-load sediment transport rate, rapidly changing sites of local erosion and deposition, and, consequently, rapidly changing alluvial stratigraphy."
>
> (p 3)

Sandar form the most prominent recent features of the high arctic landscape not directly due to ice action.

Although a large number of descriptions of glacial outwash deposits appeared in the early literature (Church, 1972, p 4-5), the first measurements of sandur processes were not reported until 1939, when Thorarinsson published the results of his glaciological research on Hofellsjokull in southern Iceland. Part of the important postwar Scandinavian contribution to the study of glaciofluvial geomorphology included a follow-up to Thorarinsson's work in the form of a compendium of ten papers on different aspects of the Hofellsandur, published by Hjulstrom, Arnborg, Jonsson and Sundborg (1954). A notable contribution to the study of braiding was also made by Sundborg (1956) in his study of the River Karälven. The work of another Scandinavian, Krigstrom (1962) provides a starting point for the discussion of recent contributions in this field.

The formation of landforms consisting of coarse clastic alluvium is not restricted to proglacial and periglacial areas. The requisite environmental factors appear to include :

1. an abundant supply of coarse material;
2. a river regime characterized by frequent floods; and
3. a steep valley floor conducive to the transport of coarse bed material.

Such conditions also pertain in certain arid and semi-arid (Doeglas, 1962) and mountain (Church, 1972) environments, and some important contributions to the study of braiding and coarse clastic deposits have been made in areas remote from the present periglacial/proglacial realm (for example : Doeglas, 1951, 1962, Blissenbach, 1952, 1954; Chien, 1961; Bluck, 1964, Allen, 1965; Hooke, 1967; Smith, 1970). It is generally accepted, however, that the denomination "sandur" refers exclusively to glacial outwash features.

Krigstrom (1962) distinguished between valley sandur or valley trains (dalsandar) and plain sandar (slattlandsandar). Apart from the aforementioned Scandinavian studies and work by Price (1969, 1971) on sandur development, destruction and kettling in Iceland, and Boothroyd's (1970, 1972) sedimentological studies on Alaskan sandar, recent investigations have concentrated on the former type. Krigstrom himself contributed a discussion of the formation, stratigraphy and classification of bar deposits, topics that have formed the primary concern of a large body of recent North American literature : Ehrlich and Davies (1968), Williams and Rust (1969), Boothroyd (1970, 1972), McDonald and Banerjee (1970, 1971), Rust (1972), Gustavson (1974), Hein (1974) and Smith (1974) have all approached the study of sandar from a sedimentological viewpoint.

Three important general conclusions can be drawn from these studies :

1. mean and maximum clast size tend to decrease down-sandur;
2. gravel facies predominate; and
3. sandur bar deposits normally exhibit crude horizontal bedding.

A more comprehensive study was made by Fahnestock (1963) who also discussed hydrology, hydraulic geometry, flow characteristics, sediment transport and sources and morphological change with reference to the White River sandur, Mt. Rainier. Patterns of braiding and the morphology of Alaskan sandar formed the concern of papers by Fahnestock (1969) and Bradley (1973); Smith (1973) considered rate of aggradation of sandur deposits in Alberta; and Bradley, Fahnestock and Rowekamp (1972) have discussed the transport of coarse sediment on Knik River sandur in Alaska. Interest in the characteristics of sandar is clearly expanding rapidly.

In the Canadian arctic, St-Onge (1965) has described and mapped dalsandar on Ellef Rignes Island, and a recent study by Bennett (1975) considered aspects of process, morphology and deposits on a small sandur on Ellesmere Island. By far the most important study carried out for any environment was that made by Church (1972) on several sandar on Baffin Island. Five major conclusions emerged from Church's work, namely :

1. snowmelt, glacier melt and rainstorm floods are all effective in generating bedload movement;
2. bedload constituted the most important component of total load;

3. velocity accomplished a major hydraulic adjustment to discharge increase, indicating bed form change associated with increasing sediment transport and a decline in flow resistance;
4. long profiles tend to parabolas, and competence declined downsandur; and
5. sandar are aggradational features, but with little internal structure on account of the rapid succession of erosional and depositional events.

Comparison of active sandar with Pleistocene relicts, as attempted in the present dissertation, has been carried out on a very broad scale by Frodin (1954), and a stratigraphic comparison has been presented by McDonald and Banerjee (1971). However, the potential for this type of comparison remains largely unexplored.

1.6 SUMMARY

The evidence presented in the above review of literature argues strongly for the importance of fluvial activity in arctic periglacial environments, but also indicates substantial gaps in our understanding of the evolution, hydrology, processes and landforms of arctic rivers. The aim of this dissertation is to examine these four elements in the context of the partly-glacierized "Schei River" basin. Of necessity, this investigation was selective. Where possible, hitherto unresearched aspects of arctic rivers were examined, although some of the results obtained proved directly comparable with those of earlier studies. Such results are valuable for establishing the degree of generality of earlier findings.

The evolution of the "Schei River" system was investigated through the study of lateglacial and postglacial landforms, in order to establish the extent to which the system is of late Pleistocene/Holocene age, an approach not previously attempted in high arctic areas. The hydrologic research is centred around comparison of the runoff regimes of glacial and nival tributaries, and their influence on the hydrologic behaviour of the "Schei River", again an untried approach. As indicated above, the nival component of runoff from a glacierized basin has often been ignored. The results of the present study allowed some evaluation of the importance of the nival component, and yielded considerable additional information on the nature of high-arctic glacial and nival stream regimes. The sediment transport observations are comparable with those made in earlier high arctic studies, and a particular attempt was made to explain variations in the concentration and load of suspended and dissolved and materials and to define the nature of sediment supply in the proglacial zone. The landform study concentrated on a little-researched aspect of the sedimentology of present-day and relict sandar.

This selection of topics leaves many potential areas for research untouched. It is hoped, however, that the results presented below will contribute in some measure to the growing but very incomplete body of work on arctic fluvial processes to bring about a fuller understanding of the role of the river in the arctic landscape.

CHAPTER 2

CHARACTERISTICS OF THE STUDY AREA

2.1 INTRODUCTION

The river basin under investigation is located near the head of Vendom Fiord, Ellesmere Island, Northwest Territories, Canada, at latitude 78°02'N, longitude 82°05'W (figures 1.1 and 2.1). The "Schei River", which drains this 91.2 km^2 catchment, has been the subject of scientific investigations since 1972 as part of a McMaster University programme of hydrologic research. Some of the findings of this programme appeared in a number of reports (McCann et al, 1972, 1974, 1975, Woo, 1975a) theses (Bennett, 1975; Blachut, 1975 in preparation) and papers (Ballantyne, 1975; Cogley and McCann, 1975; McCann, 1975; Woo, 1975b). A further season of research in the area is planned for the summer of 1975, after the completion of this dissertation.

2.2 GEOLOGY

The geologic sequence in the basin of the "Schei River" consists of an unconformable series of sedimentary formations of widely-differing ages overlying a Precambrian basement of gneisses, granites and migmatites which does not outcrop in the study area. The sedimentary sequence at the head of Vendom Fiord was first investigated by Norris (1963), as part of the "Operation Franklin" project, and the major units have since been mapped by Thorsteinsson (1972) at a scale of 1 : 250,000. The following

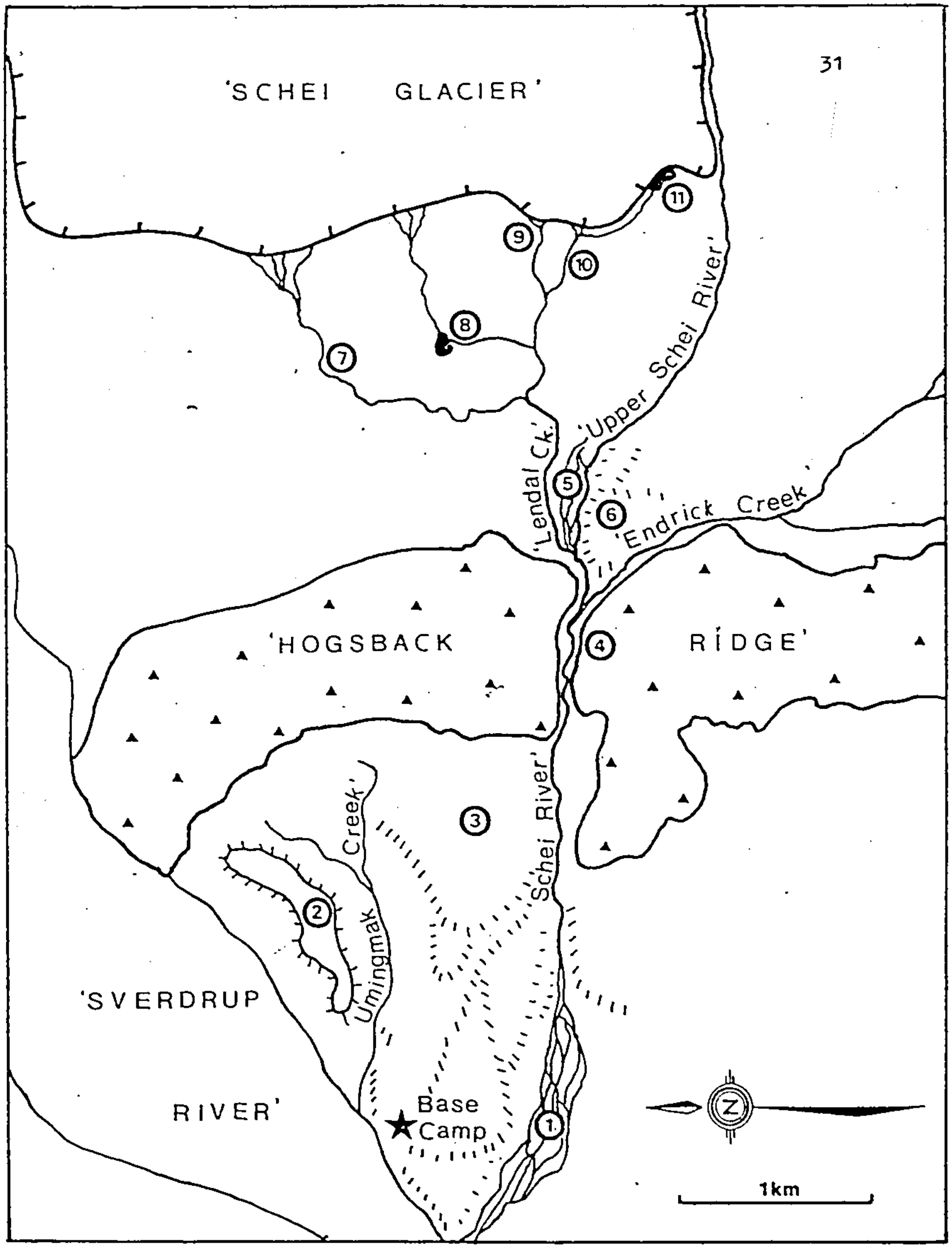


Figure 2.1: Features of the lower "Schei River" basin.

1. "Schei Sandur"
2. "Sverdrup Moraine"
3. High (T_{1X}) terrace
4. "Schei Gorge"
5. "Upper Schei Sandur"
6. "Endrick" relict sandur
7. "North Lendal Creek"
8. "South Lendal Creek"
9. "Thistle Creek"
10. "Cave Creek"
11. Ice-margin pond

account of the stratigraphy is largely based on these sources.

The lowest members in the sequence belong to the middle Ordovician Cornwallis group and consist of the limestones and interbedded shales of the Irene Bay Formation and the limestones of the Thumb Mountain Formation. These rocks outcrop extensively in the eastern third of the Schei Basin (figure 2.2) and probably underlie much of the glacierized portion of the catchment. They are conformably overlain by the dolomites and sandy and shaley limestones of the Allen Bay and Read Bay Formations, which were uninterruptedly deposited throughout the Silurian period. The lowermost members date from the upper Ordovician; the uppermost from lower Devonian time. The "Hogsback Ridge" which rises to c 300 m in the lower part of the basin is composed of these relatively resistant strata, which also outcrop in the central part of the basin. Unconformably overlying the Allen Bay and Read Bay rocks are the sandstones, siltstones and conglomerates of the lower or early middle Devonian red-bed Vendom Fiord Formation, (Kerr, 1967) which outcrops both in the central part of the "Schei" basin and on either flank of the "Hogsback Ridge".

Mesozoic strata such as those found west of Vendom Fiord are entirely absent from the basin so that the geologic column from the lower Devonian to the Tertiary is unrepresented. The topmost member of the local sedimentary sequence is the Tertiary Eureka Sound Formation, a heterogeneous and poorly-bedded assemblage of shales, sandstones, limestones, lignitic coal and lithified wood exhibiting distinct nonmarine facies and described as "soft" and "incompetent" by Norris (1963).

The sedimentary formations describes above are widely overlain by

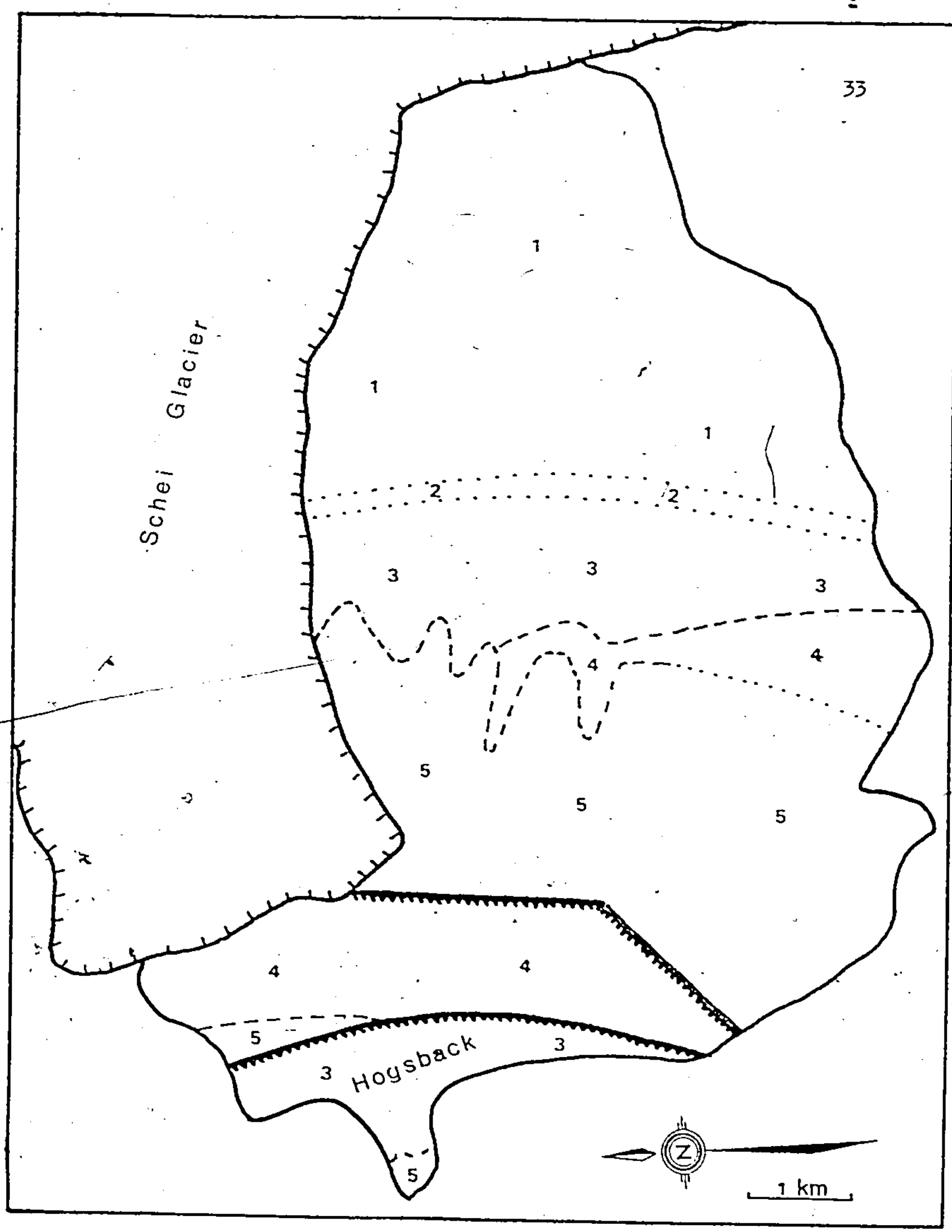


Figure 2.2: Bedrock Geology of the "Schei River" basin.
1 - Thumb Mountain Formation; 2 - Irene Bay Formation; 3 - Allen Bay and Read Bay Formations; 4 - Vendom Fiord Formation; 5 - Eureka Sound Formation.
Jagged lines represent thrust faults with the teeth on the upthrust side.

(after Thorsteinsson, 1972)

Quaternary deposits, particularly till, outwash and alluvium. The distribution, nature and significance of these deposits are discussed in chapter 3.

The "Schei" basin-Vendom Fiord area, in common with much of Ellesmere Island, displays a complex structural history (Thorsteinsson and Tozer, 1960). As part of the Arctic Lowland and Franklin Miogeosynclinal structural provinces the area was subject to severe faulting and folding during the late Devonian Ellesmerian Orogeny. The tilting and upthrusting of the strata comprising the "Hogsback Ridge" are probably attributable to this event. During the Tertiary Eurekan orogeny extensive thrust- and normal-faulting occurred, resulting in local stratigraphic displacement.

2.3 TERRAIN

35.5% of the "Schei River" basin above the main gauging site is covered by the "Schei Glacier", an outlet lobe of the Ellesmere Land Ice Cap. The main characteristics of this glacier have been described in McCann et al. (1974), the principal feature of hydrologic interest being the impermeable nature of the apparently "sub-polar" ice. In common with most other high latitude glaciers, free water movement is confined to the glacier surface or margins. There is no evidence of an englacial or subglacial drainage network except at the glacier margin, where meltwater was observed to flow through the ice for a few metres before leaving the glacier.

Comparison of photographs taken in 1972, 1973, and 1974 indicates that the "Schei Glacier" has advanced up to 5 m in the course of two years.

This evidence is supported by the discovery of recent vegetation over-ridden by the ice, and by the development of new meltwater channels in the till parallel to the west margin of the glacier. Examination of E.R.T.S. imagery for 1974 (frame number 1760 - 18013, band 6, August 22) indicated that the firn line lies outside the area drained by the "Schei River", so that all of the glacierized portion of the catchment lies in the ablation zone of the Ellesmere Ice Cap.

The upper reaches of the unglacierized portion of the basin consist of an essentially undissected plateau, between 400 m and 700 m in elevation, which rises gradually eastwards towards the ice cap. This extensive surface may represent a Tertiary remnant corresponding to the Braskeruds Plain, 16 km further north (Hodgson, 1973). West of the "Schei Glacier" front the basin is mantled by till and large areas are covered by relict sandur deposits, spreads of coarse gravels which apparently relate to the waning of the Pleistocene Innuitian ice-sheet. Further west still the basin narrows as the "Schei River" enters the "Schei Gorge", cutting through a hogsback ridge of Allen Bay and Read Bay strata.

Permafrost underlies all of the ice-free terrain, and active layer depths do not normally exceed 50 - 70 cm. Vegetation on the plateau consists mainly of mosses and lichens; in the area west of the glacier grasses, dwarf willow and various flowering plants are common; the "Hogsback Ridge" is barren.

2.4 CLIMATE

Winter circulation in the Canadian arctic archipelago is dominated by the presence of an anticyclonic system which initially develops in the western arctic then expands eastwards. This anticyclone is responsible for the cold, dry weather which persists throughout the arctic winter, and does not begin to weaken until May, when eastward-moving depressions adopt a more northerly track. Relatively few of these depressions pass north of the Parry Channel (Meteorological Branch, Department of Transport, 1970), but those that do are responsible for a significant part of total annual precipitation. Relatively high precipitation is also characteristic of September and October, when the passage of cold air from the Arctic Ocean and Beaufort Sea triggers atmospheric instabilities. As a result, the autumn months are stormy until the winter anticyclone becomes re-established.

The dominance of summer and autumn precipitation is clear from the mean precipitation records of the meteorological stations nearest to the study area (figures 2.3 and 2.4), those at Eureka (450 km distant) Dundas (750 km), Resolute (945 km) and Isachsen (970 km). The record for the coastal Eureka station (80°00'N; 85°56'W; elevation 2.4 m) is probably the most representative for the Vendom Fiord area. Eureka has the lowest mean annual precipitation for any station in Canada with 58.4 mm (Dundas 105 mm; Resolute 136 mm). 49.6% of the mean annual total precipitation at Eureka falls as rain during the months of June, July and August, the remainder accumulating as snow or ice throughout the winter then melting in late June or early July. The 100 mm mean annual total precipitation isohyet passes south of Vendom Fiord, so an average annual precipitation total of 80 - 100 mm seems not unreasonable for the study

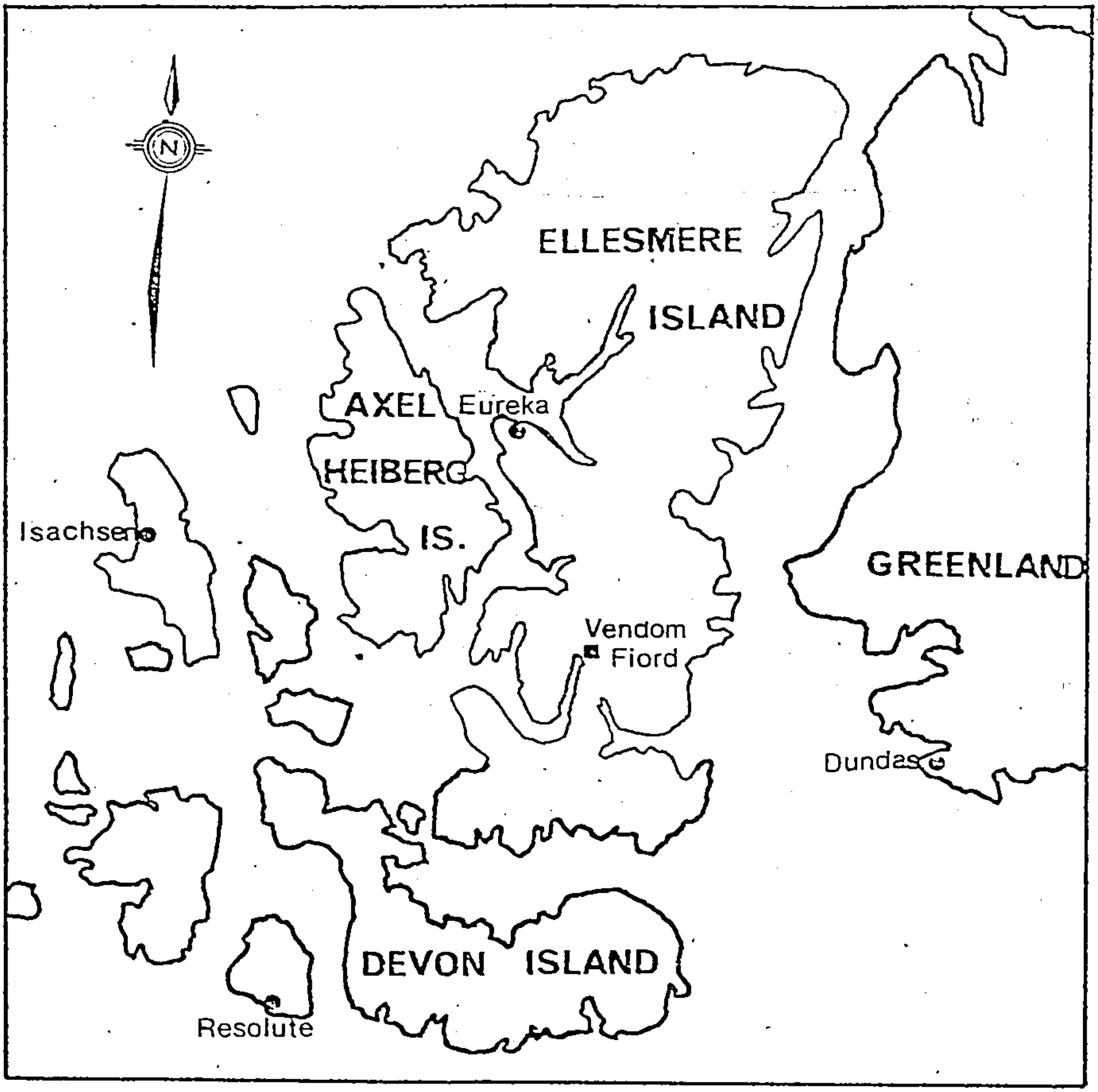


Figure 2.3: Location of nearest weather stations.

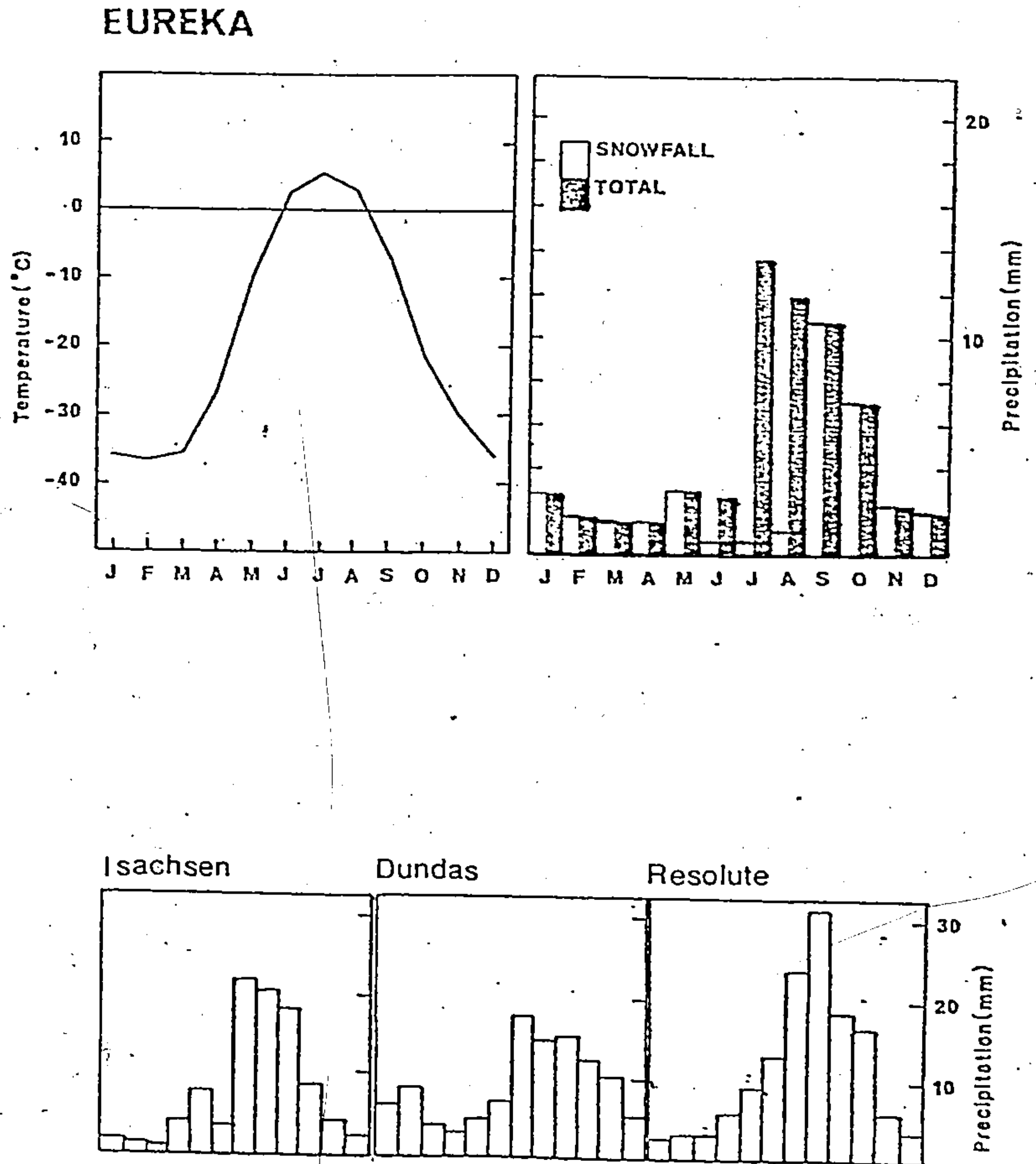


Figure 2.4: Mean monthly temperature and precipitation figures for Eureka; mean monthly precipitation figures for Isachsen, Dundas & Resolute.

area. This figure may require adjustment in light of the recent work by Cogley (1975, in preparation) which suggests that the Resolute total represents gross undermeasurement.

At Eureka, only one major rainfall event has been recorded (41.7 mm in 24 h, 68% of the annual total), but in addition a dramatic rainstorm in the Vendom Fiord area has been documented for July 21 - 23 1973 by Cogley and McCann (1975). 54.66 mm of rain fell in this period, with 49.4 mm falling in 24 h on July 22. Stream runoff exhibited a striking response to this cyclonic storm.

The arctic summer at 78°N is extremely short (figure 2.4). Mean daily temperatures at Eureka rise above freezing in June then sink below freezing in September. Geomorphic and hydrologic activity are largely confined to the intervening period.

CHAPTER 3

LATEGLACIAL HISTORY AND THE EVOLUTION OF THE "SCHEI RIVER" BASIN

3.1 INTRODUCTION : THE DEGLACIATION OF THE EASTERN CANADIAN ARCTIC

The aim of this chapter is to evaluate, through an interpretation of the lateglacial history of the study area, the postglacial development of the "Schei River" system. The general pattern of deglaciation for the eastern Canadian Arctic is described, and an attempt is made to correlate the evidence for two distinct readvance stages in the Vendom Fiord area with glacial events elsewhere in the Arctic. This permits the approximate dating of a number of lateglacial and postglacial landforms, and the establishment of a time scale by which postglacial fluvial activity may be evaluated.

During the Wisconsin glaciation, the eastern Queen Elizabeth Islands nourished an independent ice sheet, alternatively referred to as the "Ellesmere-Baffin glacier complex" (Craig and Fyles, 1960) or the "Innuitian ice-sheet" (Blake, 1966). At the Wisconsin maximum this ice-sheet covered all but the westernmost islands (Porsild, 1955; Bird, 1967) and was contiguous with the Greenland ice-sheet to the east and the Laurentide ice-sheet to the south. Taylor (1956) suggested that Greenland ice over-rode the Ellesmere Ice-shed to erode the deep fiords of western Ellesmere, but Smith (1961) concluded from field evidence that this was unlikely.

A radiocarbon isochrone map of the disintegration of the Laurentide ice-sheet (Bryson, Wendland, Ives and Andrews, 1969) reveals rapid decay of the Innuitian ice-mass between 11,000 and 9,000 B.P., with the virtual extinction of glaciers west of Axel Heiberg Island and north of Parry Channel by the latter date. Thereafter retreat was slower. Radiocarbon age determinations obtained for western Ellesmere Island (Dyck and Fyles, 1964, 1965; figure 3.1) give minimum ice withdrawal dates of 8710 ± 140 ^{14}C years B.P. (GSC - 254) for Blue Man Fiord, 8480 ± 140 (GSC - 244) for Bauman Fiord and 7750 ± 150 (GSC - 175) for Strathcona Fiord. These dates indicate a retreat of only 100 km in more than 1,000 ^{14}C years. (Hodgson, 1973).

On Baffin Island, an extensive moraine system which apparently marks the limit of an important lateglacial readvance or stillstand has been mapped. Christened the Cockburn stade (Ives and Andrews, 1963), this limit has subsequently been correlated with moraine systems in Labrador, Keewatin and N.E. Manitoba (Falconer, Andrews and Ives, 1965; Falconer, Ives, Loken and Andrews, 1965). Significantly, all the relevant radiocarbon dates obtained within the Cockburn limit are younger than 8,000 B.P. (Bryson et al., 1969) and a date of circa 8,200 B.P. is now accepted for the Cockburn stade (Andrews, 1970). The Wisconsin relict ice mass on Baffin Island, now represented by the Barnes Ice Cap, is also postulated to have readvanced around 6,700, 4,700 and 2,800 B.P. (Andrews and Webber, 1964; Andrews, 1966; Loken and Andrews, 1966) and in neoglacial times, 300 years ago. No equivalent substages have been recognised in the eastern Queen Elizabeth Islands, with exception of a readvance, possibly corresponding to the Cockburn stade, proposed for west-

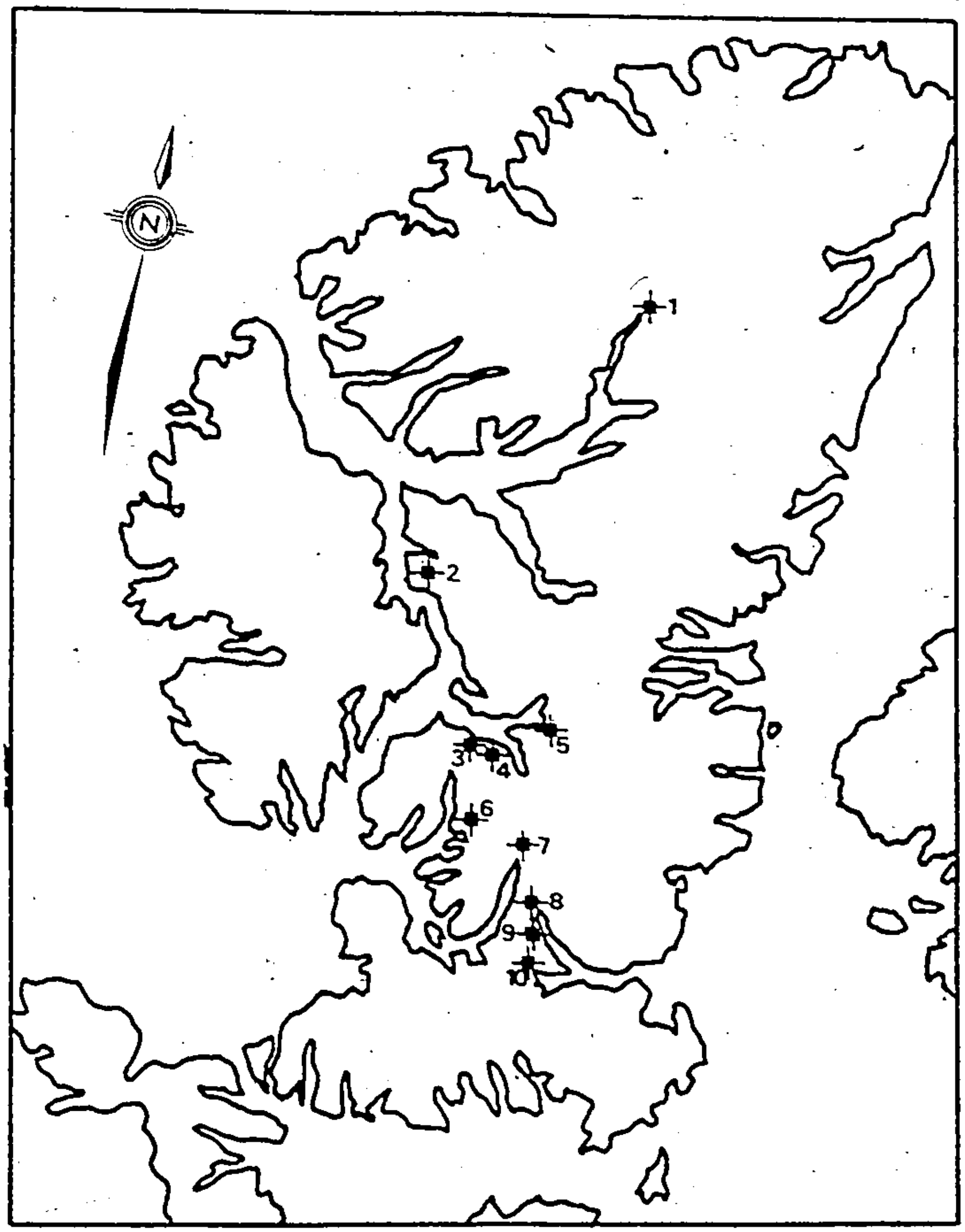


Figure 3.1: ^{14}C determinations for west-central Ellesmere Island

1. 6480 ± 200 (SI-468)
   6320 ± 120 (GSC-373)
2. 8710 ± 140 (GSC-254)
3. 7750 ± 160 (GSC-170)
4. 7680 ± 150 (GSC-175)
5. 6980 ± 80 (GSC-1858)
6. 8480 ± 140 (GSC-244)
7. 6370 ± 100 (GSC-118)
8. 7310 ± 80 (GSC-1972)
9. 2060 ± 50 (GSC-1836)
10. 8200 ± 200 (GSC-146)

central Ellesmere Island by Hodgson (1973). This situation may reflect a different response to short-term climatic deterioration under conditions of greater aridity, or merely the paucity of Quaternary studies north of Parry Channel.

Concomitant with deglaciation, all Arctic Canada was subject to pronounced isostatic rebound, which continues, at a much reduced rate, to the present day. In the eastern arctic the geomorphic legacy of uplift takes the form of flights of raised beaches, terraced deltas and perched outwash deposits. Although the postglacial uplift of Arctic Canada has been intensively studied (Andrews, 1970), few attempts have been made to relate elevated shorelines, terraces and the evidence for glacial stages to create a unified account of lateglacial events.

3.2 LATEGLACIAL HISTORY OF THE VENDOM FIORD AREA

Although observations on the Quaternary history of the Vendom Fiord area have been made by Norris (1963) and McCann et al (1972, 1974), the only detailed interpretation of lateglacial events is that of Hodgson (1973). Hodgson asserted that ice had re-occupied the "Sverdrup Sandur" (figure 3.2) at some period following the retreat of the Innuitian ice sheet. He based this conclusion on the evidence of marginal meltwater channels west of the "Sverdrup Sandur". These apparently drained through an impressive spillway (location 1 on figure 3.2) at 90 m a.s.l. The downstream end of this spillway adjoins a perched outwash delta (2) with its apex at 77 m and its distal margin at 66 m. The highest surface of an outwash deposit to the south (3) was graded to 70 m, and a rock-cut bench at the fiord head (4) was also found to be 70 m. Hodgson adopted

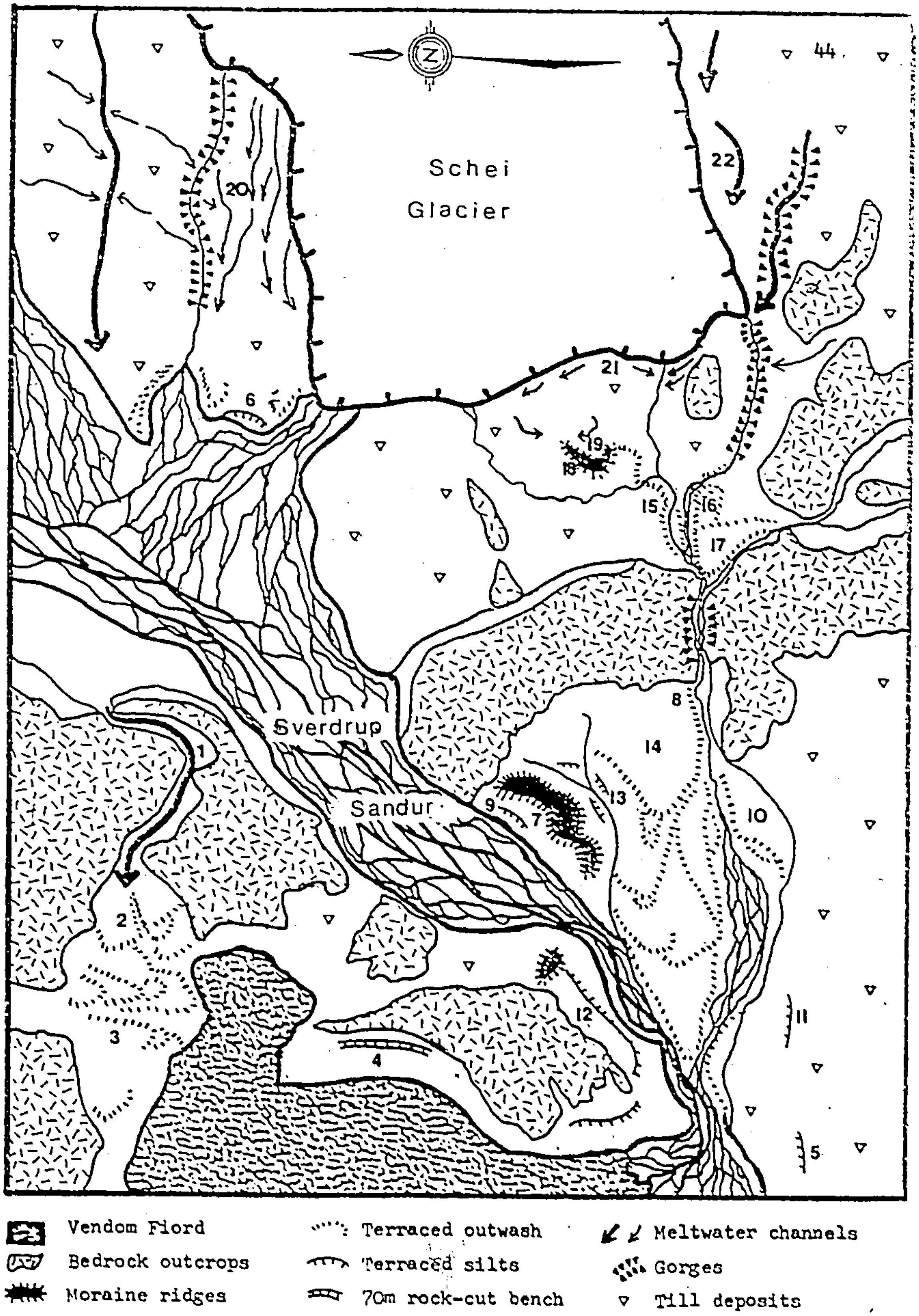


Figure 3.2: Major landforms and deposits at the head of Vendom Fiord.

this height as the local marine limit. Marine silts east of the present "Sverdrup" delta (5) did not exceed 65 m, and a height of 54 m was obtained for a similar deposit near the N.W. margin of the "Schei Glacier" (6).

As evidence for a readvance, Hodgson's findings are inconclusive. The terrace and bench heights yield limited information on local uplift, but do not themselves constitute evidence of glacial readvance. The meltwater channel evidence is ambiguous, as all of the channels mapped by Hodgson could have formed during the Innuitian retreat. Indeed, marginal meltwater channels are normally regarded as retreat-phase features (Mannerfelt, 1949; Sissons, 1960). The only positive evidence supplied by Hodgson does not concern the Vendom area :

> "Ice contact deltas graded to sea levels between 60 m and 80 m higher than present with well-defined lateral and terminal moraine ridges were noted elsewhere in central Ellesmere Island, at distances up to 20 km from the present ice caps."
>
> (p 134)

Some of Hodgson's evidence actually mitigates against a readvance. The undisturbed marine silts at (6) are inside the proposed readvance margin, which Hodgson suggested stood at the mouth of the "Sverdrup River". The surface layers of these silts contained paired bivalves of marine pelecypod shells, which would be unlikely to survive glacial onslaught. It is possible, however, that these lower (54 m) silts were deposited after retreat from the readvance maximum.

A re-examination of the glacial geomorphology of the Vendom Fiord area provided more substantive evidence for the occurrence of a lateglacial readvance. An arcuate ridge rising to 64 m on either bank of the "Sverdrup Sandur" is interpreted as an end-moraine (7). This feature is composed of rounded and sub-rounded gravels of heterogeneous composition, with a large component of allochthonous Precambrian rocks. The lack of fine material may indicate that the moraine consists of re-worked glaciofluvial sediments or, more likely, is attributable to the washing out of fines by a sea more than 64 m higher than present. A veneer of predominantly Archaen gravels which overlies nearby marine silts appears to be due to local redistribution of moraine material by wave action. Shallow pits dug at the margin of the moraine showed that the neighbouring silts were deposited against its flanks. This indicates that the moraine predates the silts.

Further evidence of a local glacial limit occurs at the distal end of the Schei Gorge" (8) where the even surface of the highest terrace (figure 3.3) ends abruptly in an area of stagnant ice deposits. Alternative interpretations of this hummocky area in terms of frost heave or ground ice action were rejected, as the adjacent area of the high terrace, though composed of similar material, is entirely hummock-free. The very district limit of the hummocky area is therefore interpreted as representing a glacial limit. Stratigraphic support for a lateglacial readvance was found in the exposed scarp of a perched delta at (10) (figure 3.4). The downstream truncation of near-horizontal beds of fine material by alternating foresets of gravel and

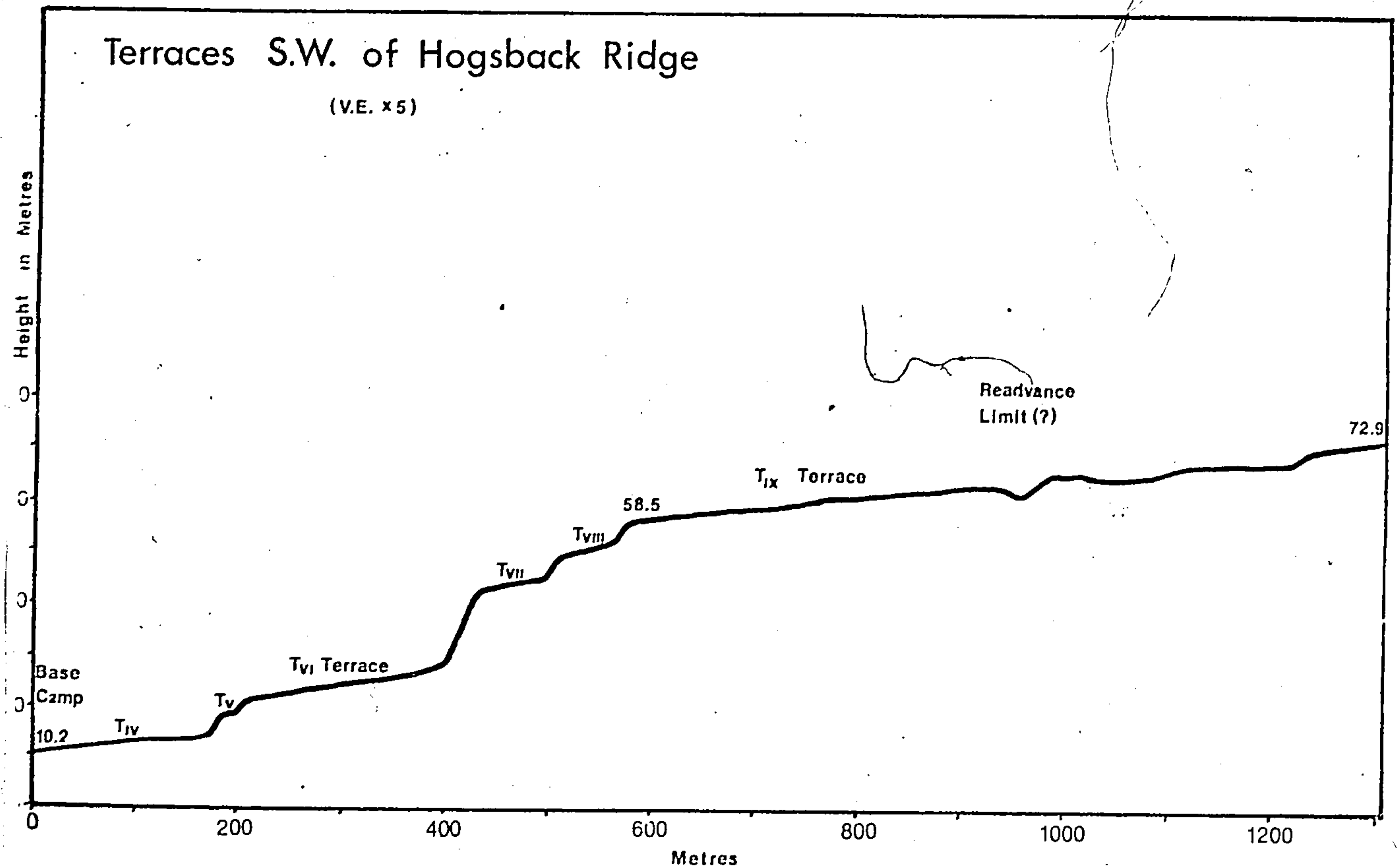


Figure 3.3: Levelled profile of sandur fragments T_{1V} to T_{1X} at the head of Vendom Fiord.

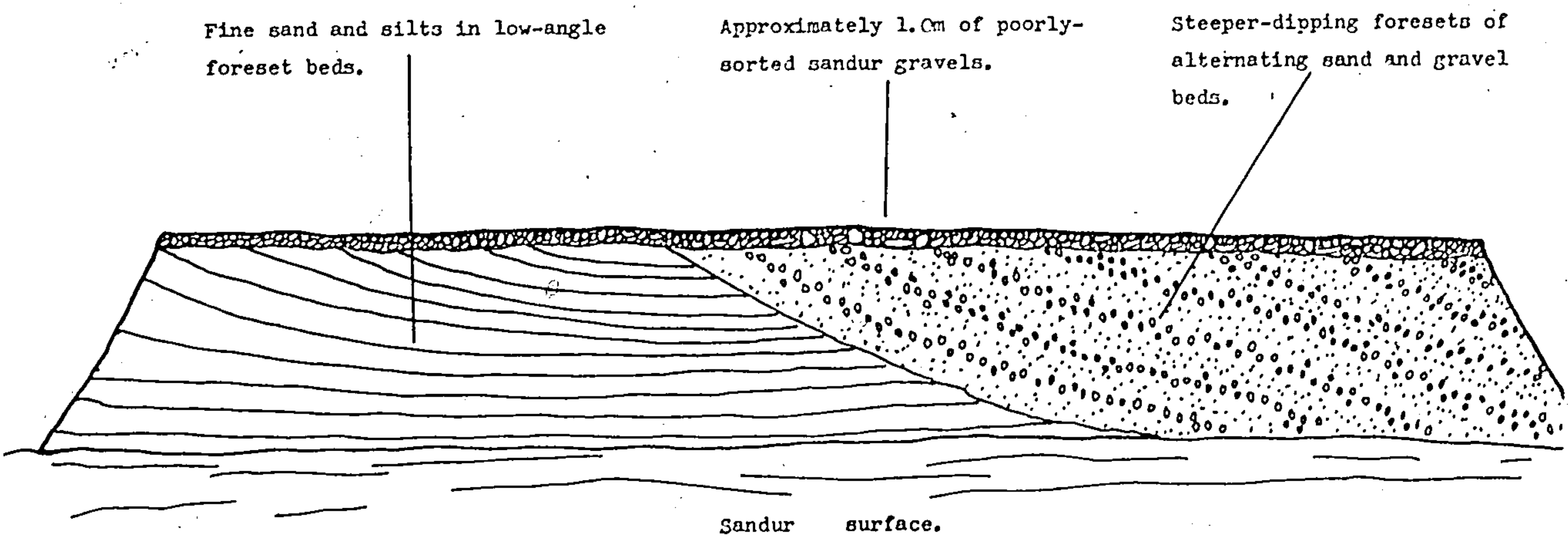


Figure 3.4: Schematic cross-section through the perched delta below the "Schei Gorge" at location 10. Height c.30m; Length c.450m.

sand indicates that quiet water deposition was succeeded by a period of sedimentation under more turbulent conditions. The higher parts of the foresets have also been truncated, and the entire exposure is topped by a metre of sandur gravels, with distinct braids visible on the surface and a small kettled area upstream. This suggests that the beds underlying the sandur gravels were deposited in a sea higher than the present level of the feature indicates. The distal end of the fragment is 58.9 m in elevation.

A fourth item of evidence favouring the existence of a readvance concerns the elevations of undisturbed marine silt deposits. Such deposits were examined at locations (5), (6), (9), (11), (12) and (13) and at a site 2 km west of (5). Paired bivalves of Hiatella arctica and Mya truncata were found on the surface of the silts at all these sites, and examples of Yoldia sapotilla were found at sites (6) and (13). The presence of paired bivalves and the lack of disturbance of rhythmites in the silt indicate that silt deposition post-dated glacial retreat. The silts at sites (5), (11) and (13), and the site west of (5) all supported a veneer of gravel, but this was apparently washed from nearby till deposits. The silts at (12) supported gravels washed from an adjacent talus slope and consisting almost entirely of local rock, carbonates of the middle Devonian Blue Mountain Formation. The lack of erratics in this last case confirms that the gravels are not of glacial origin. Theodolite survey indicated that all of the silt deposits outside the "Sverdrup Moraine" (7) achieve an elevation of over 60 m, with the exception of the degraded deposits by (13). Of the deposits inside the moraine limit, Hodgson reported those at (6) to be at 54 m, and those at (9) are well below that height. From the elevations of features associated with the postulated readvance, such as the spillway delta at 66 m, the wave-washed summit of the end-moraine (64 m) and the lower margin of the kettled sandur (65 m), the readvance can be related to a sea level of 65-70 m. However, the highest outwash surface (T_{1x}, figure 3.3), a product of the ensuing retreat phase, descends to 58.5 m. It follows that the silts outside the glacial limit must have been deposited before the withdrawal of ice from the readvance maximum. The lower silts inside the limit at (6) and (9), being undisturbed, must have been laid down after the withdrawal of ice.

Reinvestigation of the 70 m wave-cut bench at the head of the fiord (4) and the discovery of a possible degraded strandline at the same altitude above the marine silts at (12) confirmed Hodgson's 70 m marine limit. As the marine limit represents sea level at the time of Innuitian retreat, a 65-70 m readvance sea level implies that only a brief time interval separates the two events, as they occurred when uplift rate was at a maximum (Andrews, 1970). It is possible that this "Vendom Fiord stage" (Hodgson, 1973) represents a major stillstand in the retreat of Innuitian ice rather than a readvance.

The ^{14}C dates given above and a date of 8,200 ± 200 B.P. (GSC-146) for shells collected at 73 m near the south arm of Makinson Inlet suggest that ice withdrew from the fiord heads around or before 8,000 B.P. A date of 6370 ± 100 B.P. (GSC-118) obtained on the distal side of an end-moraine at Augusta Bay (Dyck and Fyles, 1964) is probably not representative of the period of Innuitian retreat, as the source material was taken from

37 m, well below the marine limit. <u>Hiatella arctica</u> shells collected by Hodgson (1973) at 52 m from the marine silts at (6), inside the "Vendom Fiord" limit, yielded a date of 6980 ± 80 B.P. (GSC-1858) and two further examples from a similar elevation yielded almost identical dates (D.A. Hodgson, personal communication). This date provides a minimum for the withdrawal of ice from the "Sverdrup" valley. Similar minimum dates for ice retreat have also been obtained for the north arm of Makinson Inlet (7310 ± 80; S.B. McCann, personal communication) and for the marine limit at Tanquary Fiord (6480 ± 200 (SI-468); 6320 ± 140 (GSC-373); Hattersley-Smith and Long, 1967). It is not known, however, if these sites lie inside a readvance limit, although this would seem not unlikely for the Makinson sample, which was taken near the present ice-front. A reasonable minimum date for the withdrawal of Innuitian ice from the area is provided by the 8200 ± 200 B.P. date for the south arm of Makinson Inlet, and Hodgson's 6980 ± B.P. date provides a minimum for ice retreat from the "Vendom Fiord" limit. The readvance would therefore appear to have occurred between 8,200 and 7,000 B.P. and may, as Hodgson suggests, correlate with the Cockburn stage on Baffin Island.

In summary, abundant evidence in the Vendom Fiord area indicates that the withdrawal of the Innuitian ice sheet was quickly succeeded by a readvance, the "Vendom Fiord stage", associated with a sea level of 65 - 70 m and occurring between 8,200 and 7,000 B.P. Hypothetical readvance limits are shown on figure 3.5. Hodgson (1973) has identified further evidence for a readvance elsewhere on west-central Ellesmere Island, and his suggestion that the "Vendom Fiord stage" is the Ellesmere correlative of the Cockburn stage on Baffin Island seems reasonable.

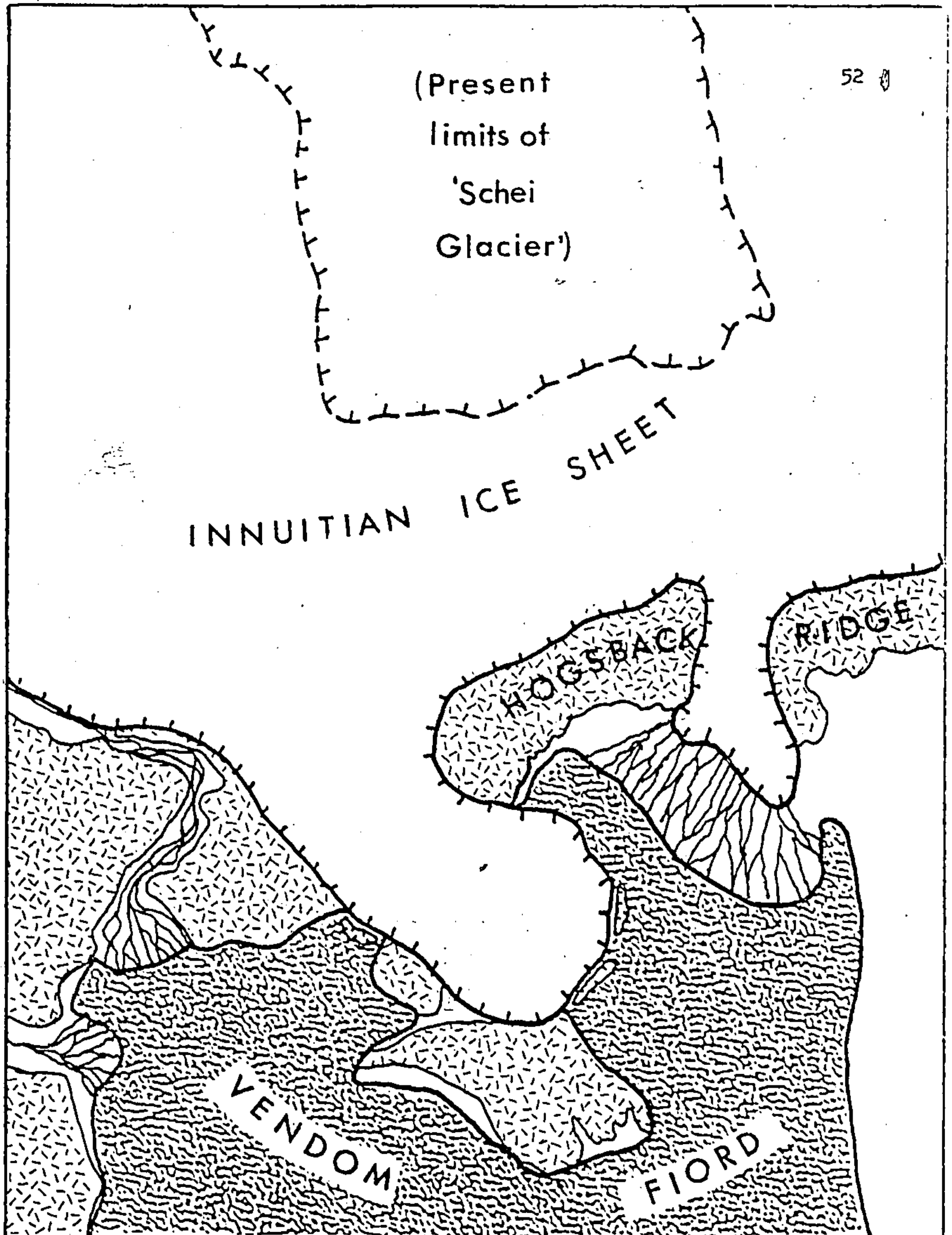


Figure 3.5: Ice margin position at the maximum of the "Vendom Fiord stage". Sea level at approximately 63m.

3.3. THE EVIDENCE FOR A NEOGLACIAL READVANCE

The highest of the Vendom terraces, T_{1X} on figure 3.3, was traced through the "Schei Gorge" and correlated with three relict sandur features (locations (15), (16) and (17) on figure 3.2; figure 3.6) in the basin west of the "Schei Glacier". Although the "Endrick" relict sandur (17) has remained largely unmodified since its deposition, little remains of the lateglacial "Lendal" (15) and "Upper Schei" (16) relict sandar. The "Lendal" sandur has clearly been eroded by a later advance of glacier ice from the east; at one point (18) an end moraine of reworked outwash gravels has been deposited on top of the old sandur surface and a small pond marks the site of a kettle-hole at (19). This suggests that a further minor readvance postdated the "Vendom Fiord stage", with the "Schei Glacier" advancing up to one kilometre west of its present position. A number of apparently recent meltwater channels, (20), (21), (22), probably mark the progressive retreat of the ice from the "Lendal" moraine (18) to its present position.

In order to relate this later readvance to sea level, and thereby gain some estimate of its age, terraces and relict sandur features associated with the readvance in the area east of the "Schei Gorge" were linked with those in the gorge itself, and these in turn were related to sandur fragments at the head of Vendom Fiord. Associating the re-advance stage with terrace features east of the "Schei Gorge" was facilitated by the presence of faint strandlines on the side of the "Lendal" moraine (18) and on the surrounding terrace scarp. These reflect the past presence of a small lake, dammed against the ice and emptying on to a younger relict sandur formed by the "Upper Schei River". Surveyed

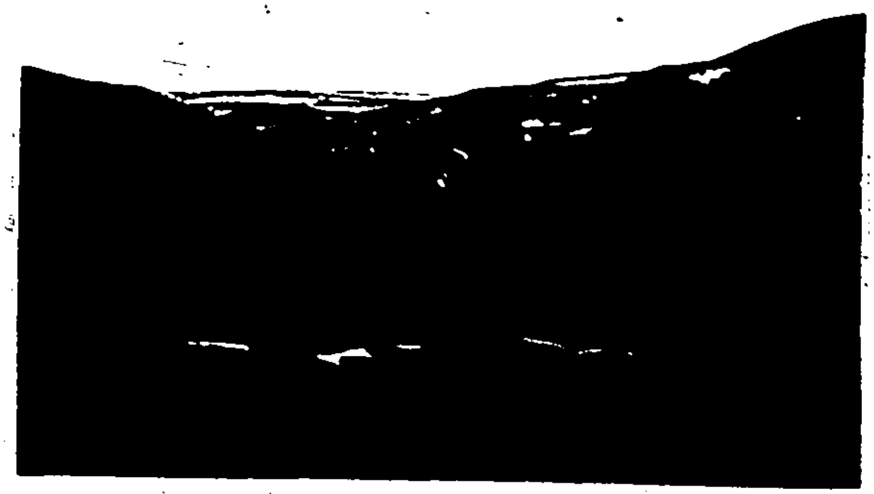


Figure 3.6 : The "Endrick" relict sandur, formed during the main Innuitian retreat phase. Active "Upper Schei" sandur in foreground.

heights on the strandlines corresponded approximately with the elevation of this sandur surface at the lake outlet (figure 3.7). It was assumed, therefore, that the lake emptied over the sandur surface and that the ice-dammed lake and this surface were contemporary features, both dating from the period immediately following the readvance maximum.

To enable this sandur surface to be related to those on the other side of the "Schei Gorge" the margins of sandur fragments above and below the gorge and the terrace fragments in the gorge were levelled (figure 3.8), as was the long profile of the lowest 8 km of the "Schei River" (figure 3.9). The fragment elevations were then plotted in the appropriate positions above the long profile (figures 3.10 and 3.11). In this way the second readvance was related to the lowest relict sandur surface, designated "T_{II}", at the head of Vendom Fiord. The accuracy of this correlation is confirmed by the equally fresh, unvegetated appearance of the T_{II} surface, the associated sandur surface east of the "Schei Gorge" and the interlinking terrace fragments. Unfortunately, little remains of the T_{II} surface below the Schei Gorge. The next highest terrace (T_{III}), however, extends to the confluence of the "Schei" and "Sverdrup" Rivers, where it is only 2.3 m above present sea level. The second readvance must have occurred when sea level was even lower, suggesting a possible neoglacial age. Such a conclusion remains tentative, however, as sea level change in the last two millenia has been very slow: a log collected at c.4 m at Makinson Inlet yielded a ^{14}C date of 2060 ± 50 B.P. (GSC-1836; D.A. Hodgson, personal communication; Blake, 1975).

In arriving at the above conclusion, the relationship between the evidence for a glacial readvance has been linked to sea level evidence to

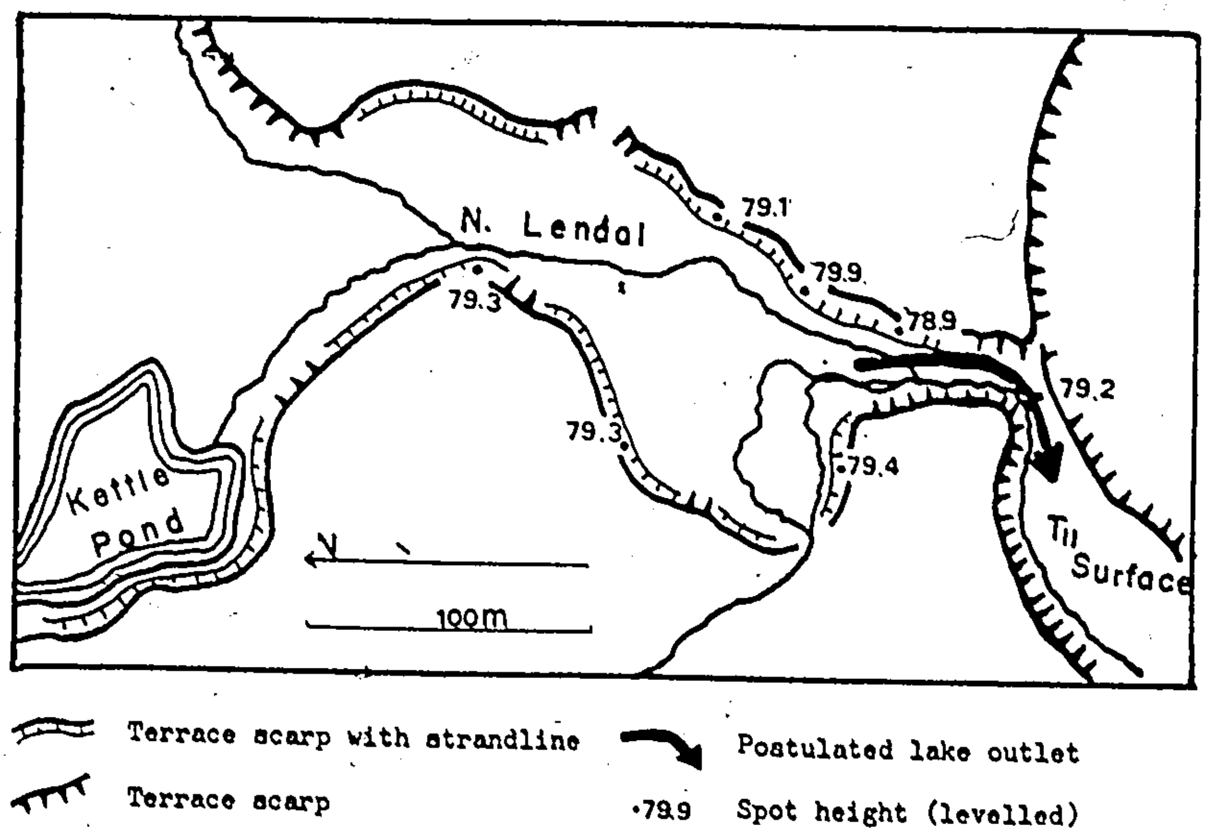


Figure 3.7: Heights levelled on a neoglacial (?) strandline and a postulated lake outlet on to the "Upper Schei" relict sandur (T_{II}) surface.

Schei River Terraces

SPOT HEIGHTS (m):

- RIVER
- ▲ T_{II}
- ▽ T_{III}
- ● T_{IV}

SYMBOLS:

- T_{III}
- T_{IV}
- OUTCROPS

0 100 m

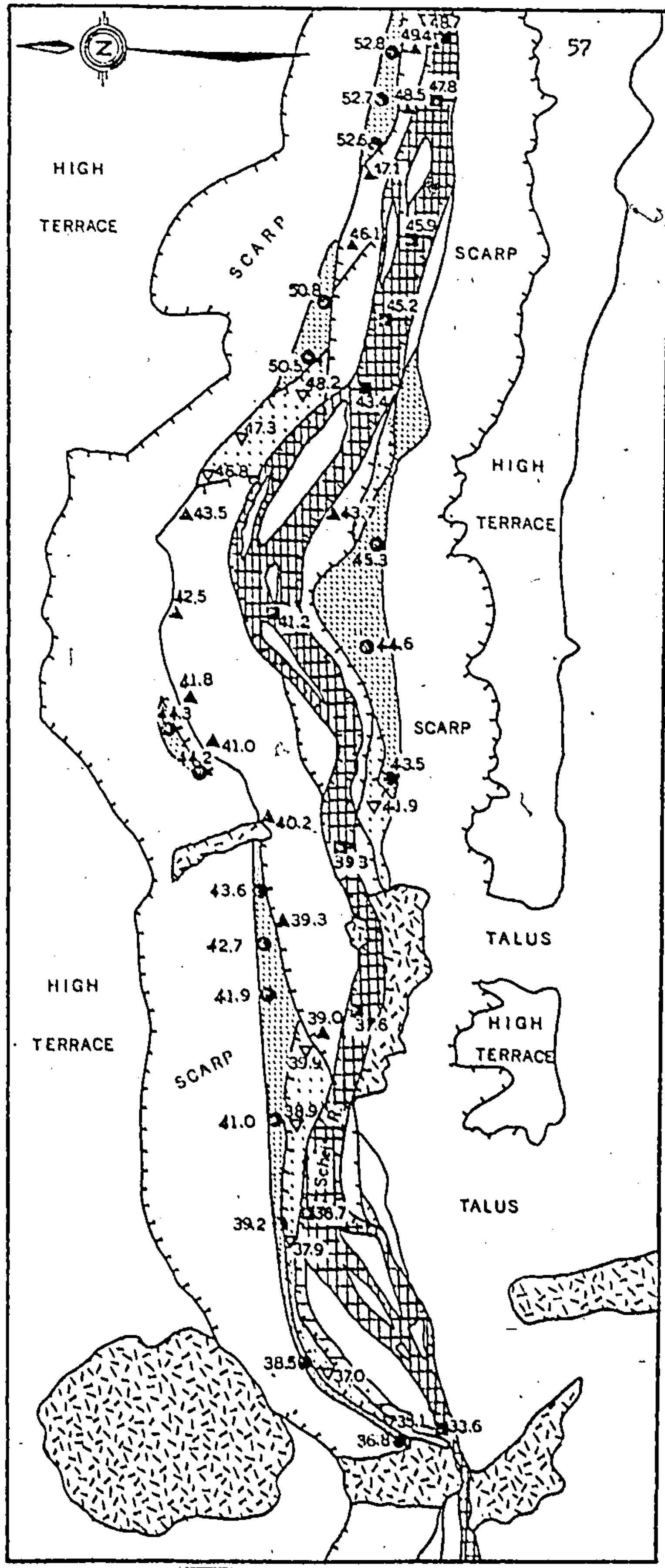


Figure 3.8:
Spot heights levelled on terrace fragments in the "Schei Gorge".

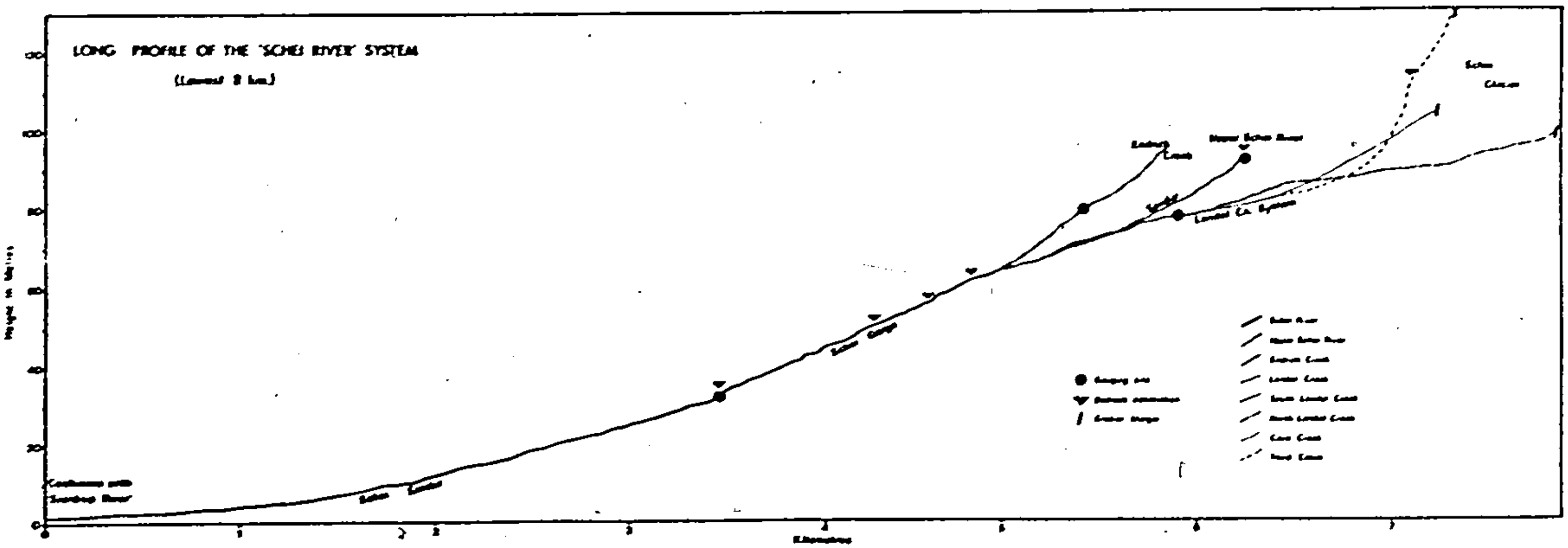


Figure 3.9: Long profile of the "Schei River".

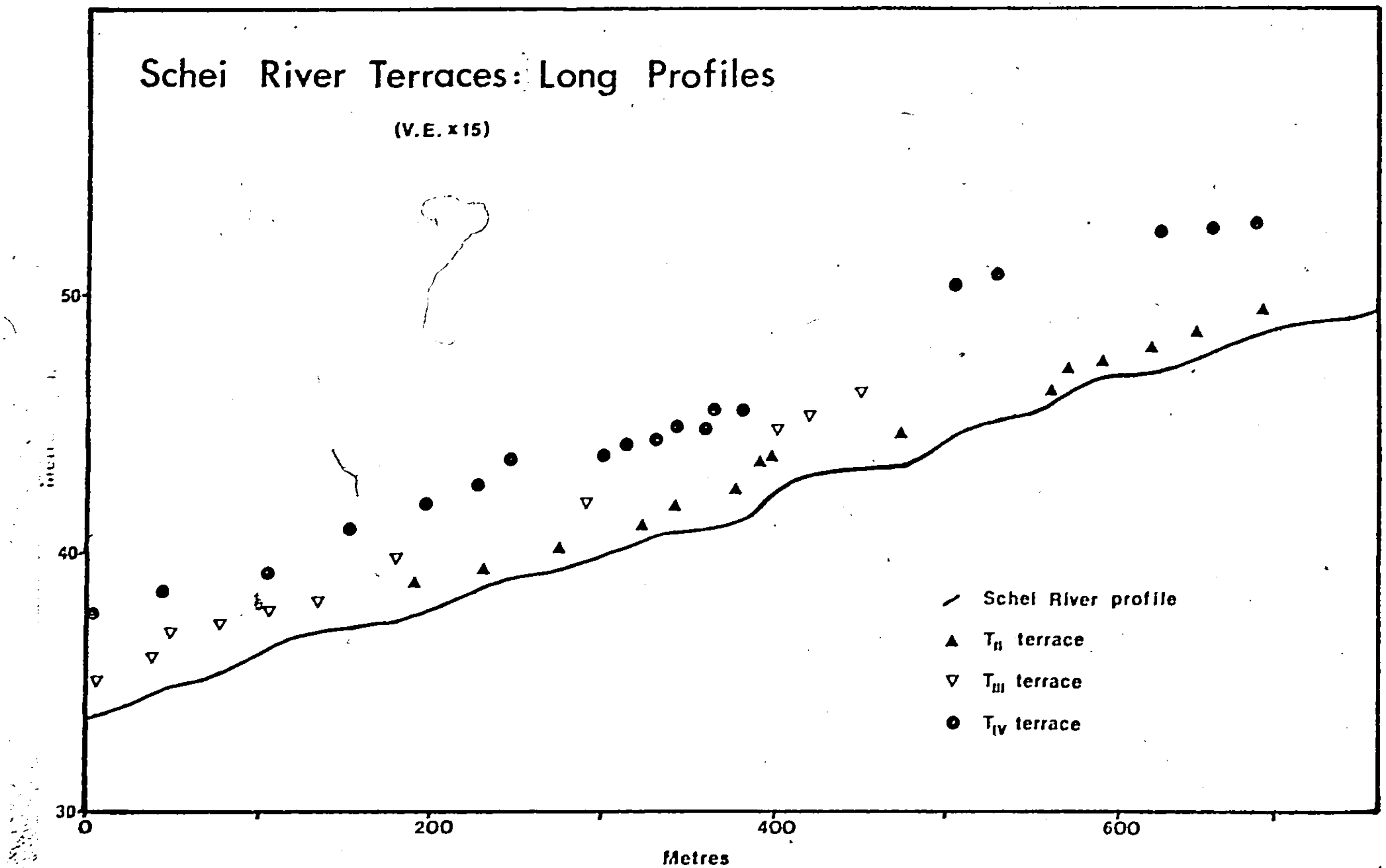


Figure 3.10: Terrace fragment spot heights plotted above present river profile
("Schei Gorge" section)

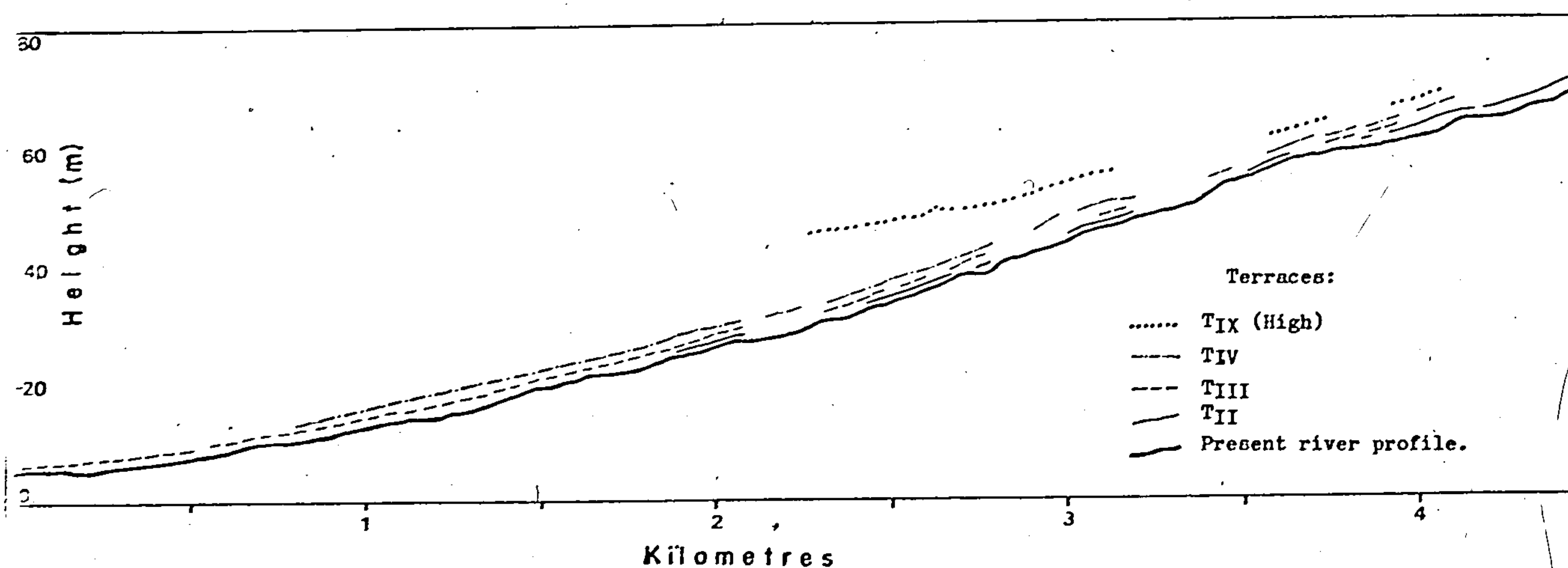


Figure 3.11: Terrace fragments plotted above the present "Schei River" profile from the "Upper Schei Sandur" to the lower "Schei Sandur"

to provide an age estimate for the former. In the eastern Canadian Arctic, where lateglacial landforms are generally well preserved, considerable potential exists for the employment of this technique to link glacial stages, sea levels and available dates in a more unified glacial history than is provided by the consideration of uplift and glacial stages in isolation.

3.4 THE POSTGLACIAL EVOLUTION OF THE "SCHEI RIVER" SYSTEM

3.4.1 The drainage network at the time of deglaciation

In the interpretation of lateglacial history given above, lateglacial, postglacial and neoglacial landforms were distinguished. In the present section, this information will be used in a reconstruction of part of the "Schei River" drainage network at the time of deglaciation, circa 7,000 B.P. The postglacial evolution of the drainage system can then be evaluated by comparing the modern drainage pattern with that which existed in the period following deglaciation.

It has been argued above that, sometime between 8,200 and 7,000 B.P., the retreating Innuitian ice-sheet readvanced to a limit bounded by the "Sverdrup" moraine (7) and the edge of the kettled sandur at (8) (figure 3.5). Sea level during the ensuing retreat phase dropped from c.65 m to c.58 m (figure 3.3). The postglacial "Schei River" was initially graded to this sea level. Below the "Schei Gorge" a wide sandur was deposited (location A on figure 3.12a), now represented by the T_{IX} terrace at the head of Vendom Fiord. The highest terraces in the gorge represent the fragments of a dalsandur or valley train (B), deposited

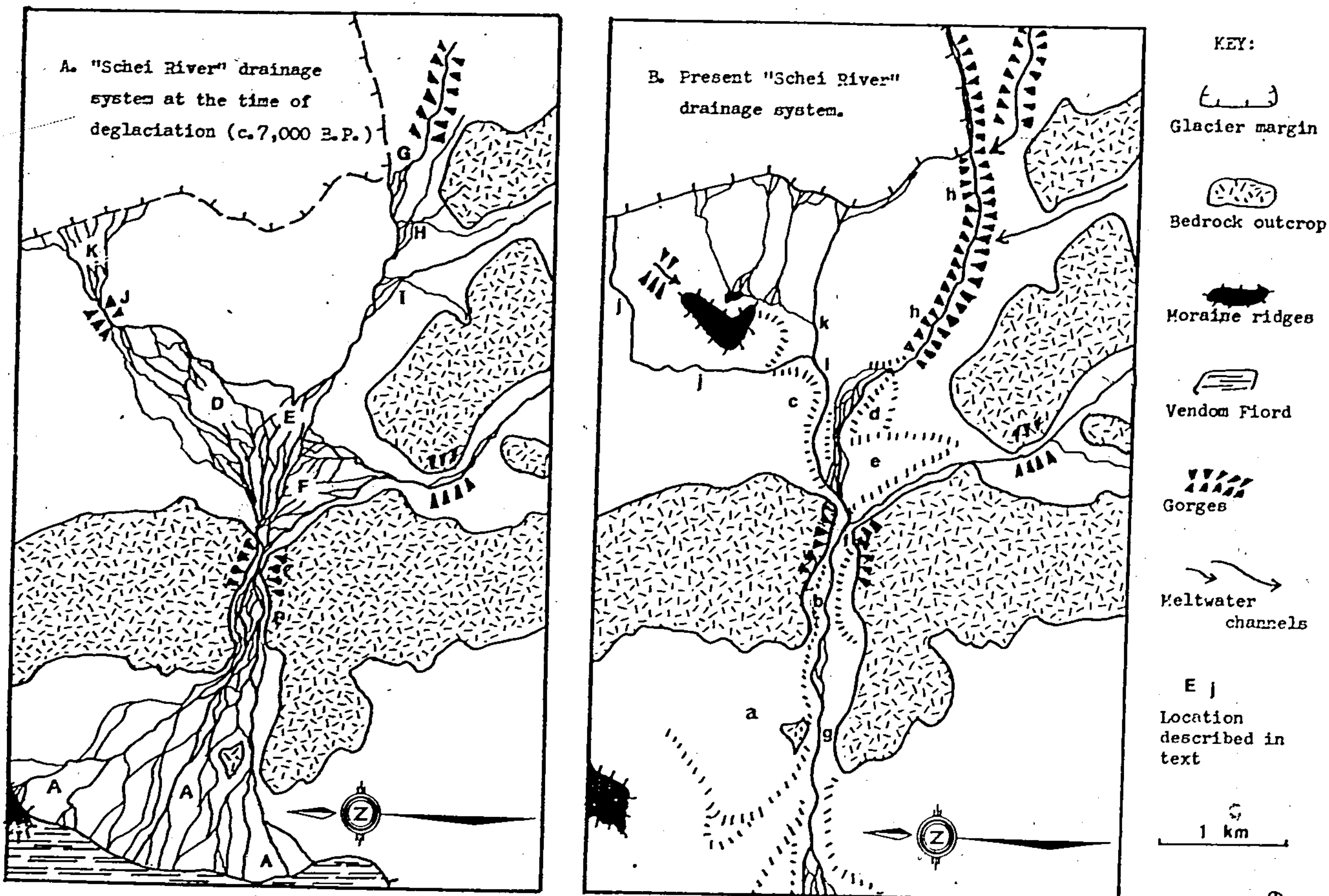


Figure 3.12: Postglacial changes in the "Schei River" drainage system.

by the retreating ice, but narrowing to a single channel (C) at a bedrock constriction. Beyond the head of the gorge existed three large sandar, each corresponding to one of the present tributaries of the "Schei River": "Lendal Creek" (D), the "Upper Schei River" (E) and "Endrick Creek" (F), all fed by glacier ice retreating to the higher parts of the basin.

The apex of the sandur at (E) was at a height of 102 m flush with the floor of the surrounding valley and indicating that no gorge then existed beyond this point. This is confirmed by the evidence of a number of tributary valleys (G), (H), (I) which were apparently the sites of lateglacial meltwater channels, and which were graded to the top of the present "Upper Schei Gorge", as evidenced by small alluvial fans perched above the gorge at the lower end of each tributary. The apex of the sandur at (D) was a now abandoned meltwater channel (J) cut deep in bedrock (figure 3.13) and draining a small upper sandur (K). The nature of the drainage system between sandur (D) and the present glacier margin is unknown, as this area was glacierized during the later neoglacial readvance.

3.4.2 Postglacial changes in the river system

From a comparison of the drainage network at the time of deglaciation with that of the present day (figure 3.12) it is evident that the major postglacial change has involved downcutting consequent on the lowering of sea level. At locations (a) and (b) on figure 3.12b the present river had downcut up to 35 m into the gravels of the postglacial sandar, intermediate stages being represented by the presence of up to seven terraces between the postglacial surface and that of the

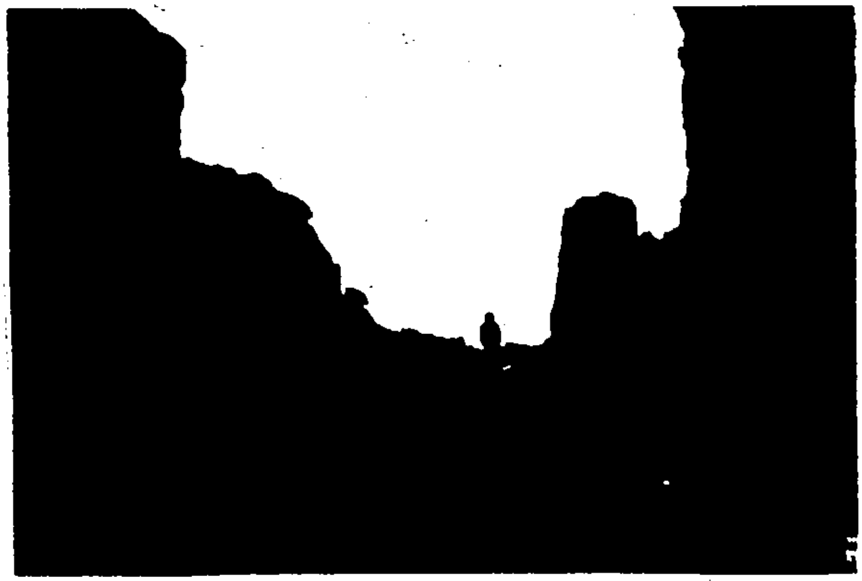


Figure 3.13 : Abandoned lateglacial meltwater channel cut through rock in the "Lendal Basin".

Figure 3.14 : The "Upper Schei" gorge, cut through competent sandstones and siltstones in the 7,000 years since deglaciation.

present day (figures 3.3 and 3.11). The three sandar east of the "Schei Gorge" (c), (d) and (e) have also been dissected, although downcutting does not exceed 9 m in this area. This suggests that the rock constriction at (f) has acted as a local base level. From the long profile (figure 3.9) it is evident that a slower rate of downcutting at this constriction has inhibited the formation of a smooth parabolic profile "graded" to present sea level. Nonetheless, postglacial incision of up to 9 m has taken place at this constriction and at (g), a further rock shelf encountered by the river as it downcut. More impressive is the postglacial downcutting accomplished (figure 3.14) at the "Upper Schei Gorge" (h). Above the apex of sandur (E) the river has downcut 12 m through the competent sandstones and siltstones of the Vendom Fiord Formation, from 102 m to 90 m. Further east, the distance from the top of the perched fans at (G), (H) and (I) to the base of the gorge exceeds 30 m. Some of this downcutting has been through weak Eureka Sound strata, but further downstream Vendom Fiord Formation rocks have been incised to a similar depth. The difference between the amount of downcutting in the "Upper Schei Gorge" and that at (f) and (g) probably reflects relative hardness of the Read Bay and Allen Bay rocks at the latter sites.

In the area drained by "Lendal Creek" (i) a substantial re-alignment of the lateglacial drainage pattern has occurred. This is largely attributable to the intervention of the neoglacial readvance. The neoglacial moraine separates the re-glaciated part of this basin form the unaffected part, indicating that the readvance caused an abandonment of the major meltwater channel at (J) and the northward realignment of drainage into "North Lendal Creek" (j). At its downstream end this new stream has cut 5 m into the lateglacial sandur surface (c) but further upstream it still flows on top of lateglacial outwash deposits, indicating that the 300(?) years since the neoglacial readvance have been insufficient for this stream to downcut to the level of adjacent tributary (k).

3.4.3 Conclusions

Three general conclusions emerge from the comparison of the lateglacial drainage pattern of c.7,000 ^{14}C years B.P. with that of the present day.

1. The modern drainage network reflects strongly the pattern of meltwater drainage as the ice-sheet retreated.
2. Substantial incision, resulting from a progressive postglacial drop in sea level, has occurred. This downcutting has resulted not only in the terracing of earlier outwash deposits but in the cutting of vertical gorges. The maximum incision into bedrock in the study area exceeded 30 m. Lithology appears to have significant control on the amount of downcutting.
3. The intervention of a glacial readvance disrupted the development of the present drainage network from that existing at the time of glaciation, with the formation of new channels and the abandonment of old.

The evidence for postglacial bedrock incision and gorge formation presented above vindicates the views of Cook (1967) and McCann and Cogley (1974) regarding the postglacial origin of steep-sided ravines on Devon and Cornwallis Islands. The evidence for bedrock incision in the

7,000 years since deglaciation also provides evidence of the effectiveness of fluvial erosion as a geomorphic agent at high latitudes, although it is apparent that the erosive energy of the "Schei River" and its tributaries since deglaciation owes much to the postglacial drop in sea level. As the rate of uplift slows (Andrews, 1970; Blake, 1975) so the importance of arctic rivers as agents of erosion, at least in the vertical direction, is likely to decrease.

CHAPTER 4

HYDROLOGY OF THE "SCHEI RIVER"

4.1 INTRODUCTION

The geomorphic efficiency of fluvial processes in the High Arctic is closely related to the nature of the arctic hydrologic regime. Early authors invoked the aridity of high latitudes and the brevity of the runoff season to diminute the geomorphic role of arctic streams, but more recent work has shown that these views are misguided. The storage of nine months' precipitation as snow or ice during the winter, and its release over a few days during the spring snowmelt freshet gives rise to high and geomorphically effective river discharges. Floods also result from summer rainfall, as runoff over glacier ice and barren permafrost terrain is rapid. Some glacierized basins may also experience floods resulting from the drainage of an ice-dammed lake.

Some attempt has been made to characterise the main elements of high arctic nival and glacial regimes (Church, 1974) but a comparison of the nival and glacial components of streamflow within a single glacierized catchment has not yet been published. Such a comparison comprises the principal aim of the present chapter. The glacionival regime of the "Schei River" for 1974 is described and compared briefly with that for 1973. The regimes of the three main tributaries of the "Schei River" are then described and analysed in terms of response to meteorologic inputs and the water-balance of the basin. These investigations allow several generalisations to be made about the hydrologic behaviour of the glacial and nival components of runoff. Some of the geomorphic implications of the differences between the two types become apparent in the ensuing (chapter 5) discussion of suspended and dissolved load.

4.2 BASIN CHARACTERISTICS

The catchment areas of the "Schei River" and its tributaries were determined by mapping watersheds from a mosaic of 1:60,000 vertical aerial photographs and measuring the enclosed area by planimeter. The 91.2 km^2 figure obtained in this way for the "Schei River" basin represents the catchment area above the main gauging point (figure 4.1); total basin area is slightly greater. An area of 32.4 km^2 or 35.5% of the gauged basin is covered by the "Schei Glacier", the main features of which have been described in chapter 2. The geologic and terrain characteristics of the basin have also been described in chapter 2.

To permit the identification and comparison of the nival and glacial components of runoff, the "Schei River" basin was divided into three sub-basins, drained by "Lendal Creek", "Endrick Creek" and the "Upper Schei River" (figure 4.1). Gauging stations were established near the mouths of the sub-basins (figure 4.2). The main characteristics of each sub-basin are listed in table 4.1.

In a high arctic basin where runoff is nourished principally by the melt of ice, snow and permafrost, altitude is an important factor in

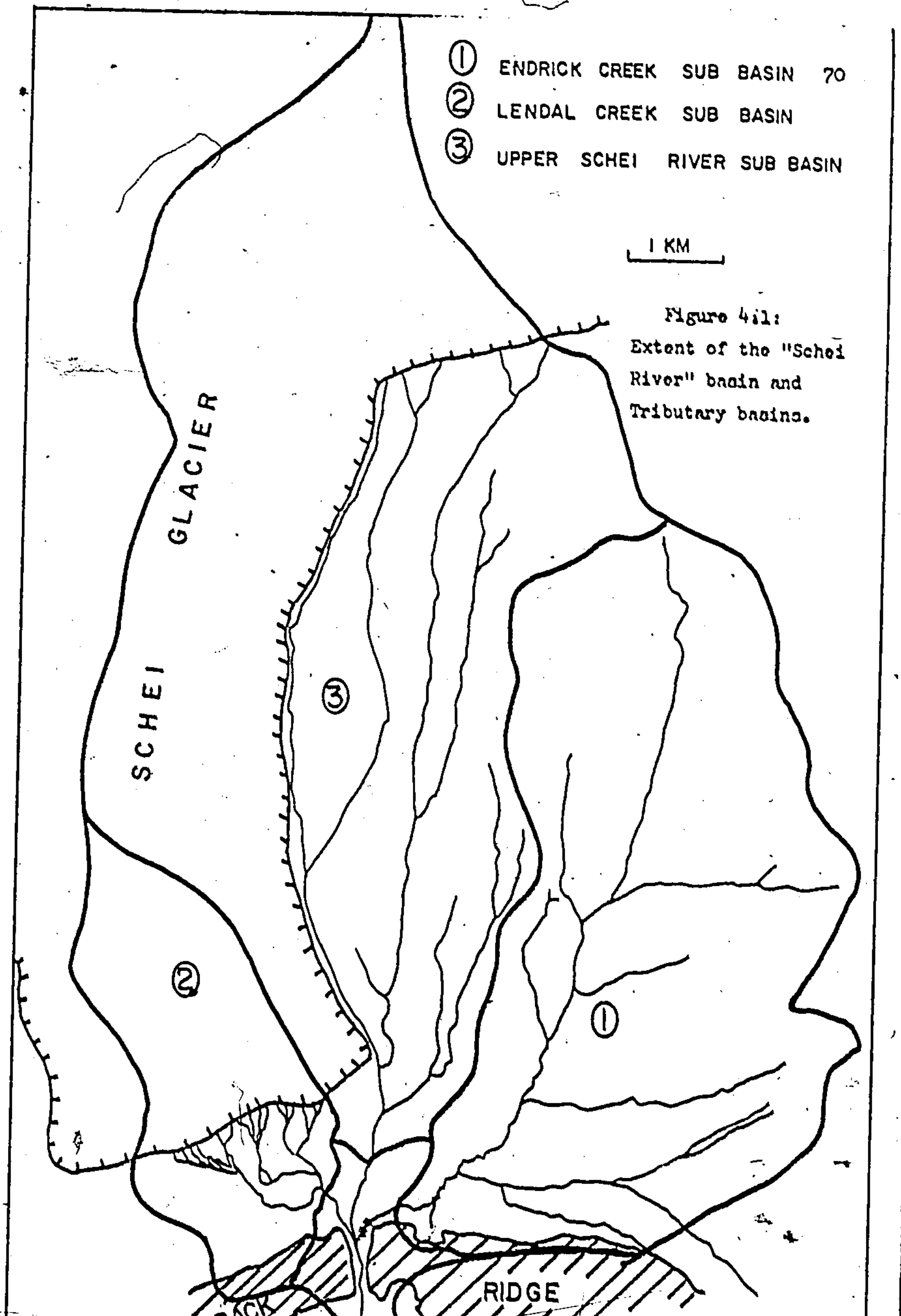


Figure 4.1: Extent of the "Schei River" basin and Tributary basins.

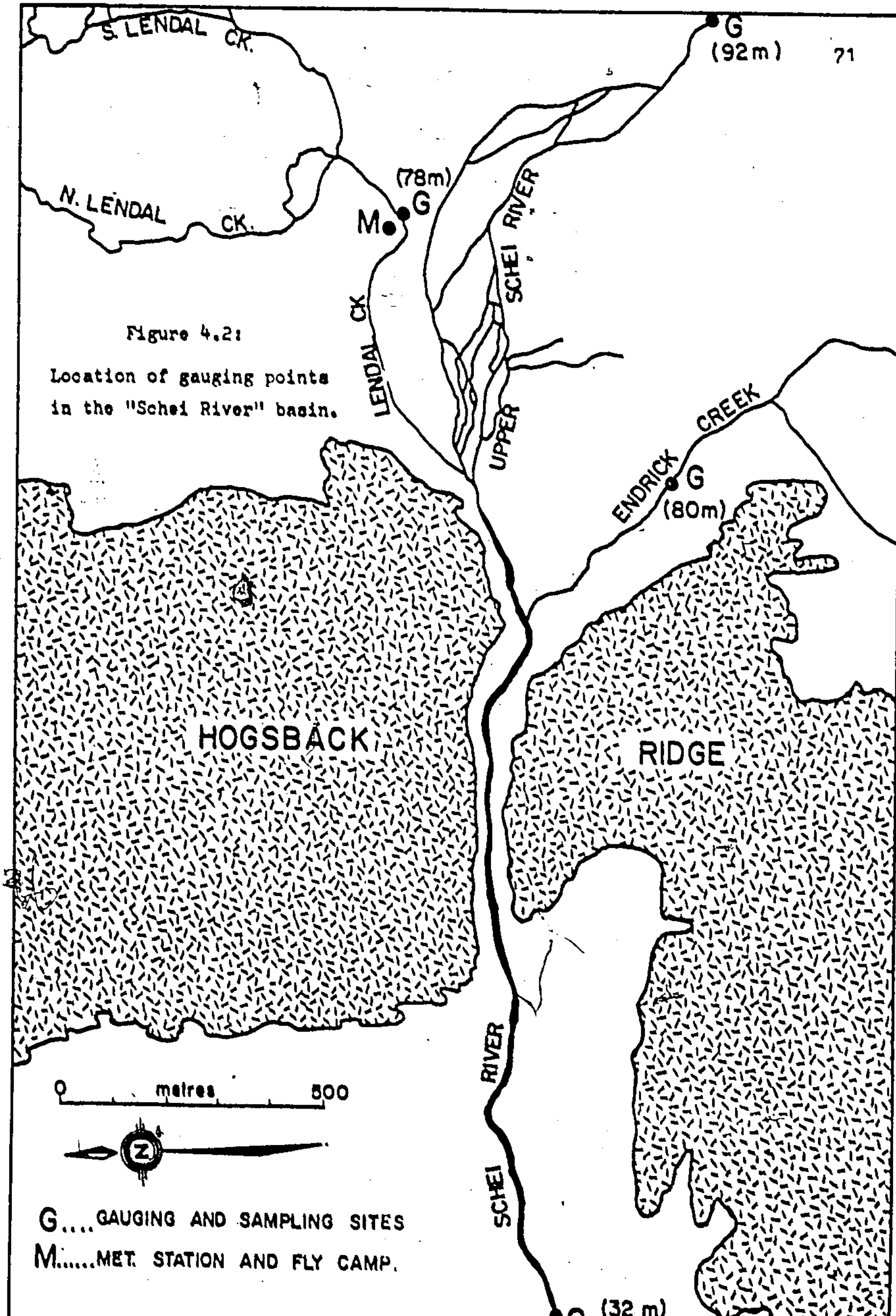


Figure 4.2: Location of gauging points in the "Schei River" basin.

TABLE 4.1

CHARACTERISTICS OF SCHEI RIVER BASIN AND SUB-BASINS[1]

Basin	Area (km^2)	% of Total Area	Glacierized Area (km^2)	% Glacierized Area	Elevation (m)		
					Min	Mean	Max
Schei River	91.2	100.0	32.4	35.5	32	360	800
Upper Schei River	49.3	54.1	26.5	54.0	92	505	800
Lendal Creek	9.4	10.3	5.9	62.8	78	275	390
Endrick Creek	29.9	32.8	0.0	0.0	80	415	650
Lower Schei River	2.6	2.8	0.0	0.0	32	100	200

<u>Notes:</u> 1. All basin measurements refer to the area above the gauging stations.

2. "Lower Schei River" is used here to refer to the area of the Schei River basin which lies downstream from the tributary gauging stations.

determining the time and duration of the spring freshet and the length of the runoff season for different parts of the basin. For example, the study of E.R.T.S. imagery for the Vendom Fiord area for the summer of 1974 showed progressive upglacier migration of the snowline as the runoff season progressed. To enable the effects of altitude to be assessed for each of the three tributary basins, hypsometric curves were constructed (figure 4.3). These were based on the contours of the 1:250,000 topographic map, and must be regarded as approximate. The most outstanding feature of these hypsometric curves is the presence of a "plateau" at an elevation 350 m to 450 m. This corresponds to the Tertiary (?) remnant surface discussed in chapter 2.

4.3 MEASUREMENT TECHNIQUES

During the 1974 runoff season, the discharge of the "Schei River" was measured at a gauging site situated at the apex of the lower of the river's two sandur reaches at an altitude of 32 m above sea level. A continuous stage record was measured using an Ott water-level recorder (figure 4.4) which was operational over the entire flow season except for the period July 12-17, when readings were obtained manually. The discharge of each of the main tributary streams was measured at gauging sites established on "Lendal Creek" at 78 m, on "Endrick Creek" at 80 m, and on the "Upper Schei River" at 92 m (figures 3.9 and 4.2). Stevens type F recorders were maintained on the "Upper Schei River" (first two weeks of season and on "Lendal Creek" (last six weeks of season; figure 4.5), and a Leupold-Stevens type E recorder was operated at the "Upper Schei River" gauging site from August 5 until measurements were

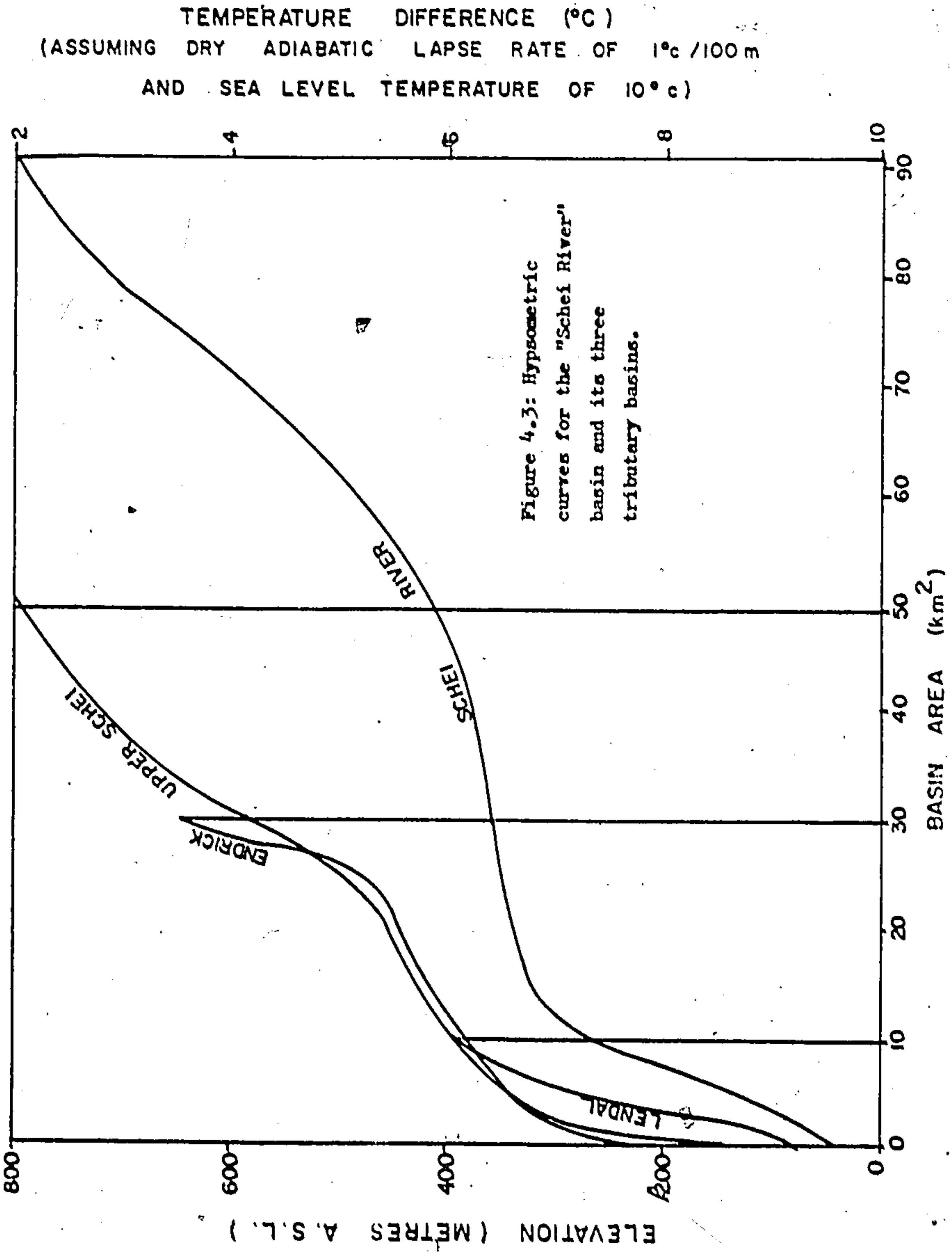


Figure 4.3: Hypsometric curves for the "Schei River" basin and its three tributary basins.

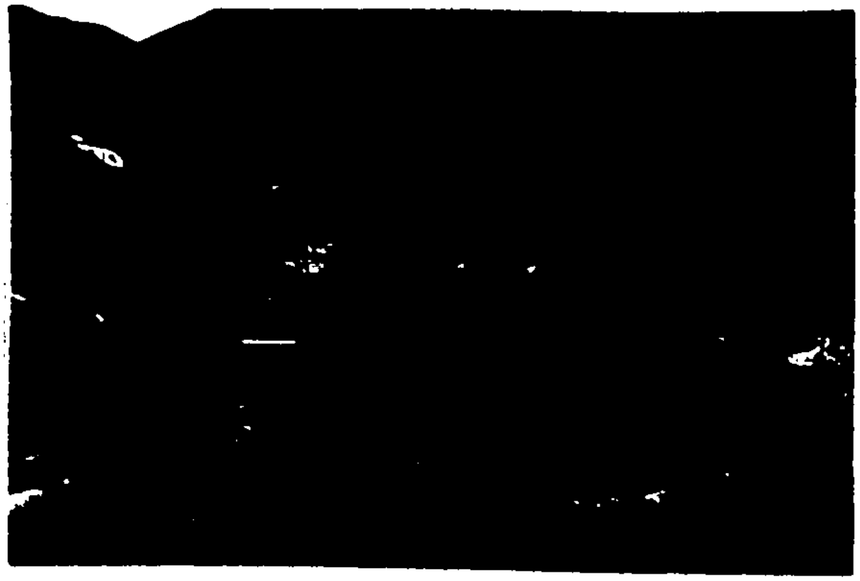


Figure 4.4 : Ott water-level recorder at the main "Schei River" gauging site.

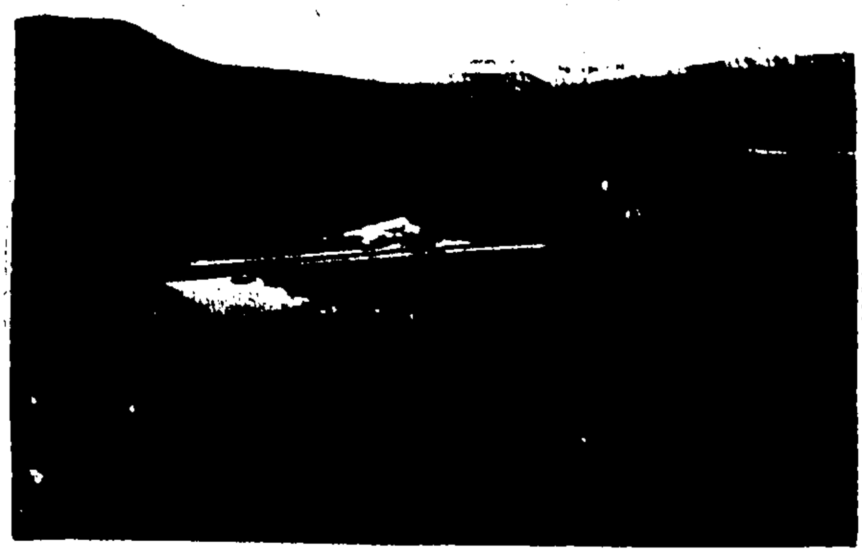


Figure 4.5 : Stevens' type F water-level recorder at the "Lendal Creek" gauging site.

terminated on August 18. These recorder measurements were supplemented by numerous manual readings with reference to fixed points. The "Endrick Creek" record was obtained entirely in this way, with an average of more than three measurements per 24 hour period.

The stage records for all gauging stations were translated into discharge hydrographs by establishing stage-discharge rating curves from which rating equations (table 4.2) were calculated. The rating curves are presented in appendix 1, along with a discussion of the discharge measurement techniques employed.

Meteorologic data for the 1974 runoff season were measured at the Vendom Fiord base camp (figure 2.1) and at the "Schei River" fly camp (figure 4.2). Precipitation measurements were made with tipping bucket raingauges, and Lambrecht thermohygrographs calibrated by daily thermometer readings were used to measure temperature. Net radiation was recorded on a Casella bimetallic actinograph, calibrated by independent solarimeter readings. The meteorologic record for the period June 21 to August 19 is shown on figure 4.6.

4.4 THE "SCHEI RIVER" HYDROGRAPH

The characteristic seasonal discharge pattern exhibited by streams with a high arctic nival regime is now well documented (Cogley, 1971, 1975 in preparation, McCann and Cogley, 1972, 1974; McCann, Howarth and Cogley, 1972; Walker, Lewis and Lake, 1973; Ambler, 1974; Church 1974). Flow normally begins in the second half of June or early July as ambient air temperatures exceed 0°C and snowmelt commences. Given clear skies.

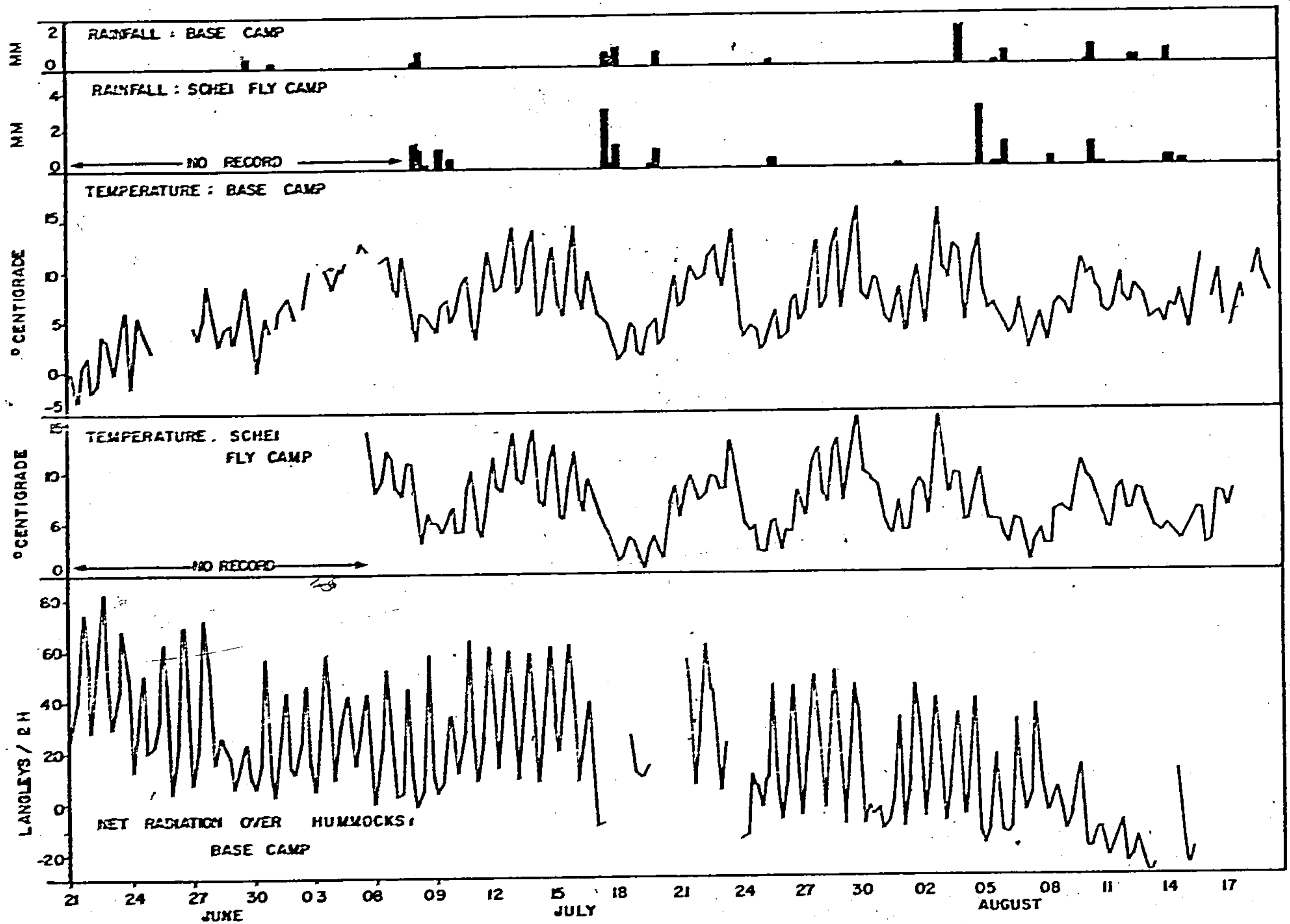


Figure 4.6: Meteorologic record, June 21 to August 19, 1974.

TABLE 4.2

RATING EQUATIONS FOR SCHEI RIVER AND TRIBUTARIES

River	Year	Rating Equation	N	r	r^2
Schei River	1973	$Q = 9.343y^{2.087}$	6	0.987	0.974
Schei River	1974	$Q = 14.306y^{1.919}$	9	0.952	0.906
Upper Schei River	1974	$Q = 5.940y^{1.957}$	9	0.985	0.960
Lendal Creek	1974	$Q = 37.730y^{1.902}$	14	0.972	0.945
Endrick Creek	1974	$Q = 4.818y^{1.963}$	10	0.956	0.914

<u>Notes:</u> Q is discharge in $m^3\ s^{-1}$.

y is stage in m.

N represents number of points in rating curve.

r is the correlation coefficient of the regression.

and sunny conditions, diurnal discharge maxima climb steadily to a peak, thereafter exhibiting a rather sharper decline as snowpack attenuation brings about a decrease in the supply of meltwater to the stream system. Unless interrupted by significant precipitation events, a pattern of low flow will prevail over the remainder of the runoff season, which lasts until temperatures drop below 0°C in late August or September, or until percolation rate exceeds rate of water supply and surface runoff ceases (Cogley, 1971). Throughout the runoff season pronounced diurnal discharge fluctuations are evident, the result of fluctuations in temperature and radiative inputs affecting snowmelt rate.

The runoff season of high arctic glacial streams also extends from June to September, but high discharges may be sustained almost to the end of the season as the supply of glacial meltwater is effectively unlimited. The period of peak discharge need not necessarily coincide with the snowmelt freshet, but appears to correspond with the period of highest mean daily temperatures. This may occur as late as early August (Anonymous, 1967; Church, 1974). Diurnal discharge fluctuations are also characteristic of glacial streams.

The "Schei River" hydrograph for the 1974 runoff season (figure 4.7) displays features typical of both nival and glacial regimes, justifying the designation of the "Schei River" regime as "glacionival". Flow began on June 22 (figure 4.8) and daily maxima displayed an irregular increase until the snowmelt freshet commenced on July 3. Peak discharge occurred on July 12, after which diurnal maxima declined rapidly in the manner typical of a nival stream. After several days of low flow (July 16-26), however, there was a renewed period of high discharge, with

Figure 4.7:

"Schei River" hydrograph, June 22 to August 18 1974.

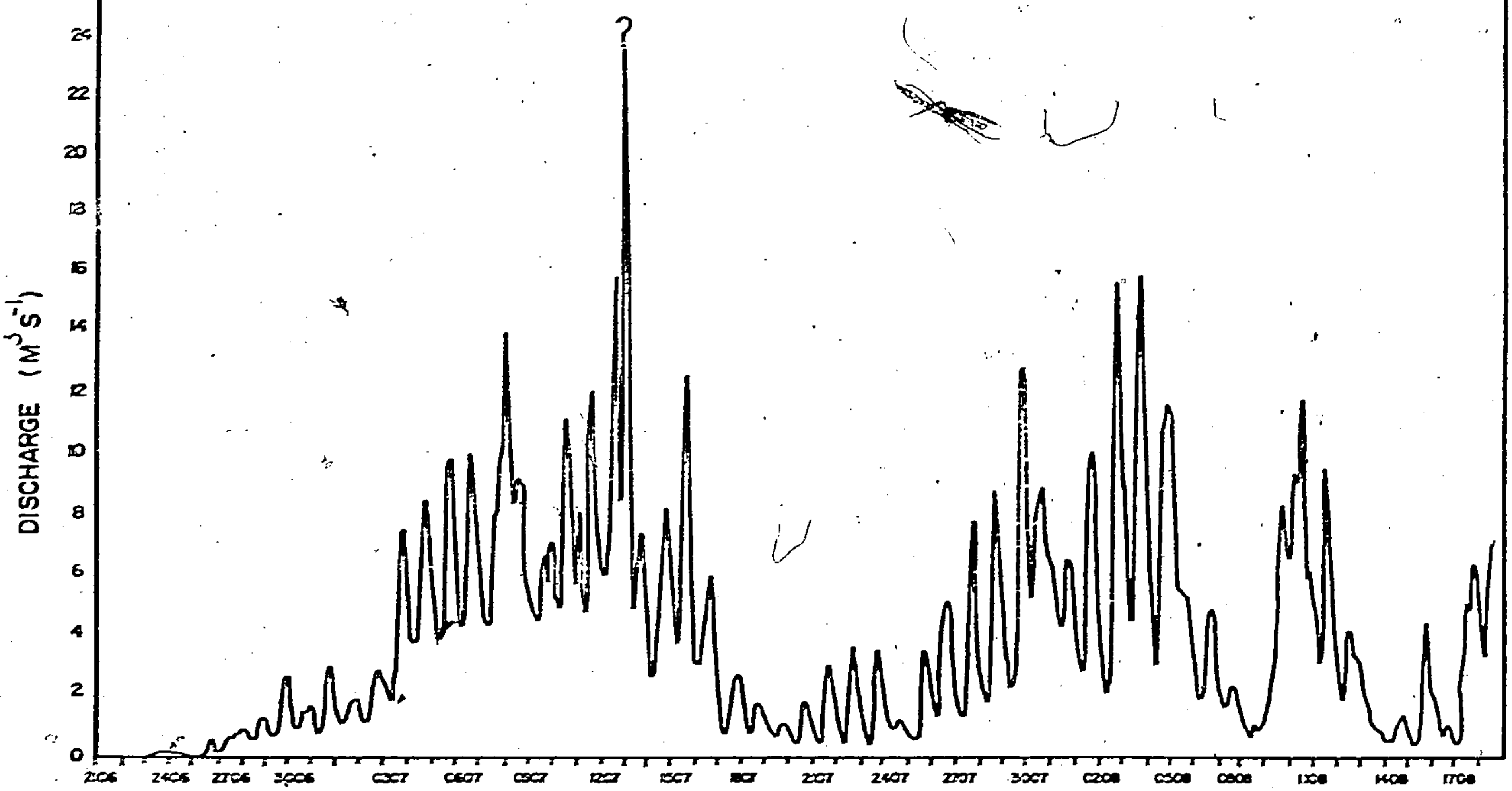


Figure 4.8 : "Schei River" in the "Schei Gorge" on the first day of flow, June 22, 1974.

extremely pronounced diurnal discharge fluctuations (figures 4.9 and 4.10). After August 5 discharge exhibited an irregular decline, with interruptions in the diurnal cycle caused by the passage of cyclonic storms. Time of freezeback is not known, but examination of E.R.T.S. satellite imagery showed that snowcover extended to sea level by mid-September, suggesting that freezeback had occurred by that date.

Although comparison of the "Schei River" discharge record with the meteorologic records for the runoff season (figure 4.11) reveals no broad correlation between radiation, temperature and discharge, some interesting details emerge. Both the period of the snowmelt freshet (July 3-July 16) and the second major period of high runoff (July 27-August 4) coincide with generally high temperatures. Two or three days of lower temperatures occur within these periods; these are reflected by drops in diurnal discharge maxima. The period of low flow in the middle of the season (July 16-26) appears to have been initiated by the return of near-freezing temperatures (July 16-19) and sustained by a further period of temperatures under 5°C (July 24-26).

On the seasonal scale, the hydrograph generally mirrors the temperature record more closely than the net radiation record. At the diurnal level, however, there was a stronger relationship between discharge and radiation than between discharge and temperature, confirming the findings of earlier investigators (Adams, 1966; Cogley, 1971; Cogley and McCann, 1971; McCann and Cogley, 1972). This topic is discussed more fully in section 4.6, along with that of response to rainfall.

The 1974 seasonal discharge pattern for the "Schei River" differs markedly from that measured for 1973 (McCann et al., 1974; figure 4.12).

Figure 4.9 : "Upper Schei River" at "Upper Schei Sandur", July 29, 0900 h.
$Q_u = 2.0\ m^3\ s^{-1}$

Figure 4.10 : "Upper Schei River" at "Upper Schei Sandur", July 29, 1800 h.
$Q_u = 12.0\ m3\ s\text{-}1$

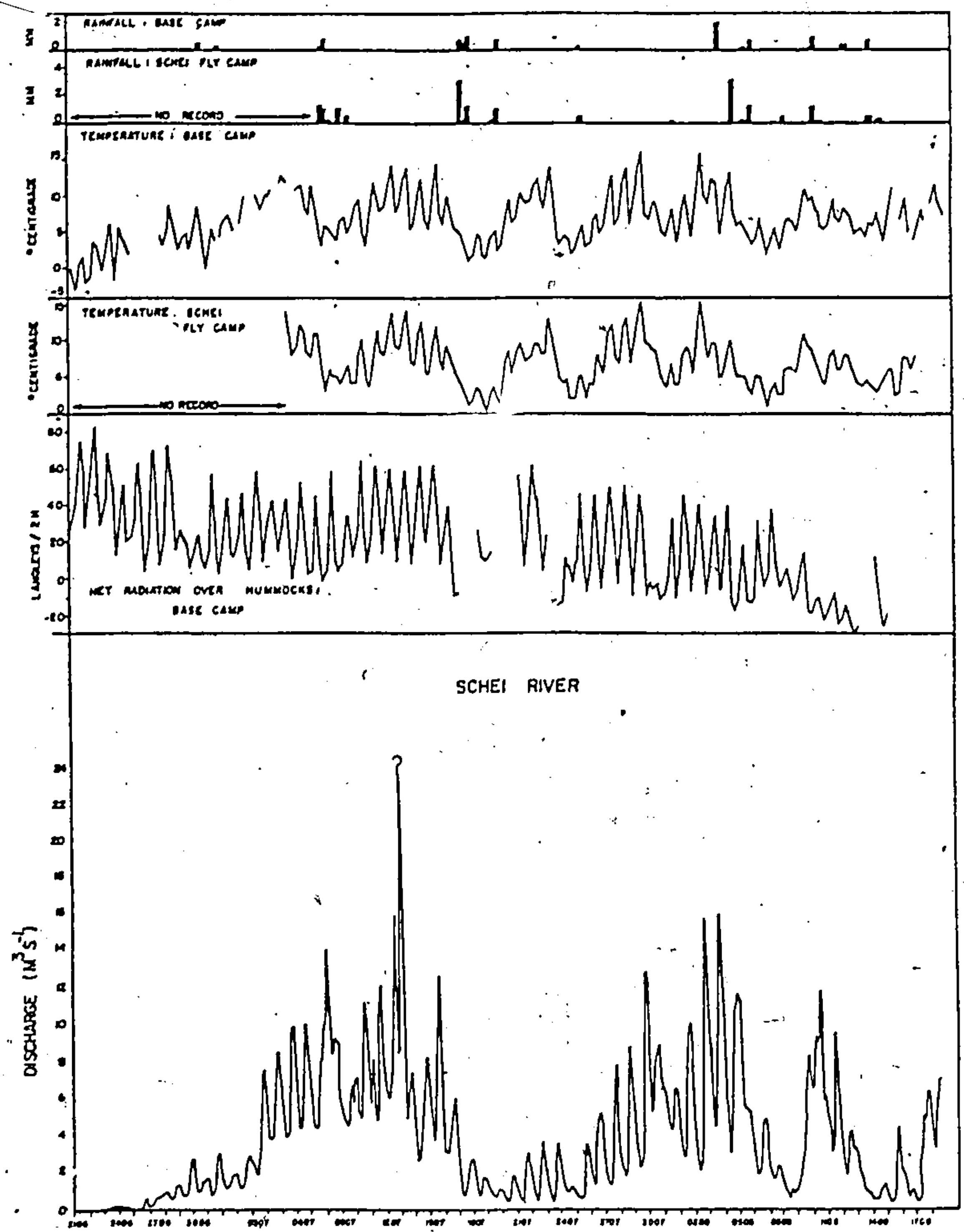


Figure 4.11: Meteorologic record and "Schei River" discharge, 1974.

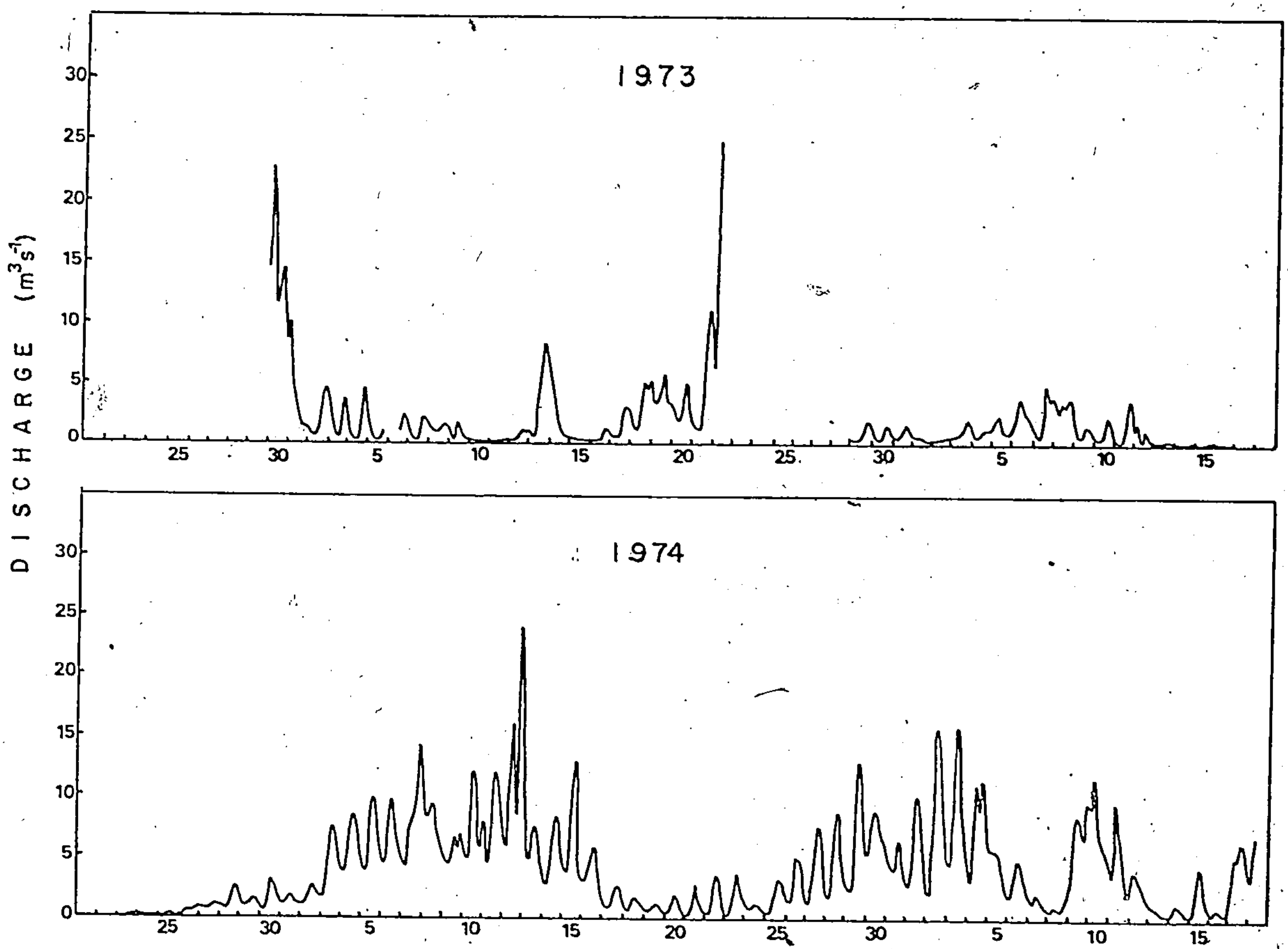


Figure 4.12: "Schei River" discharge, 1973 and 1974 (adapted from Bennett, 1975)

The main characteristics of the two runoff seasons are shown in table 4.3. The most striking difference between the hydrographs for 1973 and 1974 concerns the lack of glacial regime characteristics in the former. No meteorologic data accompanied the hydrograph in McCann et al. (1974), but it appears that this difference is attributable to the occurrence of persistantly cool, overcast weather during July and August, 1973 (J.G. Cogley, personal communication). As a result, relatively little glacier melt occurred, and the seasonal hydrograph assumed a typically nival pattern. The extremely high discharge of June 22, 1973 resulted from an exceptional rainstorm during which nearly 50 mm of rain fell in 24 hours (McCann et al., 1974; Cogley and McCann, 1975).

Using the flow-frequency curve of figure 4.13, the total discharge of the "Schei River" for the period June 22-August 18 1974 was calculated as $19.2 \times 10^6 m^3$, or 210 mm distributed over the entire basin, J.G. Cogley (personal communication) arrived at a total discharge figure of 16.9 x $10^6 m^3$ for the period June 23-August 20 1973 (185 mm distributed over the basin), but the derivation of the latter figure involved a number of large assumptions, and comparison of the hydrographs (figure 4.12) suggests that it is an overestimate.

In summary, comparison of discharge data for 1973 and 1974 indicates that weather patterns exert a considerable influence on the seasonal form of the "Schei River" hydrograph, but no simple relationship is apparent between net radiation, temperature and discharge. The hydrologic analysis of the three tributary basins reported in the following two sections further explores the influence of meteorologic variables on discharge, in particular their effect on runoff from the glacierized

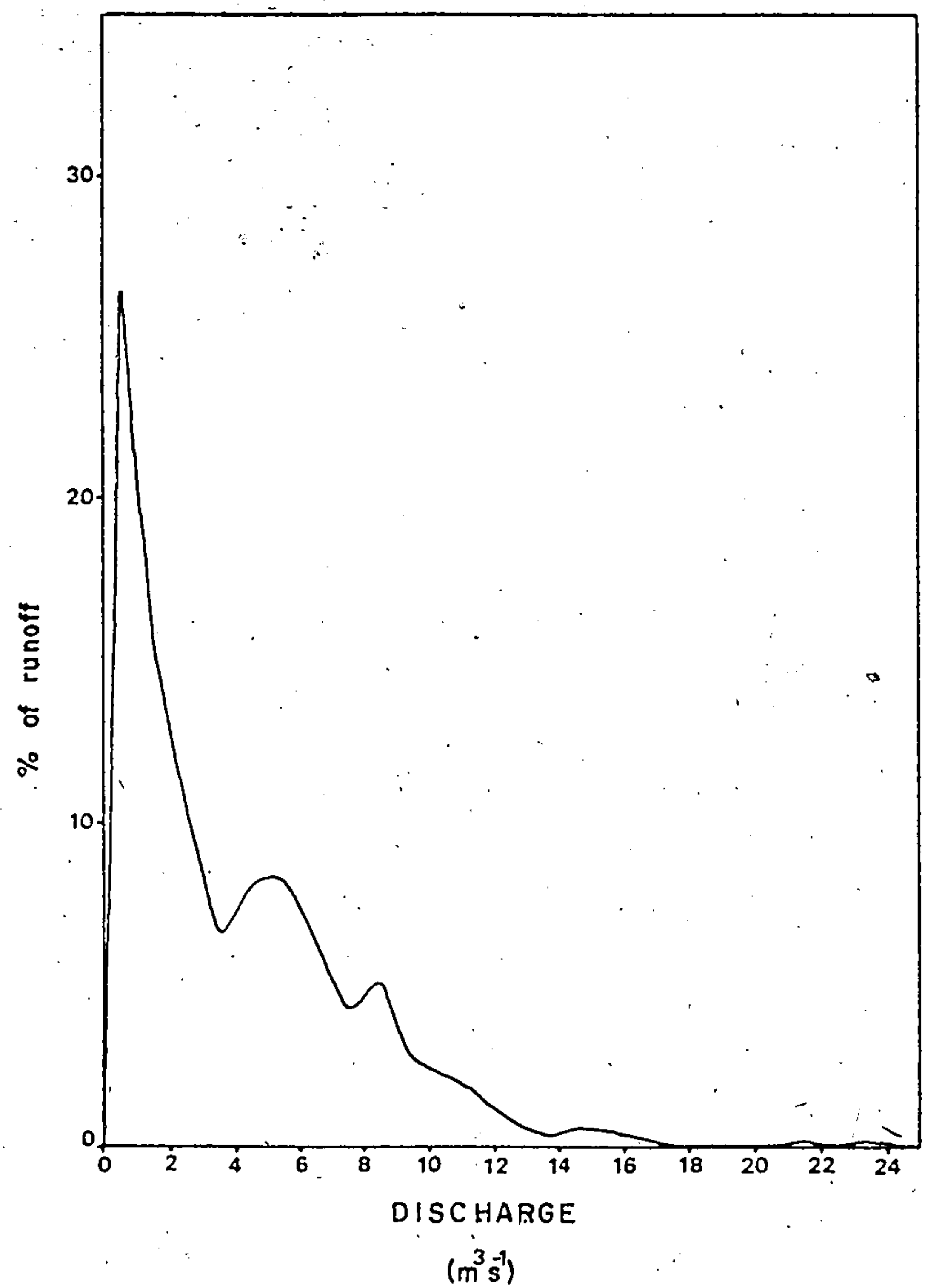


Figure 4.13: Flow frequency curve for the "Schei River", 1974.

TABLE 4.3:

"SCHEI RIVER" : MAIN CHARACTERISTICS OF RUNOFF IN 1973 and 1974

	1973	1974
First day of flow	before June 22	June 22
Last day of flow	unknown	unknown
Main periods of flow	before June 30; July 22 - 24	July 3 - 16 July 26 - August 5 August 9 - 11
Day of maximum discharge	July 22 (rainstorm)	July 12 (jokulhlaup)
Maximum recorded discharge ($m^3\ s^{-1}$)	? 50 (rainstorm)	? 24 (jokulhlaup)
Maximum recorded discharge ($m^3\ s^{-1}\ km^{-2}$)	0.549 (rainstorm)	0.260 (jokulhlaup)
Total discharge ($10^6 \times m^3$)	16.9[1] (June 23 - Aug. 20)	19.2 (June 22 - Aug. 18)
Total discharge ($10^4 \times m^3\ km^{-2}$)	17.434 (June 23 - Aug. 20)	21.051 (June 22 - Aug. 18)

1. J.G. Cogley, personal communication.

portions of the basin.

4.5 THE TRIBUTARY HYDROGRAPHS

The fluctuations of discharge recorded on the "Schei River" hydrograph represent the output from an extremely complex hydrometeorologic system. Comparison of the hydrograph with the meteorologic record permitted only very limited analysis of this system. In order to investigate the regime of the "Schei River" in greater detail, its basin was subdivided into three major units (figures 4.1 and 4.2; table 4.1). The hydrographs of the streams draining these three tributary basins are shown on figure 4.14, and the main characteristics of the 1974 flow regimes are summarised in table 4.4.

Flow was first observed on all three tributaries on June 22, the first day of flow of the "Schei River" itself. The two glacier-fed tributary streams, the "Upper Schei River" and "Lendal Creek", maintained high levels of discharge up to and beyond the end of observations on August 18. By August 17, however, the waters of "Endrick Creek" were no longer reaching the "Schei River" by the surface flow. Although runoff was observed on the higher reaches of "Endrick Creek" on that date, this water was percolating into the gravels that form the bed of the downstream end of the river. A similar phenomenon has been documented for a small nival stream on Devon Island by Cogley (1971). The virtual cessation of flow from the unglacierized "Endrick Basin" results from the depletion of meltwater sources (snow, ground ice and permafrost) before the onset of the freezeback period. The glacier-fed streams, however, continued to receive a substantial input from the melting of glacier ice

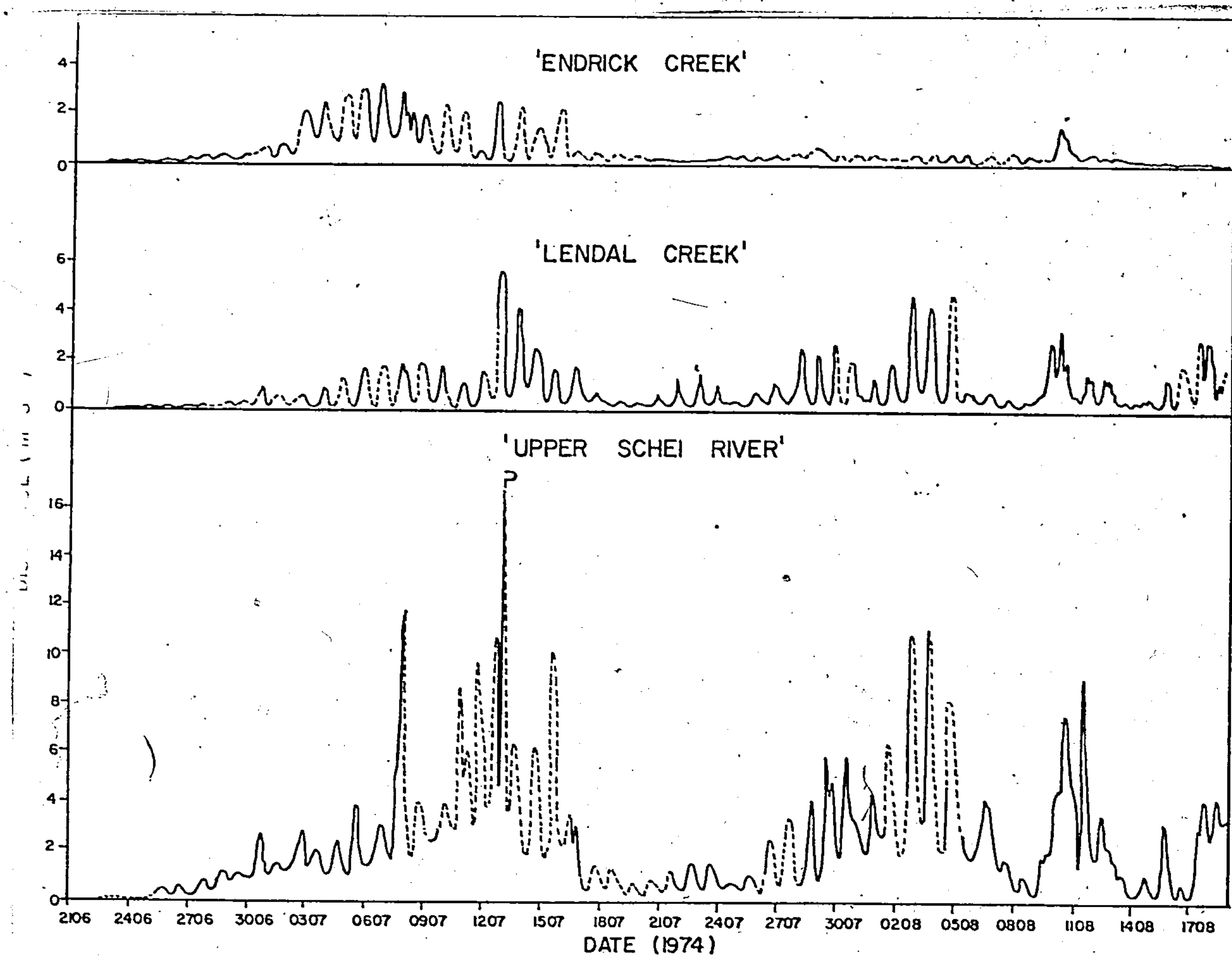


Figure 4.14: "Schei River" tributary hydrographs.

TABLE 4.4

SCHEI RIVER TRIBUTARIES:
MAIN CHARACTERISTICS OF 1974 RUNOFF SEASON

	Upper Schei River	Lendal Creek	Endrick Creek
First day of flow	22 June	22 June	22 June
Last day of flow	unknown	unknown	17 Aug.
Main periods of flow	30 June-17 July; 26 July-7 Aug; 9-13 Aug.	5-16 July; 28 July-4 Aug; 10-12 Aug.	2-17 July
Day of maximum discharge	13 July	12 July	6 July
Maximum discharge ($m^3\ s^{-1}$)	16.0+	5.6	3.2
Maximum discharge ($m^3\ s^{-1}\ km^{-2}$)	0.325	0.596	0.107
Total discharge, 22 June-18Aug. ($10^6 \times m^3$)	12.094	4.568	2.948
Total discharge, 22 June-18 Aug. ($10^4 \times m^3\ km^{-2}$)	24.552	48.600	9.860

until the return of sub-zero temperatures.

The highest recorded discharge for each tributary occurred during the snowmelt freshet, which extended from June 30 to July 16. "Endrick Creek" achieved a maximum discharge in excess of 3.2 m^3s^{-1} on July 6 (figure 4.15), and "Lendal Creek" reached a maximum of 5.6 m^3s^{-1} on July 12. The July 12-13 maximum of the "Upper Schei River" reflects an exceptional flood event of a type discussed below, as does the high discharge recorded on July 7-8. The highest recorded "normal" discharge of the "Upper Schei River" (10.5 m^3s^{-1}) did not occur until August 4, but it is likely that discharge on July 12 would have been higher, even had runoff not been affected by exceptional circumstances.

The seasonal hydrograph pattern of the "Schei River" is strongly reflected by the major tributary, the "Upper Schei River", and to a lesser extent by the glacier-nourished "Lendal Creek" (figure 4.16). All three hydrographs display a gradual buildup to the snowmelt maximum of July 12, a period of relatively low discharge between July 17 and July 25, then a steady increase in daily maxima up to a second peak on August 3-4. This was followed by four days of declining flow, then three days of high discharge, August 9-11. A further five days of low flow preceded the moderately high discharges recorded on the last two days of measurement, August 17 and 18.

Like that of the glacier-fed tributaries, the discharge of "Endrick Creek" displayed a gradual rise to a snowmelt maximum, in this case on July 6, and an abrupt drop in discharge after July. Unlike that of the other tributaries, however, "Endrick" discharge exhibited no response to the high temperatures of late July or early August, but remained generally

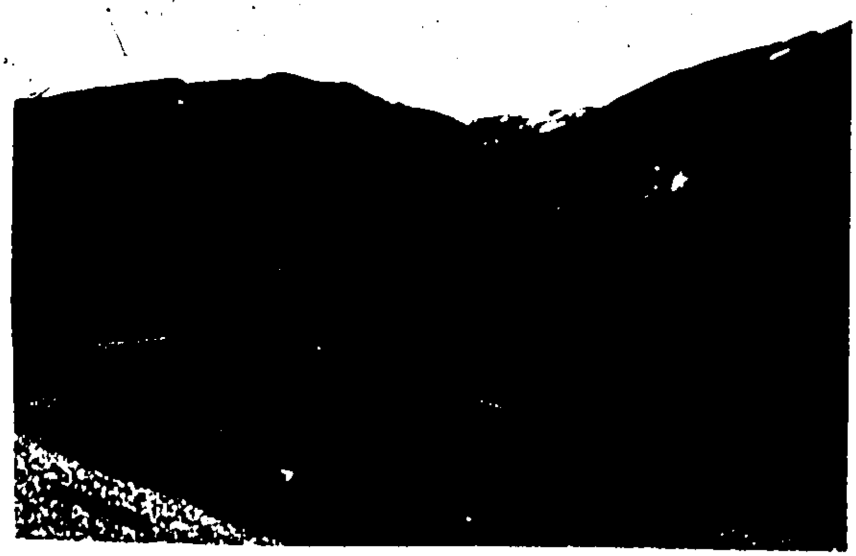


Figure 4.15 : "Endrick Creek" near the peak of the snowmelt flood period.
Q_e = c 3.0 m^3 s^{-1}

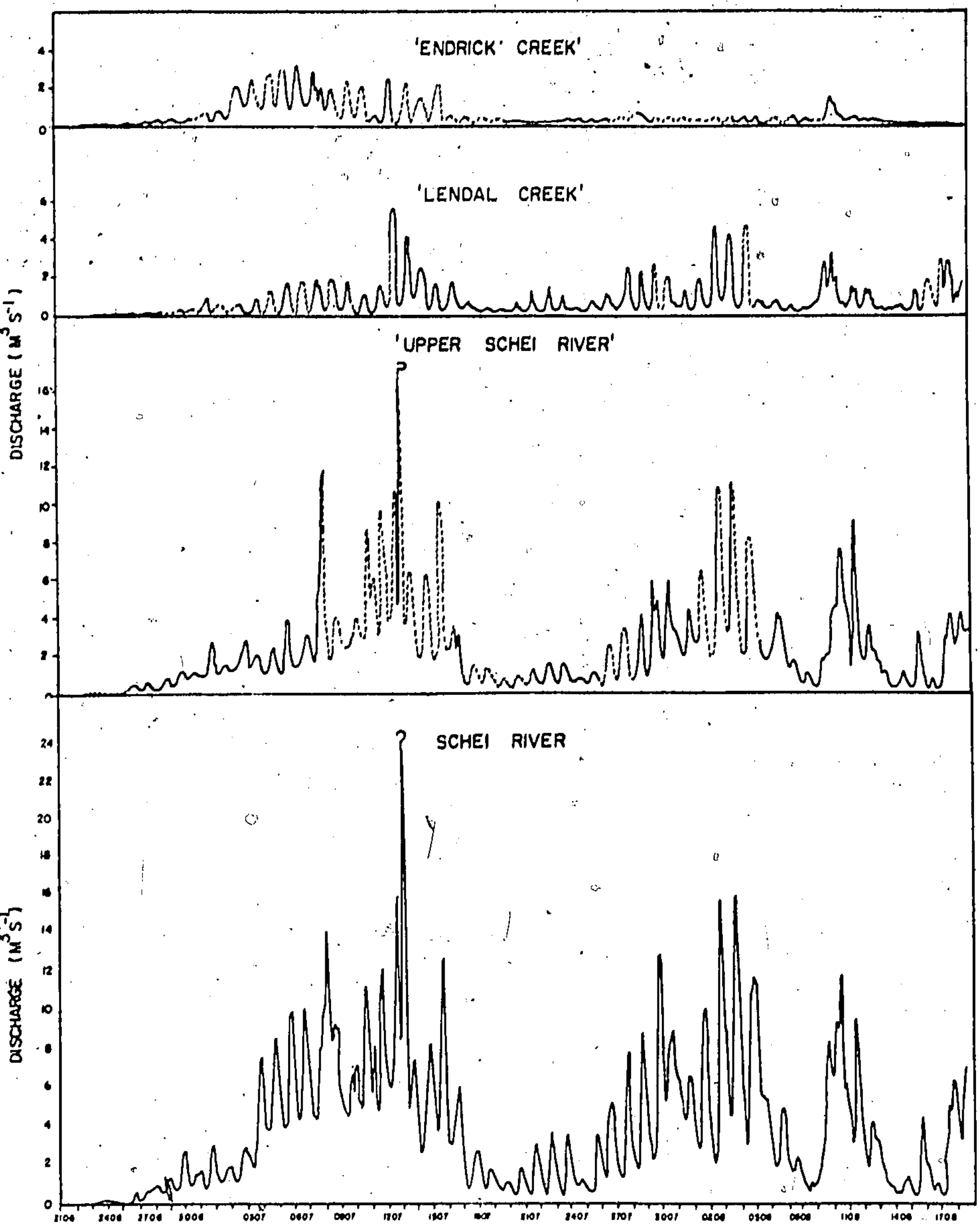


Figure 4.16: Hydrographs for the "Schei River" and its three main tributaries, 1974.

below 0.5 m^3s^{-1}. This difference is attributable to the virtual disappearance of snow in the "Endrick" basin by July 16. After that date flow was sustained by precipitation and the melting of ground ice, neither capable of sustaining discharges in excess of 1.5 m^3s^{-1}. The "Endrick Creek" hydrograph for 1974 is characteristic of a small, high-arctic, unglacierized catchment (Cogley, 1971; McCann and Cogley, 1972, 1974; McCann, Howarth and Cogley, 1972; Walker, Lewis and Lake, 1973; Ambler, 1974; Vieira-Ribeiro, 1975).

The construction of flow-frequency curves (figure 4.17) for the three tributaries highlights the importance of glacial meltwater as a streamflow input. Other factors being equal, the positive skewness of flow-frequency curves is a function of basin size, with the smallest basin in an area exhibiting the greatest positive skewness. In figure 4.17, the "Endrick" curve, reflecting the discharge from a 29.9 km^2 basin, is more skew than that of the "Lendal" which drains a basin only 9.4 km^2 in area. This implies that the supply of water per unit area from the "Lendal" basin is <u>at least</u> three times higher than input per unit area from the "Endrick" basin. The difference can be ascribed to the fact that the "Lendal" basin is 62.8% glacierized, although differences in hypsography (figure 4.3) may also be influential.

4.6 RESPONSE TO METEOROLOGIC INPUTS

Diurnal discharge fluctuations are characteristic of glacial streams (Adams, 1966; Anonymous, 1967) and all but the largest nival streams (McCann and Cogley, 1972, 1973; Church, 1974). Such diurnal fluctuations have been considered a response to variations in radiative

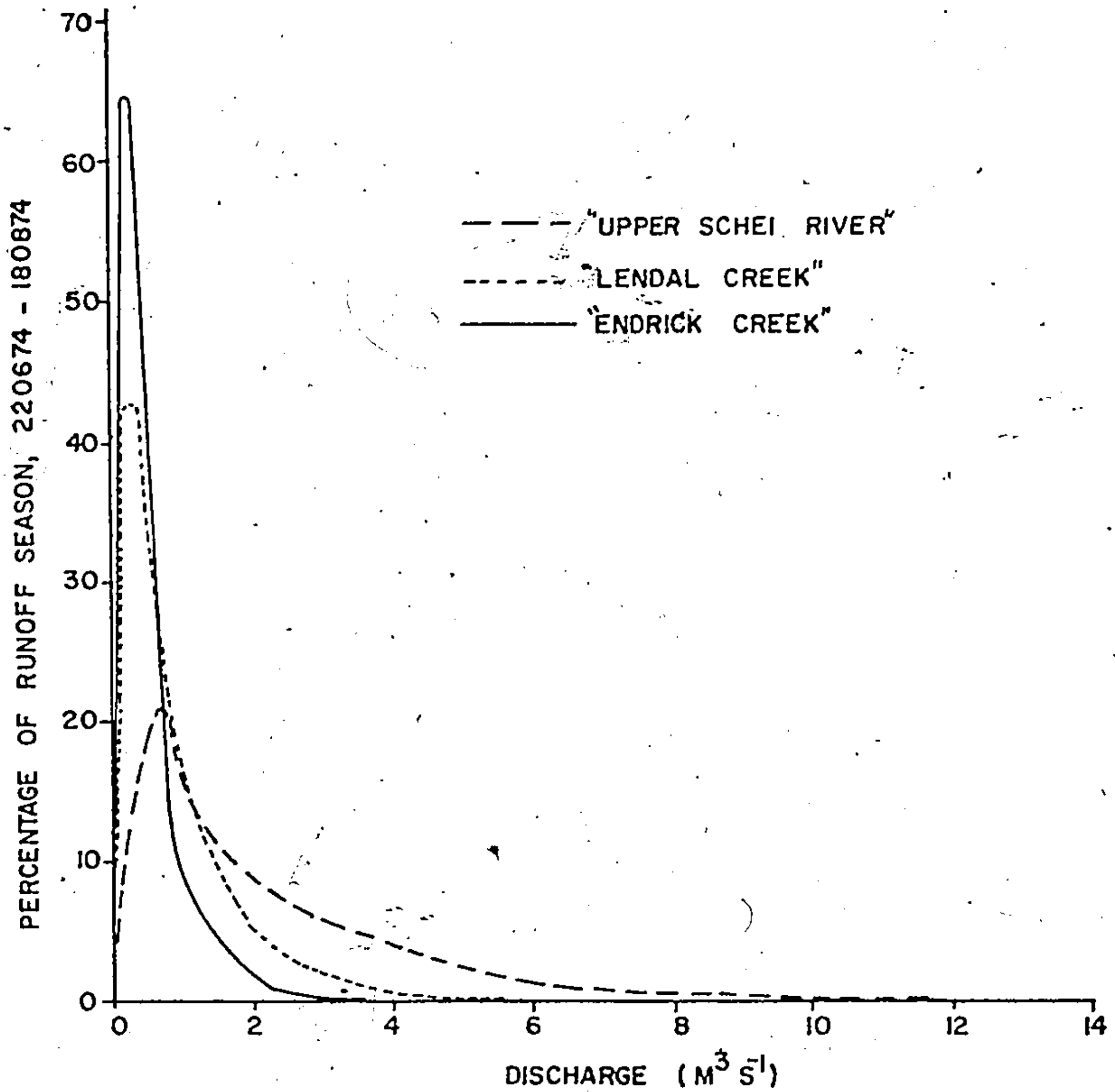


Figure 4.17: Flow-frequency curves for the "Schei River" tributaries, 1974.

input rather than temperature (Adams, 1966; Cogley and McCann, 1971). The diurnal cycle of discharge of the "Schei River" and its tributaries is clearly exhibited by all the hydrographs (figure 4.16 and 4.18). Although there is obvious temporal correspondence between the fluctuations of radiation and temperature and the respondent changes in discharge, (figure 4.11 and 4.19), little correlation is apparent in the values of these three variables, even in the case of the glacial rivers where water supply is effectively unlimited. The hydrographs for the "Upper Schei River" and "Lendal Creek" display two marked periods of high discharge, whereas the temperature graph shows a surprisingly regular cycle roughly ten days from peak to peak and the net radiation data obtained at base camp displays a declining trend throughout the season. The lack of correspondence in these patterns suggests that factors other than temperature and radiation inputs are important in determining short term variations in discharge, factors such as precipitation, antecedent meteorological conditions, snow and ice conditions and temporary retardation of runoff by damming. Some of these factors are discussed in detail below, although precise evaluation of their relative importance is constrained by the data available. In particular, the following considerations hamper assessment of the factors affecting runoff amount.

1. Hydrograph data obtained by irregular stage readings are obviously incomplete, although the closeness of the total seasonal runoff figure calculated from the "Schei River" hydrograph ($19.2 \times 10^6 m^3$) to that obtained by summing the seasonal runoff from the tributary basins and adding a representative factor for the "Lower Schei Basin" (19.8

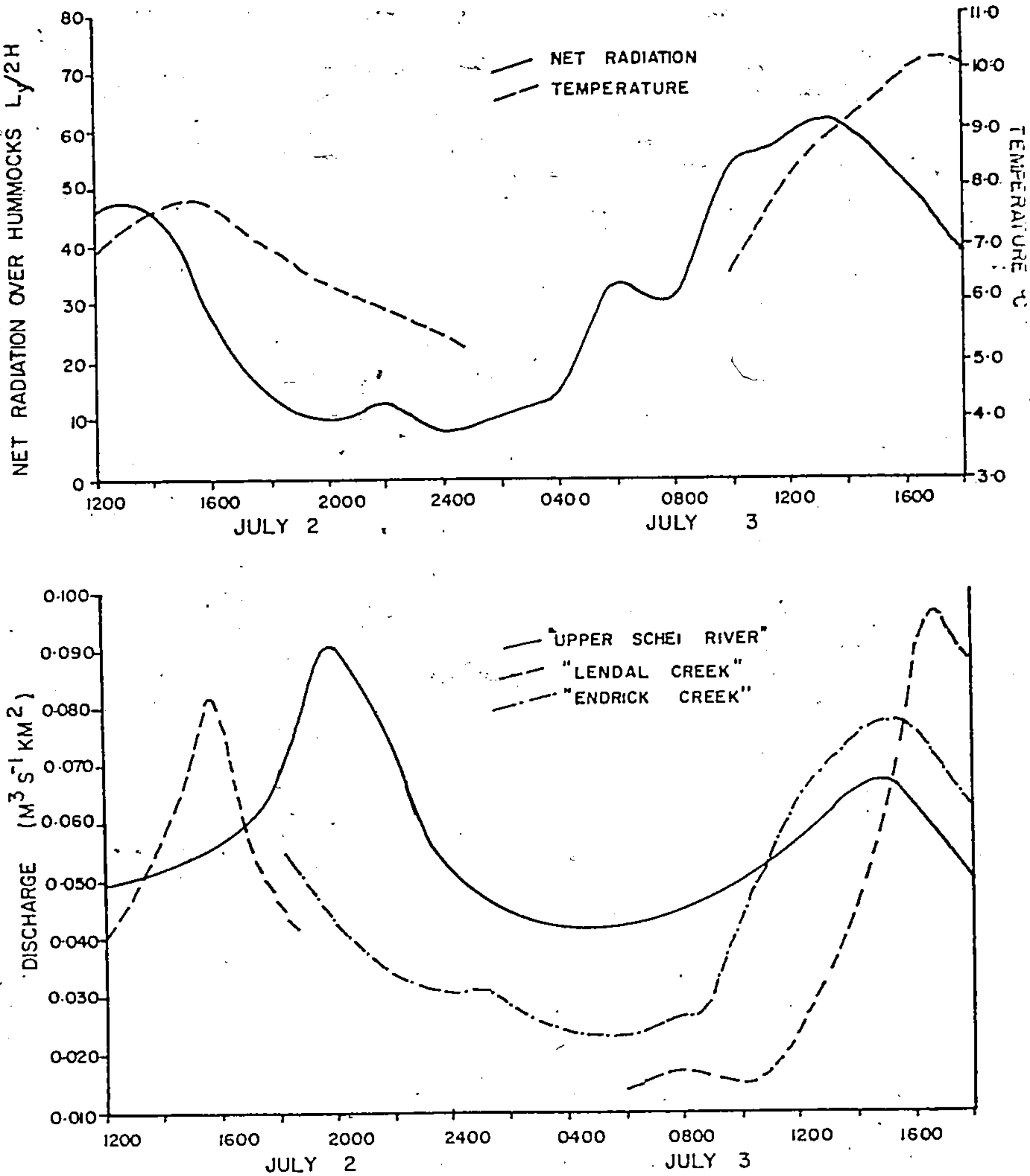


Figure 4.18: Net radiation, temperature and discharge of the "Schei River" tributaries, July 2 - 3, 1974.

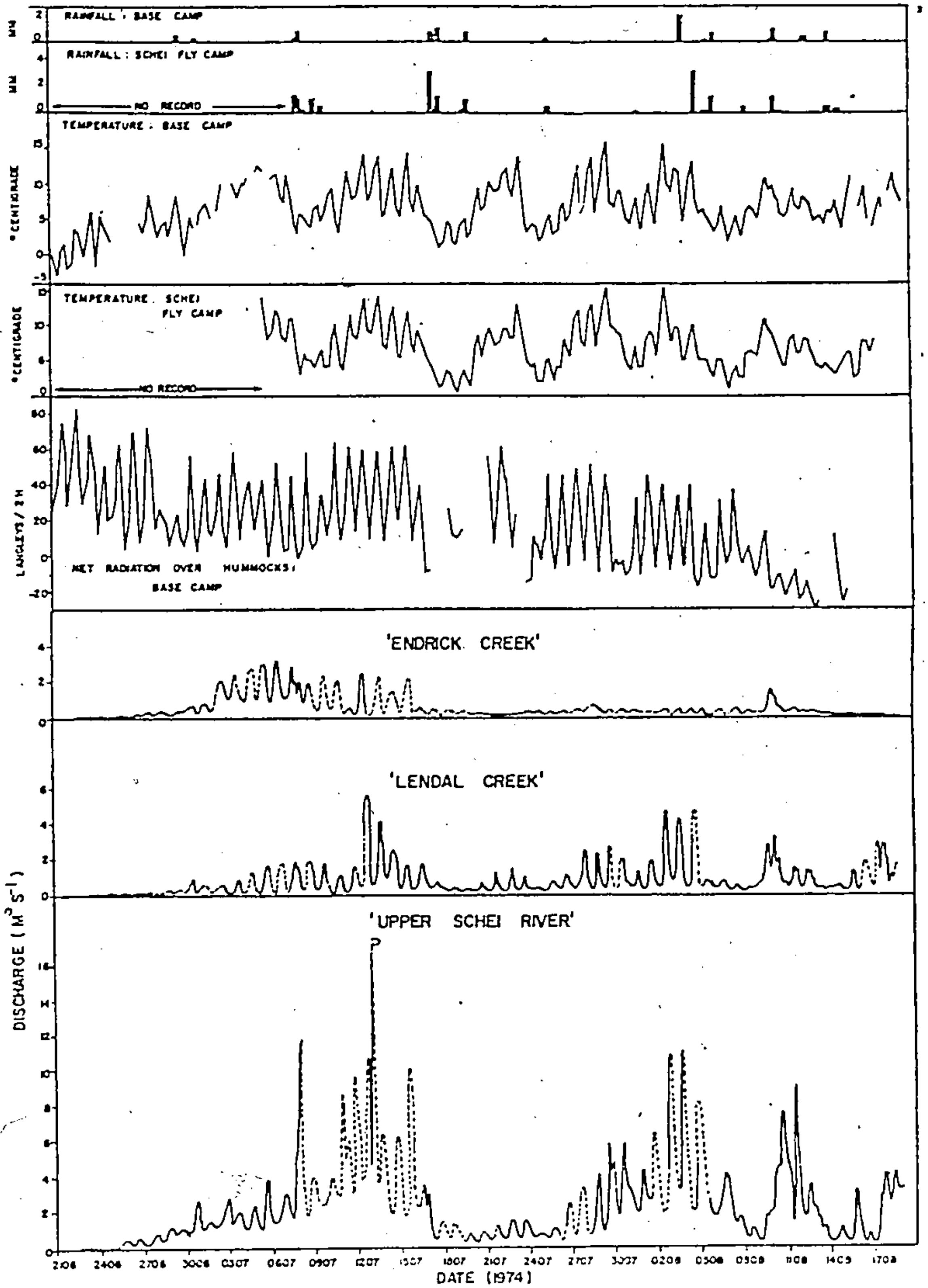


Figure 4.19: Discharge of the "Schei River" tributaries and the meteorologic record for 1974.

x 10^6 m^3) is encouraging.

2. The net radiation data were measured over a level hummock surface at an altitude of 10 m. This location is only partly representative of the "Schei River" catchment.
3. The most representative temperature record was obtained at the fly camp site, 1 km from the glacier at an altitude of 78 m, a location unlikely to reflect conditions in the upper reaches of the basin.
4. The precipitation records are probably underestimates due to the relatively low elevation of the gauges (10 and 92 m) and the tendency of the tipping-bucket apparatus to under-measure precipitation, particularly when rainfall is accompanied by high winds.
5. No information was gathered on snow and ice conditions.

To analyse the time (τ) response of discharge to radiative and temperature inputs, the lags of peak river discharge *(Q max)* behind peak diurnal radiation (R_n *max)* and maximum daily temperature *(T max)* were tabulated (table 4.5). Discharge data for dates when anomalous flow events obscured the diurnal maximum are excluded, and the dates on which precipitation possibly affected the time of peak discharge τ*(Q max)* are indicated.

These tables indicate that τ*(T max)* was relatively unimportant in determining time of maximum discharge as τ*(Q max - T max)* was often negative, implying the occurrence of peak discharge before the maximum diurnal temperature was recorded. It was found that the mean discharge lag behind radiation τ*(Q max -* R_n *max)* for the "Schei River" was 5.15 h

TABLE 4.5

SCHEI RIVER SYSTEM:
LAG OF PEAK DISCHARGE BEHIND PEAK RADIATION AND TEMPERATURE

DATE	SCHEI RIVER				UPPER SCHEI RIVER				LENDAL CREEK				ENDRICK CREEK			
	Lag behind R_n		Lag Behind T		Lag behind R_n		Lag behind T		Lag behind R_n		Lag behind T		Lag behind R_n		Lag behind T	
	H	Min	H	Min	H	Min	H	Min	H	Min	H	Min	H	Min	H	Min
June 25					5	00										
26					0	00										
27	9	30			9	30										
28	7	30			1	00										
29	9	00			8	00										
30*	4	45			3	45										
July 1	7	00	4	00	9	00	4	00								
2*	4	00	3	30	7	00	6	30	1	20	-2	40				
3	5	40	1	00	4	40	0	00								
4	5	15	1	15	5	00	-1	00								
5	6	00	6	20					4	10	-1	50				
6	2	40	3	00												
7	6	05	4	25					4	20	-1	45	3	25	1	45
8	4	10	3	30												
9*	4	40	3	30												
10*	4	20	1	00												
11*	4	50	0	50												
12	5	30	1	30	5	00	1	00	5	15	-0	15				

Continued..

Table 4.5 Continued...

July 13																
14																
14																
16									2	35						
17*																
18*									8	50						
19			2	30												
20*	2	10	2	40					4	25	2	15				
21	5	10	4	30					5	40	0	30				
22	7	30	2	30	6	50	1	50	8	00	0	30	18	00	13	00
23			1	50			0	30							14	30
24	5	20	4	30					5	50	0	30				
25	5	55	2	45					5	50	-0	05				
26*	4	10	-0	20					4	30	0	20				
27	7	10	1	50					8	30	1	20				
28	5	00	1	50					6	20	1	20				
29																
30	0	30	1	30	-0	15	0	45	-1	00	-1	30				
31			-0	50			-1	35			-1	00				
Aug. 1	5	20	-0	20					5	20	-0	20				
2									5	45	1	15				
3	4	30	4	00					3	30	3	00				
4*																
5*									3	00	-0	10				
6	4	10	3	20					4	00	-3	10				
7	4	30	3	10					4	00	2	40				
8					2	20	-6	40	0	40	-8	20				
9	5	15	1	05					0	50	-3	20				
10*					3	10	2	40								
11					0	00	-1	10	1	20	0	10				
12					1	00	-2	40	3	00	-0	20				
13							2	30			0	00				
14*	6	40	1	30	5	25	-0	15	4	20	-0	50				

Continued...

Table 4.5 Continued...

Aug. 15 16 17 18		3 15 4 00		1 45 1 20		1 45	
Mean of all Measurements:	5 10	2 25	4 15	0 20	4 15	-0 20	Not Calculated
Standard Deviation (σ):	2 05	1 40	2 35	2 50	2 20	2 15	Not Calculated

(σ = 2.25 h), for the "Upper Schei River" 4.25 h (σ = 2.6 h), and for "Lendal Creek" also 4.25 h (σ = 2.3 h). Too few values were obtained for "Endrick Creek" to warrant an equivalent calculation. These figures suggest that the lag of *Q max* behind *R max* is similar for "Lendal Creek" and the "Upper Schei River", and that average travel time from the "Lendal" and "Upper Schei" gauging stations to the main "Schei River" station is 55 minutes. The high standard deviations (σ) indicate, however, that variability from the mean condition is considerable.

The difference in response lag between two gauging sites is an inverse nonlinear function of water velocity and hence discharge. This was confirmed by timing the passage of dye released at the "Upper Schei River" gauging station. The peak of a wave of one litre of Rhodamine dye released at 1150 h on July 31 passed the "Schei River" gauging site at 1310 h, a travel time of 80 minutes between the two points when the initial "Upper Schei" discharge was 0.6 $m^3\ s^{-1}$. The peak of a wave of dye injected when discharge of the "Upper Schei River" was 4.0 $m^3\ s^{-1}$ at 1800 h on August 1 took only 33 minutes to reach the "Schei River" gauging site.

The identical mean lag of *Q max* behind R_n *max* for the "Upper Schei River" and "Lendal Creek" is somewhat puzzling, in view of the much smaller size of "Lendal Creek" basin. The difference may reflect higher velocities of flow in the "Upper Schei River" system.

It is apparent from table 4.5 that the lag of peak discharge behind maximum net radiation varied widely, even on days when the influence of precipitation was absent. The times of maximum discharge on such days (τ *max*) are plotted on figures 4.20 and 4.21. For the "Schei River", regression of the

Figure 4.20:

Time of diurnal peak discharge on the "Schei River", 1974.

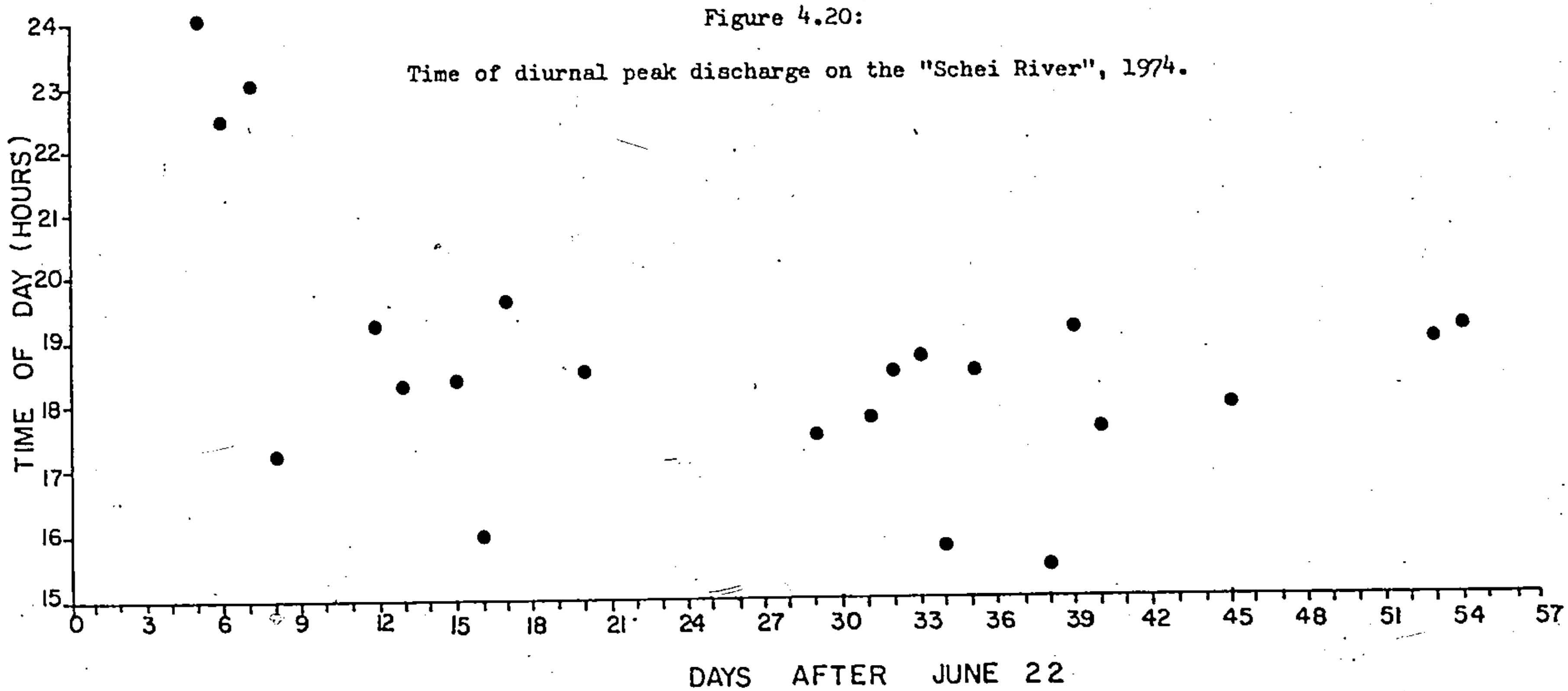
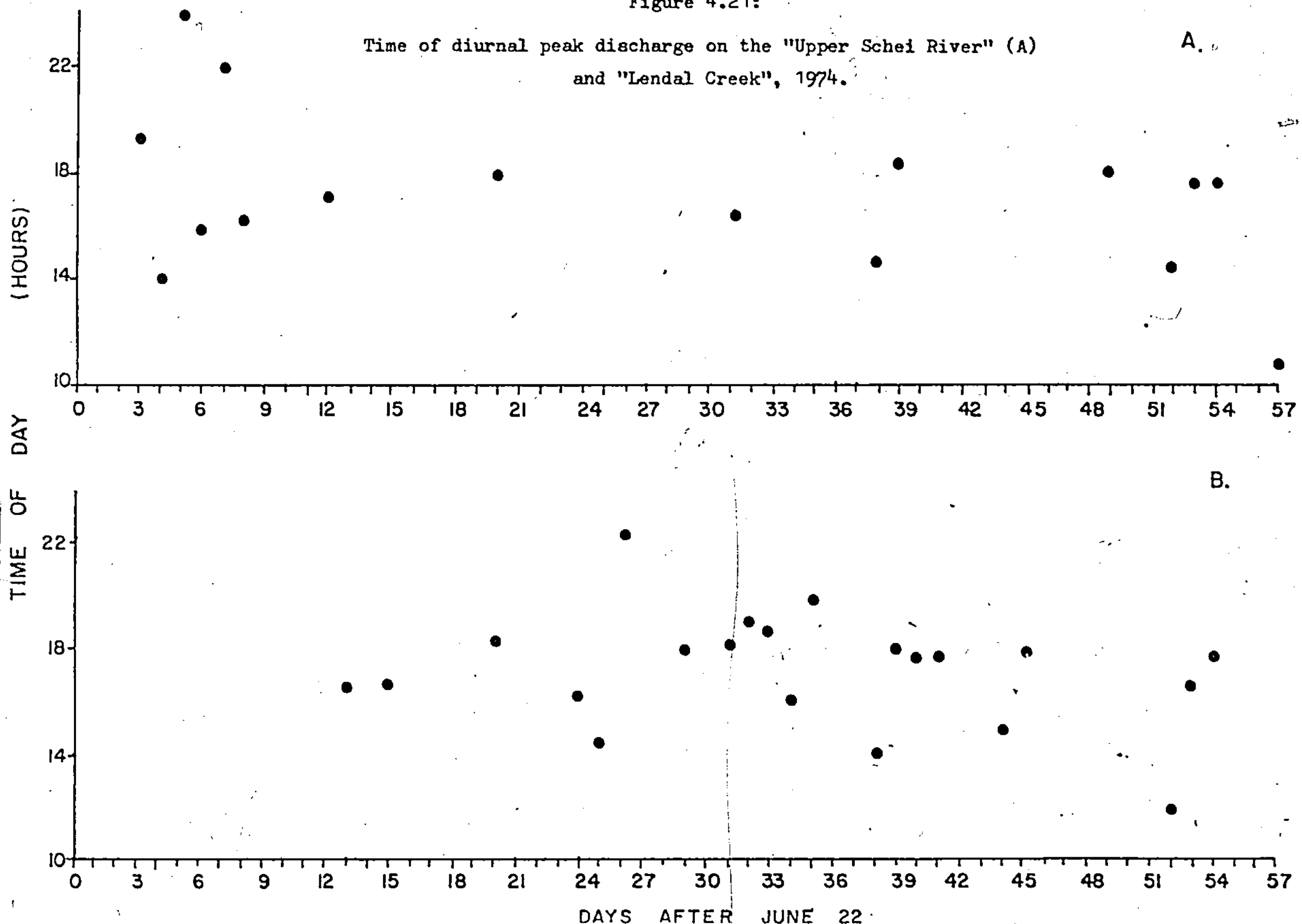


Figure 4.21:

Time of diurnal peak discharge on the "Upper Schei River" (A) and "Lendal Creek", 1974.



values of $\tau(Q_s\ max - R_n\ max)$ against time t (days after June 22) yielded a negative exponential relationship

$$\tau(Q_s\ max - R_n\ max) = 10.12t^{-0.28}$$

$$r = -0.285 \qquad (4.1)$$

significant, however, only at the 70% level and therefore probably not meaningful. A 98% significance is achieved by regressing $\tau\ Q_s\ max$ against t, giving

$$\tau\ Q_s\ max = 23.99\ t^{-0.088}$$

$$r = -0.521 \qquad (4.2)$$

With the "Upper Schei River" figures, regression of $t(Q_u\ max - R_n\ max)$ against t produces no significant relationship, although

$$\tau\ Q_u\ max = 20.37\ t^{-0.065}$$

$$r = -0.371 \qquad (4.3)$$

is significant at the 80% level. Insufficient data are available on the discharge of the "Lendal" and "Endrick" to allow the same analysis to be carried out for these rivers.

An increasingly early diurnal peak discharge has been observed in several studies of glacial runoff (e.g. Wendler, Trabant and Benson, 1973), and has been attributed to the increasing efficiency of the glacier channel network as the ablation season progresses. The simple negative exponential relationship found in the present instance may however obscure a slightly more complex pattern: it seems possible from figures 4.20 and 4.21 that the mean response time of the glacial streams follows a steeper exponential decrease than those obtained by regressing the entire data set, then levels off at some constant mean lag value. Unfortunately the data are insufficient to allow this to be tested. If this were the case, however, the end of the period of decreasing lag would appear to lie (figure 4.20) roughly between the 18th and 27th days of runoff (July 10 - 19) thereby corresponding with the apparent termination of the snowmelt freshet, and suggesting that with the disappearance of snow from the glacier surface, the mean lag of meltwater discharge becomes constant. The correspondence may be coincidence, however.

Although the precise time of diurnal peak discharge on "Endrick Creek" was only measured on three occasions, data on periods of increasing and decreasing discharge indicated that the lag of discharge in response to radiative and temperature inputs increased very considerably throughout the season, being 18 hours on July 30, but less than 4 hours on July 7. No information is available on the manner of increase, however, although in the similar circumstances of a nival catchment on Baffin Island, Vieira-Ribeiro (1975) found a linear increase in lag throughout the runoff season. This is probably due to the retreat of the snowline, such that snowmelt input is received from progressively higher parts of the basin. It may also be due in lesser measure to the generally lower discharge (hence lower velocity) of the stream after the end of the snowmelt freshet.

Rainfall in excess of 0.25 mm was recorded at base camp on 13 days during the 1974 season (figure 4.6), but the magnitude of individual falls was low. The largest precipitation event recorded at the "Schei" fly camp occurred on July 17 when 4.58 mm of rain were recorded between 0500 h and 2350 h. Although a single raingauge provides an inadequate measure of precipitation in a basin 91.2 km^2 in area the precipitation record provides an explanation for at least some of the observed fluctuations in river discharge.

The complexity of the meteorologic processes generating runoff in the "Schei River" basin hinders the identification of response to specific precipitation inputs (figures 4.18 and 4.22). For this reason no attempt was made to subtract baseflow from precipitation response, and runoff ratios for the "Schei River" tributary basins were not calculated. Some comments on the response to precipitation of these streams may however be made from an examination of figures 4.17 and 4.22. The influence of precipitation is difficult to discern on the hydrographs of the glacier-fed rivers, but its effect on the runoff of "Endrick Creek" after the period of snowmelt flood might be expected to be apparent. Interestingly, this is true of the rainfall of August 10, but not that of August 4. A good stage record was obtained for "Endrick Creek" on both occasions. These differences in basin response to rain may result from the condition of the active layer at the time of rainfall. The rain on August 4 may have recharged the moisture content of the active layer, with minimal immediate contribution to runoff. On August 10, a much higher proportion of the rain may have run off rapidly in consequence of the active layer being near-saturated from the earlier fall. Alternatively, the rainfall record may not reflect accurately the precipitation over the "Endrick Creek" basin during these periods. A third possibility is

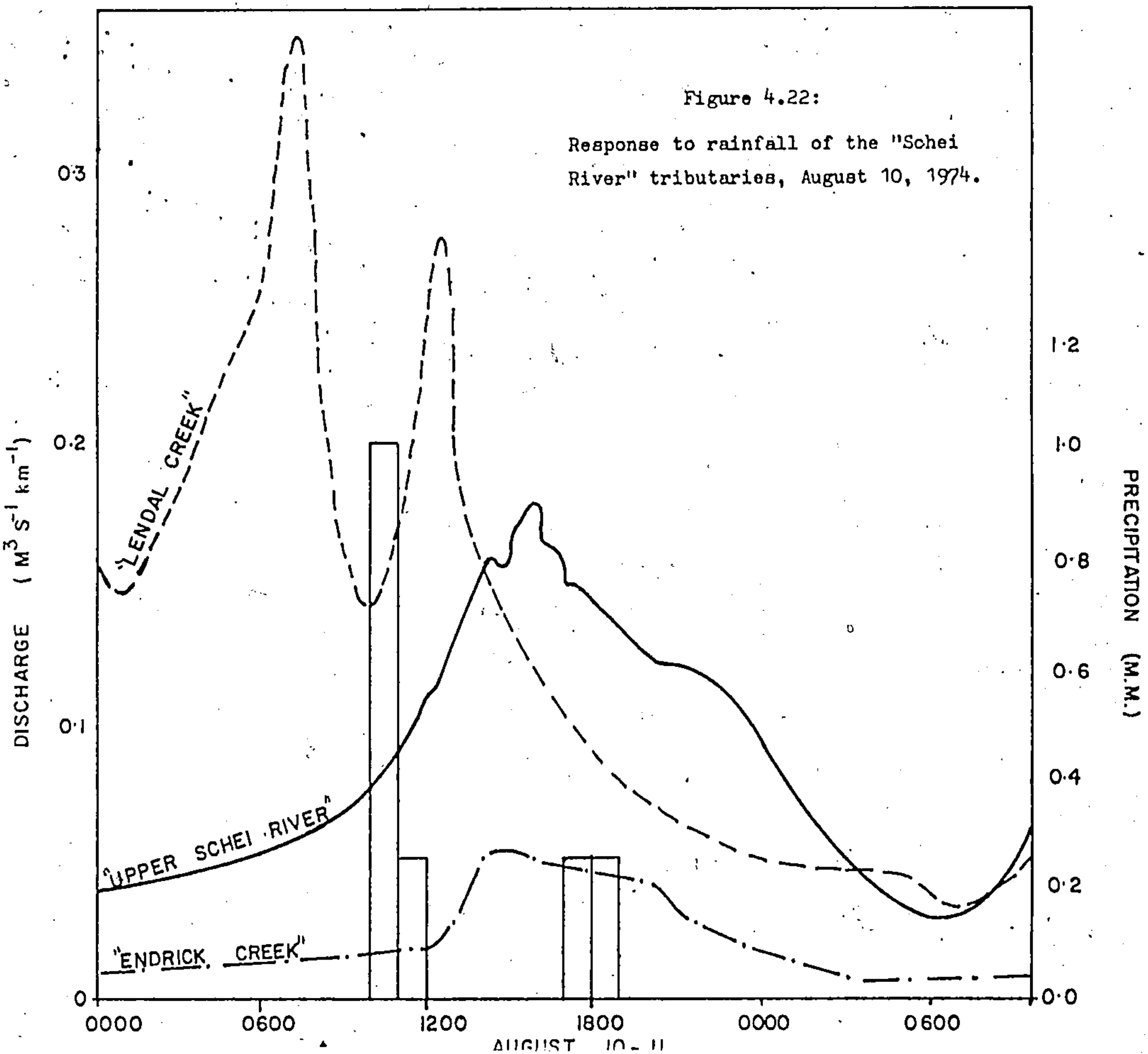


Figure 4.22:

Response to rainfall of the "Schei River" tributaries, August 10, 1974.

is that some of the rain on August 4 fell as snow on the high ground of the basin (figure 4.3), melting only gradually.

The precipitation of August 10 and the three tributary hydrographs for that date are plotted in figure 4.22. The trace of "Lendal Creek" discharge is highly irregular, possibly reflecting the influence of antecedent rainfall not recorded at the Schei fly camp and no conclusions about river response can be drawn from the "Lendal" hydrograph. "Endrick Creek" and the "Upper Schei River" show a clearer response to precipitation, with lag times of 4-5 h and 5.5 h respectively, although the peak discharge of the "Upper Schei River" at 1540 probably reflects in part the "normal" lag of the system to melwater inputs. The 5.5 h lag of discharge behind precipitation of the"Upper Schei River" system suggests that lag in response to rainfall resembles lag behind other runoff-generating inputs.

4.7 DISCHARGE IRREGULARITIES

4.7.1 Introduction

Up to this point, the discussion of discharge has centred around the influence of meteorologic inputs on streamflow. Other factors may however influence short-term hydrograph patterns. Analysis of the fine structure of the "Schei River" hydrograph for 1973 by McCann et al (1974) revealed the occurrence of irregular fluctuations in discharge not attributable to meteorologic influences. Explanations for these irregularities were offered in terms of "resonance phenomena associated with the shape of the basin in which the stilling well was located" and "contributions of meltwater from different parts of the drainage basin which arrive at the recorder site as separate waves of water." The cause of similar discharge irregularities recorded on the "Schei River" in 1974 forms the subject of the present section.

4.7.2 Discharge irregularities on the "Schei River", 1974

The first observed flow anomaly on the "Schei River" in 1974 occurred on July 7-8. Around midnight a flood wave passed down the "Upper Schei River". Discharge at the "Upper Schei" gauging station increased rapidly from 4.0 m^3s^{-1} to 10.0 m^3s^{-1} before the stilling well was destroyed, and at the main "Schei River" gauging station a peak discharge of 14.0 m^3s^{-1} was recorded. Both "Endrick Creek" and "Lendal Creek" were under constant surveillance at the time of the flood, and neither exhibited anomalous behaviour. These observations made it clear that the explanations for discharge irregularities offered in McCann et al (1974) were inadequate, as the flood event of July 7-8 was apparently caused by the rapid release of water in the "Upper Schei" basin.

This was confirmed by observations made at the onset of the flood of July 12, during which the highest discharge of 1974 runoff season was measured. At the "Upper Schei" gauging site discharge peaked at 10.3 m^3s^{-1} at 1830 h, then dropped dramatically to 8.0 m^3s^{-1} by 1910 h, and 4.5 m^3s^{-1} by 1940 h. At 2000 h, however, discharge was 6.2 m^3s^{-1} and rising rapidly, with large blocks of glacier ice being carried downriver. No further observations were made that evening at the "Upper Schei" station, but shortly afterwards the stilling well of the main "Schei River" gauge was destroyed by water discharges in excess of 20 m^3s^{-1}. Again, flooding was restricted to the "Upper Schei" tributary.

Four types of discharge irregularities were identified on the

"Schei River" stage record for 1974 (figure 4.23):

1. An abrupt drop in discharge followed immediately by an equally abrupt rise to levels above "normal".
2. An abrupt fall of discharge followed by an abrupt rise, but with a period of "normal" flow intervening.
3. An abrupt drop in discharge succeeded by a period of rapid, small discharge fluctuations.
4. An abrupt rise in discharge not preceded by a previous fall.

All four types were recorded on the "Upper Schei River" stage record, but no similar events occurred on the other tributaries. The periods of abruptly rising discharge were invariably accompanied by the appearance of blocks of glacier ice on the river (figure 4.24).

The above evidence suggested that the cause of the discharge irregularies was some form of temporary retardation of flow down the "Upper Schei River". Reconnaissance of the south margin of the "Schei Glacier" on August 9 revealed that the "Upper Schei River" had cut a slot of up to 25 m under the glacier margin, and in several places masses of glacier ice 15-20 m thick had collapsed into the river. Such ice collapses afford an explanation for the short-lived damming of the "Upper Schei River", and the stage record patterns 1 to 3 described above. The type 1 stage trace probably represents the ponding of water by ice collapse and its immediate release when the ice dam breaks. The intervening "normal" flow of type 2 suggests overtopping of the ice dam before breaching occurs. The lack of a significant flood following a drop in discharge (type 3) indicates slow destruction of the ice dam, possibly by melting, with no

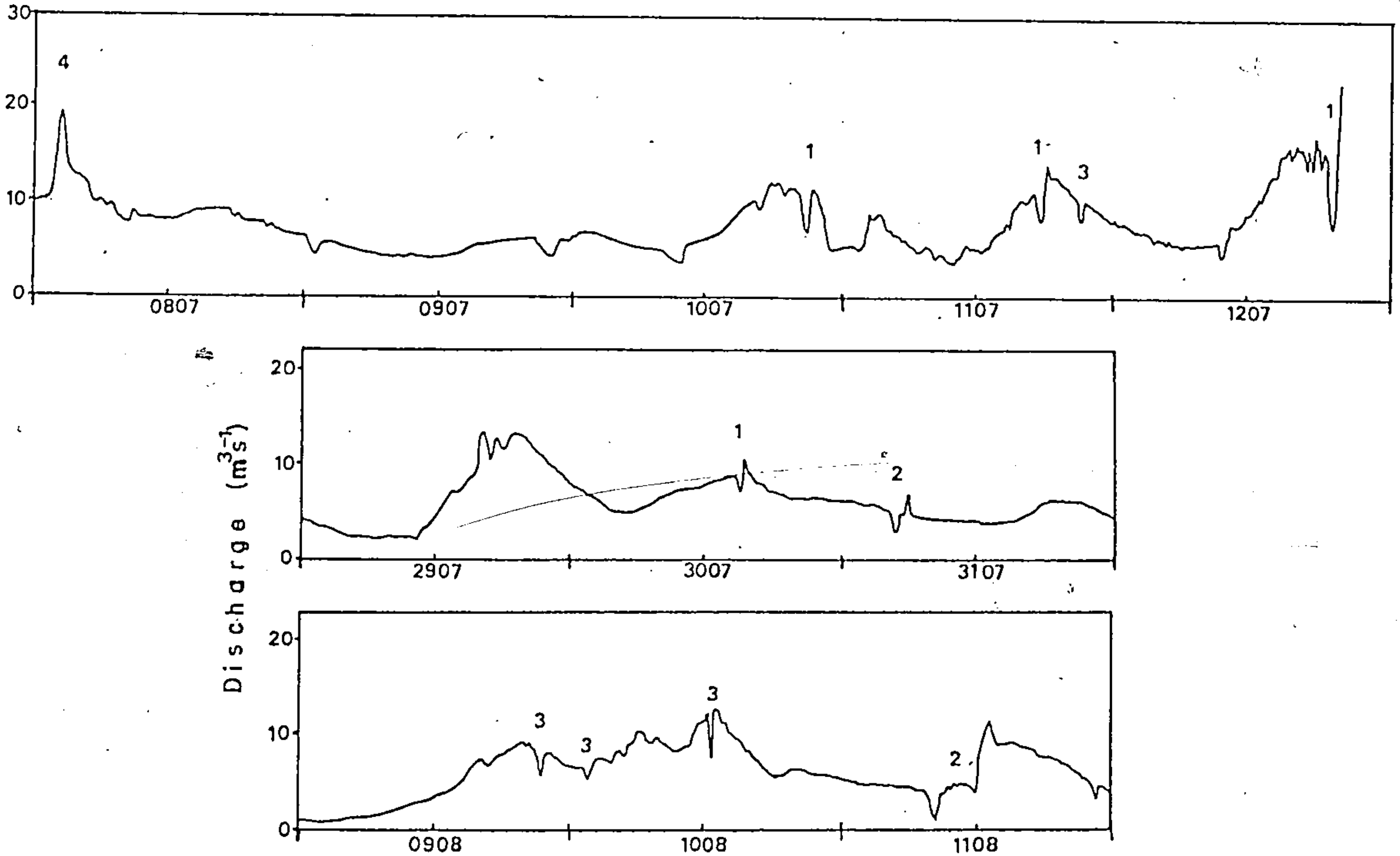


Figure 4.23: Short-term discharge irregularities on the "Schei River", 1974. Hydrographs for July 8-12, July 29-31 and August 9-11 showing examples of each type of discharge irregularity.

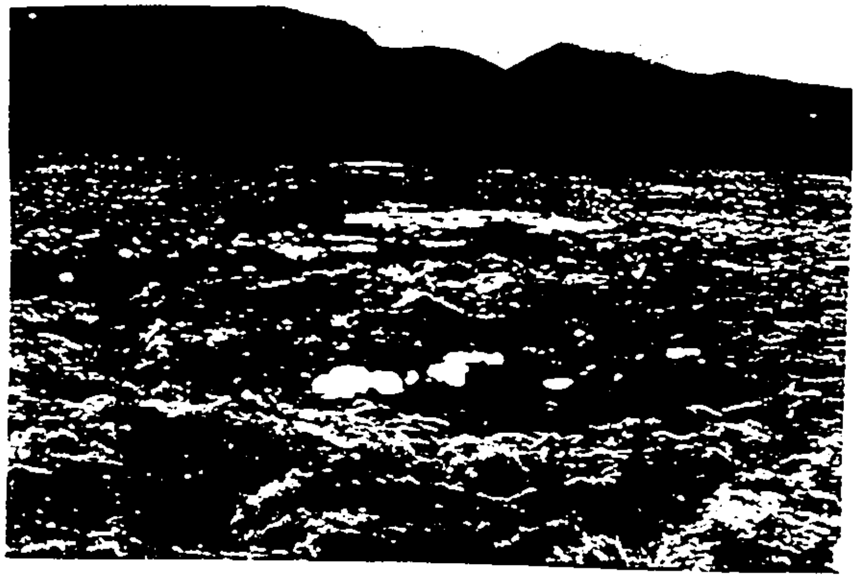


Figure 4:24 : Blocks of ice being carried down the "Upper Schei River" following the breaking of a marginal ice dam, July 30, 1974.

Figure 4.25 : Site of drained ice-dammed lake at the south margin of the :Schei Glacier" (photograph by S.B. McCann)

single abrupt release of water.

Only the flood of July 7-8 fell into category 4, being unheralded by a preceding drop in discharge. The probable explanation for this flood lay in the discovery of a small lake dammed against the south margin of the glacier. Helicopter observations on July 2 revealed the presence of a considerable body of water ponded at the glacier margin. Aerial photographs taken on July 10 show a lake at least 160 m long and 25 m wide at this location, but when visited on August 9 the lake was only 45 m long and less than 1 m deep (figure 4.25). Stranded icebergs extended to a height of 3 m above the lake surface, however, and strandline evidence suggested that, when full, the lake would be 200 m long, 150-200 m wide and over 15 m deep, with a maximum capacity of 6×10^6 m^3. Only a fraction of this capacity would be necessary to nourish a small jokulhlaup like that of July 7-8.

Analysis of the size and frequency of the discharge irregularities on the "Schei River" hydrographs for 1973 and 1974 revealed three tendencies:

1. Damming at the glacier margin was more frequent towards the end of the runoff season, presumably after the river has melted a deep slot under the glacier margin.
2. Damming was most frequent in the latter half of the day when discharge was generally higher.
3. There was an inverse relationship between the frequency of events and their magnitude.

4.7.3 Discussion

The temporary damming of marginal glacial streams by an ice, slush or snow barrier has been documented by Adams (1966), Church (1972) and Wendler, Trabant and Benson (1973). The 1961 hydrograph for the marginal Ermine River on Axel Heiberg Island (Adams, 1966) shows several characteristically abrupt discharge fluctuations that mark the occurrence of temporary damming.In all of the documented cases, however, the size of individual irregularities has been small in comparison with other flood events, such as peak snowmelt discharges. This is also true for most of the 24 discharge irregularities noted on the 1974 "Schei River" hydrograph. Of greater interest and geomorphic significance are larger flood events, such as those of July 7-8 and July 12.

It is likely that both of these flood events involved the sudden release of water dammed in the small ice-marginal lake described above, as at no other point on the south margin of the "Schei Glacier" could sufficient water have been dammed by falling ice to produce floods of such magnitude. The hydrograph of the "Schei River" for July 7-8 has all the attributes of a typical "jokulhlaup" or glacier burst (Maag, 1969; Blachut and Ballantyne, 1975 in preparation) with a period of exponentially increasing discharge ending abruptly as lake level drops below the level of the lake outlet. On July 12, however, a sudden drop in discharge preceded the flood, indicating that ice collapse near the outlet caused blockage of the marginal stream and a build-up of lake water which was subsequently released as a jokulhlaup. A large block of collapsed ice, 20 m in height, was observed at the lake outlet during the August 9 reconnaissance.

Although the occurrence of jokulhlaups is well documented for lakes dammed by temperate ice, the work of Maag (1963, 1969) on Axel

Heiberg Island suggested that the normal mode of drainage for lakes dammed by polar glaciers involves outflow along supraglacial or ice-marginal channels. Drainage is maintained through the constant down-melting of such channels until the end of the runoff season. A small pro-glacial pond in the "Lendal" basin (figures 4.26, 4.27 and 4.28) conformed to this drainage pattern, draining gradually throughout the season as the outlet waters melted a progressively deeper "slot" under the damming ice. The catastrophic drainage of a lake dammed by polar ice has recently been reported from West Central Ellesmere Island (McCann et al., 1974, 1975; Blachut, 1975 in preparation), but the circumstances under which this took place were exceptional.

Three drainage mechanisms have been proposed to account for the occurrence of jokulhlaups from lakes dammed by temperate ice : floatation of the ice barrier (Thorarinsson, 1939b; Aitkenhead, 1960; Marcus, 1960; Howarth, 1968); melting of an outlet tunnel (Liestol, 1956; Mathews, 1965, 1973; Gilbert, 1971, 1972); and plastic deformation of the ice dam induced by hydrostatic pressure (Glen, 1954). The theory behind the last demands the existence of a water body at least 200 m deep, and this explanation can be discounted in the present case. The rapidity with which the flood of July 12 followed the fall in discharge (hence ice collapse) indicates that melting is unlikely to be important, except in the case of "type 4" floods. Floatation offers a more plausible explanation, but it seems equally likely that the ice-dam, being no longer "attached" to the glacier, simply ruptured as pressure built up behind it. Such an occurrence would explain the abundance of glacier ice carried down the river during the rise ~~of~~ discharge to the flood peak.

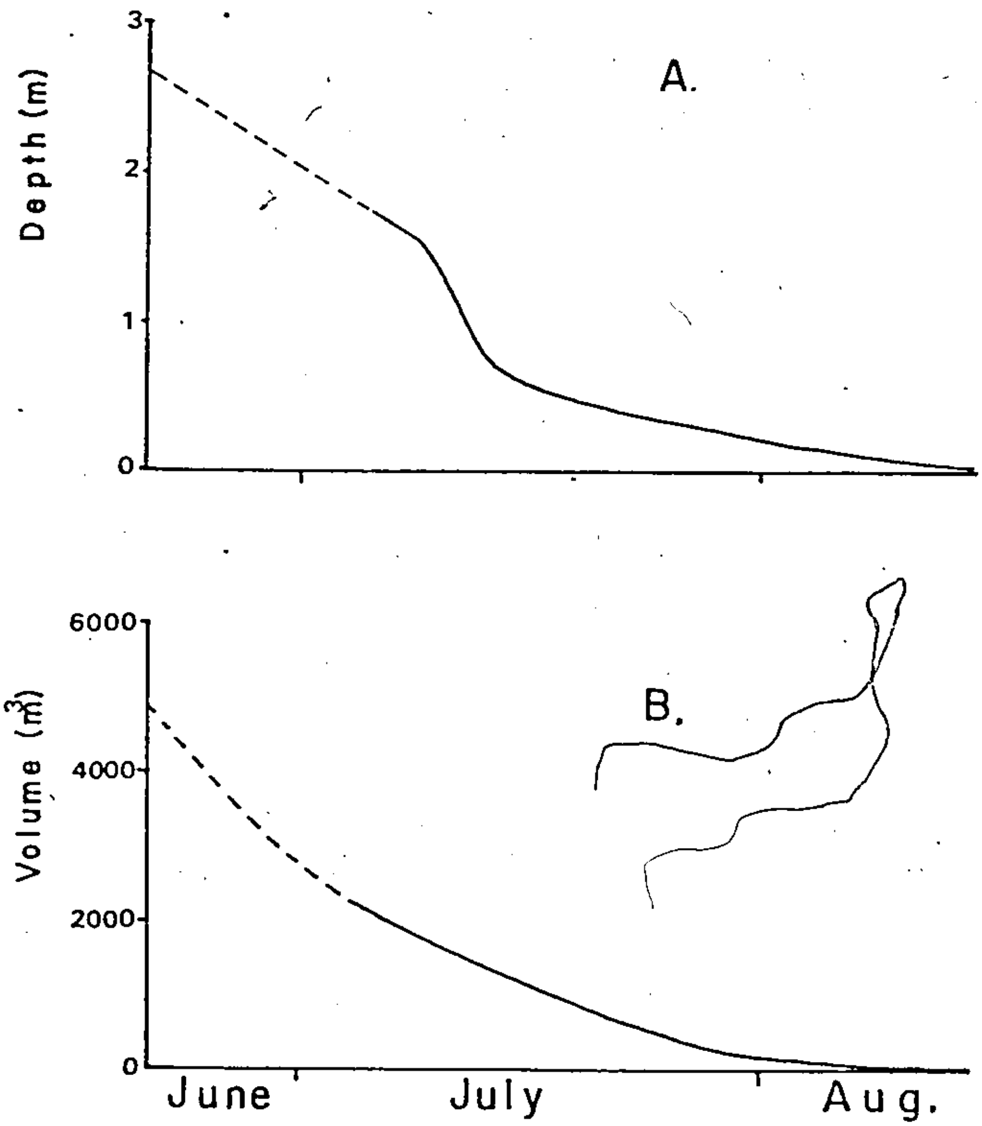


Figure 4.26: drainage of an ice-marginal pond in the "Lendal" basin, 1974. A is drop in water level; B is estimated drop in volume.

Figure 4.27 : sub-marginally draining ice-dammed pond in the "Lendal" Basin, July 5, 1974.

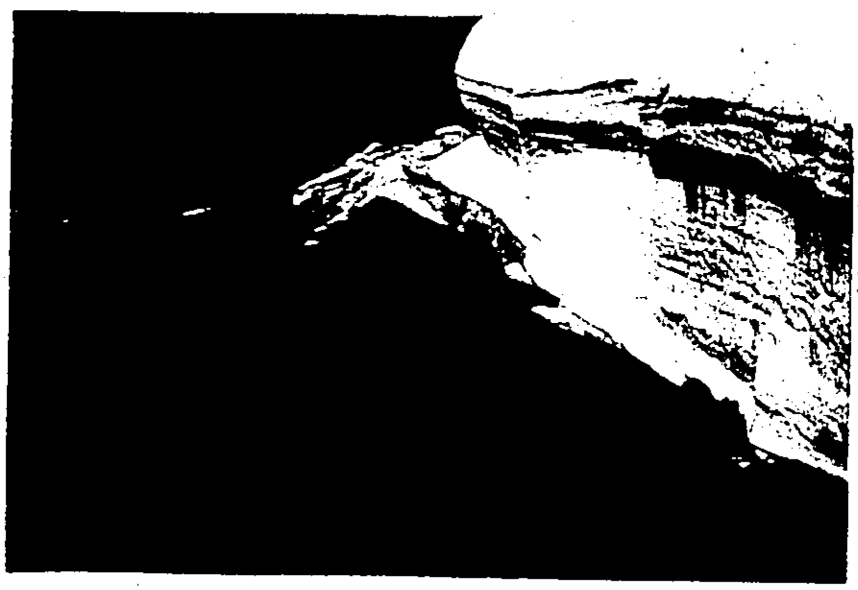


Figure 4.28 : sub-marginally draining ice-dammed pond in the "Lendal" Basin, August 13, 1974.

The temporary retardation of flow by ice collapse at a glacier margin constitutes a rather different situation from that under which jokulhlaups normally occur. The result, however, is the same : catastrophic floods of unpredictable timing and magnitude. Such floods deserve future study as events of considerable geomorphic and hydrologic significance.

4.8 WATER BALANCE OF THE "SCHEI RIVER" SYSTEM

The inputs to streamflow in the "Schei River" system can be expressed as a simple equation

$$G_s + S_s + P_s + M_s - E_s = QT_s \qquad (4.4)$$

where: G_s is total input of water from the glacier (glacier melt, snowmelt and precipitation on to the glacier surface);

S_s is snowmelt in the non-glacierized portion of the basin;

P_s is precipitation over the non-glacierized portion of the basin

M_s is permafrost melt;

E_s is flux of water vapour (evaporation); and

QT_s is the total discharge of water passing the "Schei River" during the study period, June 22 to August 18.

The subscript s denotes "Schei River" system. A lake storage term has been omitted from the equation as storage in lakes is negligible in the "Schei River" basin.

Similar simple water-balance equations for the three tributary catchments can be written as

$$G_u + S_u + P_u + M_u - E_u = QT_u \tag{4.5}$$

$$G_\ell + S_\ell + P_\ell + M_\ell - E_\ell = QT_\ell \tag{4.6}$$

$$S_e + P_e + M_e - E_e = QT_e \tag{4.7}$$

where the subscripts u, ℓ, and e refer respectively to the "Upper Schei River", "Lendal Creek" and "Endrick Creek" basins. It should be noted that equation (4.7) contains no G term; this equation pertains to an unglacierized basin.

Accurate assessment of the relative importance of the terms in these equations is precluded by the inadequacy of the data available, but some generalisations may be hazarded. Of the terms in the general equation

$$G + S + P + M - E = QT \tag{4.8}$$

E and M may be considered with reference to the similar environment of the Mecham River basin, Cornwallis Island, where Cogley (1975 in preparation) has found that evaporative loss exerts a very substantial influence on the water balance. Evaporation may however be ignored in the present context, which primarily concerns inputs to streamflow. Cogley has also argued that although the contribution of permafrost melt (M) may be significant, the active layer is likely to retain an equal amount of precipitation, such that the net contribution from permafrost melt over the runoff season is small or even negative. With the exclusion of these two terms, the general equation becomes

$$G + S + P = QT \tag{4.9}$$

It is possible to arrive at an estimate of the seasonal discharge from the glacierized portion of those basins with partial ice cover by considering the total (seasonal) discharge per unit area of the "Endrick Creek" basin $(QT_e A_e^{-1})$ as representative of the unglacierized portions of the "Schei River", "Upper Schei River" and "Lendal Creek" basins. If A represents basin area and A' is area under ice cover, then

$$G_s = QT_s - \frac{QT_e (A_s - A'_s)}{A_e} \tag{4.10}$$

$$G_u = QT_u - \frac{QT_e (A_u - A'_u)}{A_e} \tag{4.11}$$

$$G_\ell = QT_\ell - \frac{QT_e (A_\ell - A'_\ell)}{A_e} \tag{4.12}$$

The values of G_s, G_u, and G_ℓ obtained from these equations are shown in table 4.6.

The main sources of error in this calculation arise (1) from the assumption that runoff per unit area of the "Endrick Creek" basin $(QT_e A_e^{-1})$ approximates the runoff per unit area of the other tributary basins, and (2) from the unreliable nature of the "Endrick Creek" discharge record. If these deficiencies are accepted, the calculation permits a rough calibration of the difficult G term of the water balance equations.

Equally broad estimates may be made for the P term by considering the Schei fly camp rainfall total as representative of the entire "Schei River" catchment area. The total was first adjusted using the base camp record to represent the full period June 22 – August 18. The calculated

TABLE 4.6

TOTAL WATER INPUT FROM THE GLACIERIZED PORTION OF SCHEI RIVER BASIN, JUNE 22-AUGUST 18

River	Subscript	Total Discharge (TQ) in $10^6 \times m^3$	Glacial Discharge (G) in $10^6 \times m^3$	QTA^{-1} in 10^6 m^3 km^{-2}	GA^{-1} in 10^6 m^3 km^{-2}	G/QT(%)
Schei	s	19.867	14.069	0,218	0.434	71%
Upper Schei	u	12.094	9.856	0.246	0.371	81%
Endrick	e	2.948	-	0.099	-	-
Lendal	l	4.568	4.223	0.486	0.716	92%
[Lower Schei]		0.256	-	0.099	-	-

total for this period becomes 21 mm, so that PA^{-1} is 21,000 m^3km^{-2}, assuming that all precipitation is translated to runoff, and none is stored as ice or snow or in the active layer.

An estimate for the final unknown term, S can now be simply calculated from

$$S = QT - G - P \tag{4.13}$$

The value of SA^{-1} calculated in this way is 77,500 m^3km^{-2}, representing the melt of 77.5 mm water equivalent of snow.

As values have now been obtained for PA^{-1} and SA^{-1} for each basin, it is possible to use these values to break up the G term into its component parts, namely glacier melt, snowmelt and precipitation. If the terms of the general equation are redefined, such that

γ is glacier melt;

ρ is all precipitation falling in a basin between June 22 and August 14; and

δ is snowmelt

the general equation can be rewritten

$$\gamma + \rho + \delta = QT \tag{4.14}$$

The values for every term are available as a result of the calculations described above and are given in table 4.7.

Despite the crudeness of these estimates, some interesting points emerge. Although the "Schei River" basin is only 35.5% glacierized, melt of glacier ice accounted for 55% of total recorded discharge in 1974. Precipitation between June 22 and August 18 is estimated to have contributed

TABLE 4.7

CALCULATED VALUES FOR THE
TERMS OF THE WATER BALANCE EQUATION $\gamma + \rho + \sigma = QT$

a) in 10^3 m^3

River	TQ	γ	ρ	σ
Schei (s)	19,867	10,884	1,915	7068
Upper Schei (u)	12,094	7,238	1,035	3821
Endrick (e)	2,948	-	627	2319
Lendal (l)	4,568	3,643	197	728
[Lower Schei]	256	-	55	201

b) in 10^3 m^3 km^{-2}

River	TQ	γ	ρ	σ
Schei (s)	218	119.5	21	77.5
Upper Schei (u)	246	147.5	21	77.5
Endrick (e)	99	-	21	77.5
Lendal (l)	486	387.5	21	77.5
[Lower Schei]	99	-	21	77.5

only 10%, and the remaining 35% is largely attributable to snowmelt. In his studies of the Ermine River, which drains the White Glacier on Axel Heiberg Island, Adams (1966) also found that glacier melt provided the dominant input to streamflow (table 4.8). The contribution of glacier melt to the discharge of the "Ermine River" was however much lower in 1961, a relatively cool, cloudy season, than in the summer of 1960. A similar contrast exists for the "Schei River" between the summers of 1973 and 1974; runoff on the latter year was greater, despite the heavy rainfall of 1973, as the relatively warm, sunny conditions of 1974 promoted faster ablation of the "Schei Glacier".

The size of the glacier-melt contribution to discharge has important geomorphic repercussions. It is evident from the figures presented in tables 4.7 and 4.8 that, for relatively warm seasons at least, the ice-melt input to streamflow considerably exceeds the snowmelt input. There is however no evidence that ice-melt discharge per unit area exceeds snowmelt discharge per unit area. The geomorphic importance of the ice-melt contribution lies in the fact that high glacial discharge may persist throughout the season, whereas high nival discharges terminate at the end of the snowmelt freshet. As a result, the geomorphic effectiveness of glacial streams probably exceeds that of their nival counterparts, as the erosion and transportation which accompany high discharges will operate over a longer period.

In the above section, discharge records from glacierized and unglacierized portions of the "Schei River" basin and the summer precipitation record were employed to measure the size of different streamflow inputs. From these data it was possible to isolate the input of glacier

TABLE 4.8

Comparison of streamflow inputs to the "Schei River" and its tributaries with those calculated for the Ermine River, White Glacier, Axel Heiberg Island.

River	Year	Basin Area (km^2)	%Area Glacierized	QT ($10^6 m^3$)	%γ	%ρ	%δ
Ermine [1]	1960	54	75	31.85	81	9	10
Ermine [1]	1961	54	75	13.4	63	15	33
"Schei" [2]	1973	91.2	35.5	16.9	-	-	-
"Schei" [3]	1974	91.2	35.5	19.8	55	10	35
"Upper Schei"	1974	49.3	54	12.1	60	8	32
"Lendal"	1974	9.4	63	4.6	80	4	16
"Endrick"	1974	29.9	0	2.9	0	21	79

Notes:

1 from Adams (1966)

2 QT obtained by J.G. Cogley (pers. comm). Probably a high estimate

3 QT obtained by summing the contributions of three tributaries.

melt from that of snowmelt on the glacier surface, thus overcoming one of the main problems inherent in the hydrologic assessment of ablation rate (Ostrem, Bridge and Rannie, 1967). This technique may be of value in glacier mass balance studies.

4.9 SUMMARY AND CONCLUSIONS

In this chapter the discharge of the glacionival "Schei River" has been described and analysed by comparing the hydrologic behaviour of its three main tributaries. Several generalisations emerge from this analysis, mainly concerning differences between the behaviour of the glacierized and non-glacierized parts of the basin. The most important of these generalisations are as follows:

1. Nival streams differ in behaviour from glacial streams after the end of the snowmelt freshet, in that reserves of water in unglacierized catchments become depleted whereas the supply of glacial meltwater is effectively limitless. In consequence, high discharge of nival streams after the snowmelt flood may only be expected after rainfall. High discharge of glacial streams may however occur on warm, sunny days at any time during the runoff season.
2. The time of the diurnal peak discharge of glacial streams advances progressively throughout the runoff season. Time of diurnal peak discharge of nival streams is progressively retarded however, as sources of meltwater retreat to the higher parts of ice-free basins.
3. The magnitude of river response to rainfall on unglacierized terrain is apparently dependent on antecedent moisture conditions in the active layer.

4. Diurnal fluctuations in discharge can be related to fluctuations in temperature and net radiation. No simple relationship was found between the values of these three variables, however.

5. The lag of glacial discharge in response to rainfall is of similar magnitude to the lag of peak diurnal discharge behind maximum daily net radiation.

6. The contribution to runoff of the glacierized part of the "Schei Basin" was greater in 1974 than that of the ice-free area. Glacier melt comprised roughly 55% of total discharge, snowmelt 35%, precipitation 10%. The high level of the glacier-melt contribution reflects the relatively warm and sunny conditions of the 1974 runoff season.

7. The ponding of water behind collapsed ice at the margin of a glacier may represent a considerable hydrologic hazard, as a sudden flood of unpredictable timing and magnitude may result from the breaching of the ice dam.

Although many high arctic streams drain both glacierized and ice-free terrain, the arctic glacionival regime is poorer documented than either of its component types. The present study is a first attempt to assess the influence of each component on the pattern of glacionival discharge. It must be emphasised that the potential for more rigorous analysis of glacionival systems is great.

CHAPTER 5

SEDIMENT TRANSPORT AND SUPPLY

5.1 Introduction

Present knowledge of the concentration and load of sediment transported by high arctic rivers is limited to the measurements reported in a small number of recent studies (for example : Church, 1972; McCann, Howarth and Cogley, 1972; McCann and Cogley, 1974). In the present chapter measurements of the concentration of suspended and dissolved material transported by the "Schei River" in 1974 are reported. Variations in the level of concentration are discussed, with particular reference to the effect of changing discharge and to the sediment contribution of different parts of the basin. The results obtained are then compared with those of previous studies, and conclusions drawn as to geomorphic implications.

Several geomorphologists (for example : Corbel, 1959; Smith, 1969, 1972; Douglas, 1973) have manipulated measurements of sediment discharge to derive a quantitative index of denudation rate. This technique assumes that all of the material eroded from a basin is subject to eventual removal by river, hence measurements of river load over a given time period can be used to express denudation per unit area during that time period. The basic equation is

$$D_t = \frac{L_t}{\rho A} \qquad (5.1)$$

where D_t is mean denudation occurring in time t, L_t is total load (mass) transported from a basin in time t, ρ is density of transported material (normally taken as 2.65 g cm^{-3}) and A is basin area. Extrapolation of D_t to estimate long-term denudation (for example, in millimetres per thousand years) is meaningless, however, as rates of denudation are apt to vary widely from year to year, and over a long time will reflect changes in climate and base level. The most promising use of this and similar indices of denudation is that of Douglas (1973), who employed multivariate analysis to identify the main influences on denudation rate in nearby catchments over the same time period.

The principal difficulty in the use of equation *(5.1)* lies in arriving at a value for the L_t term, as the three components of this term (dissolved load, suspended load and bedload) must be evaluated individually. Accurate assessment of bedload is difficult (Hubbell, 1964), and most authors either ignore this fraction of the load or resort to the uncertainties of a standard bedload formula. The evaluation of dissolved load and suspended load requires a continuous discharge record and the establishment of separate rating curves for concentrations of suspended and dissolved sediment against discharge (Campbell and Bauder, 1940). This too may be problematic, as factors other than discharge may affect sediment concentration (Guy, 1964). In particular, the availability of fine sediment is of crucial importance in determining the level of suspended inorganics carried by small to medium sized rivers.

In view of these difficulties, no attempt is made in this study to arrive at a comprehensive value for denudation rate in the "Schei Basin". Instead, attention is focussed on the factors affecting the removal of sediments in suspension and solution, and the comparison of measured concentrations with those obtained in other arctic environment.

5.2 Measurements Techniques

In order to assess the concentration and load of suspended and dissolved materials in the "Schei River", water samples were obtained daily at the main gauging station, normally around the time of peak diurnal discharge (1600 - 1800 h), using a depth-integrating DH 48 sediment sampler. Concentration of dissolved materials was assessed for all samples in terms of specific conductivity, which gives a measure of free ions in solution. Conductivity was measured with a Barnstead PM-70CB conductivity bridge; all measurements were made at a sample temperature of precisely 25°C to eliminate the influence of temperature. In addition, nine samples taken at the main "Schei River" gauge were titrated to obtain values of calcium and total (Ca^{2+} + Mg^{2+}) hardness, using a B.D.H. hardness test in the manner described by Schwartzenbach (1957). As well as throwing light on the identity of the principal ions in solution, these measurements allowed partial calibration of the conductivity values.

To measure the concentration of sediment in suspension, all water samples taken at the "Schei River" gauging station were filtered at base camp using a modified version of the rapid filtration apparatus devised by G. Østrem (Østrem and Stanley, 1969; Østrem, Ziegler and Ekman, 1970). Sample volume was measured to the nearest millilitre before filtration through cellulose acetate filters, 47 mm in diameter with mean pore size 45 μm. After filtration, these filters were stored in numbered plastic bags, and in the subsequent laboratory analysis the filters were dissolved in acetone and the bags washed out in distilled water. The acetone and distilled water were then evaporated, and the residue on the filters was placed in a muffle furnace for 30 minutes at 350° - 400°C. The residue of the filters and the bags was then weighed on a laboratory balance sensitive to 0.1 mg. Tests showed that this procedure reduced filter weight to an amount negligible in comparison with the weight of the sediment sample.

The daily sampling programme at the "Schei River" gauge was supplemented by occasional sampling at the tributary gauging points, designed to enable the comparison of the sediment contribution of each tributary, and at the glacier front, where meltwater erosion was extremely active. These supplementary samples were analysed using the filtration, conductivity measurement and hardness titration techniques outlined above. In addition, occasional measurements of pH were made at base camp using a Metrohm E-280A pH meter with manual temperature compensation.

5.3 Solute Concentration and Load

5.3.1 "Schei River" samples

Carbonate rocks outcrop fairly widely in the unglacierized portion of the "Schei River" basin. The principal calcareous series are those of the Lower Palaeozoic Read Bay and Allen Bay formations, and the Devonian Vendom Fiord formation contains strata of calcareous sandstone (figure 2.2). Moreover, the tills and outwash gravels that mantle much of the basin contain abundant carbonate erratic material, such that groundwater and surface runoff are frequently in contact with a relatively soluble calcareous regolith. Studies of solute concentrations in small streams draining high arctic limestone and dolomite terrain indicate that solute load accounts for a substantial portion of total sediment transport (McCann and Cogley, 1971; Cogley, 1972; M-k Woo, personal communication). For larger arctic streams, concentration apparently diminishes (Smith, 1969, 1972; McCann, Howarth and Cogley, 1972) but dissolved solids still constitute a significant portion of total river load. It might be anticipated, therefore, that the waters of the "Schei River" will exhibit low but not insignificant values of hardness and conductivity.

The conductivity readings for 66 water samples taken at the "Schei River" gauge in 1974 ranged from 36 to 206 μmho cm^{-1} (figure 5.1). The range obtained for the Mecham River, which drains a non-glacierized basin of similar size on the carbonate rocks of S.E. Cornwallis Island, was 88 to 210 μmho cm^{-1} for 1971 (McCann et al., 1975). For the "Schei River", nine samples titrated to give values of calcium and total (Ca + Mg) hardness demonstrated a fairly close relationship between total hardness (H_t) and conductivity (K), such that

$$H_t = 0.468K - 1.145 \qquad (5.2)$$

$$(r = 0.878)$$

where K is expressed in μmhos cm^{-1} and H_t in p.p.m. The small constant term in equation (5.2) indicates that conductivity will be virtually zero when there is no Ca^{2+} or Mg^{2+} in solution. Hence it would appear that measurements of total hardness account for almost all of the dissolved solids in the "Schei River", other ions being of negligible importance. Equation 5.2 allows conductivity measurements to be rated in terms of total dissolved solids. Calculated from this equation, concentration of dissolved materials in the "Schei River" in 1974 ranged from 15 to 97 ppm, calcite

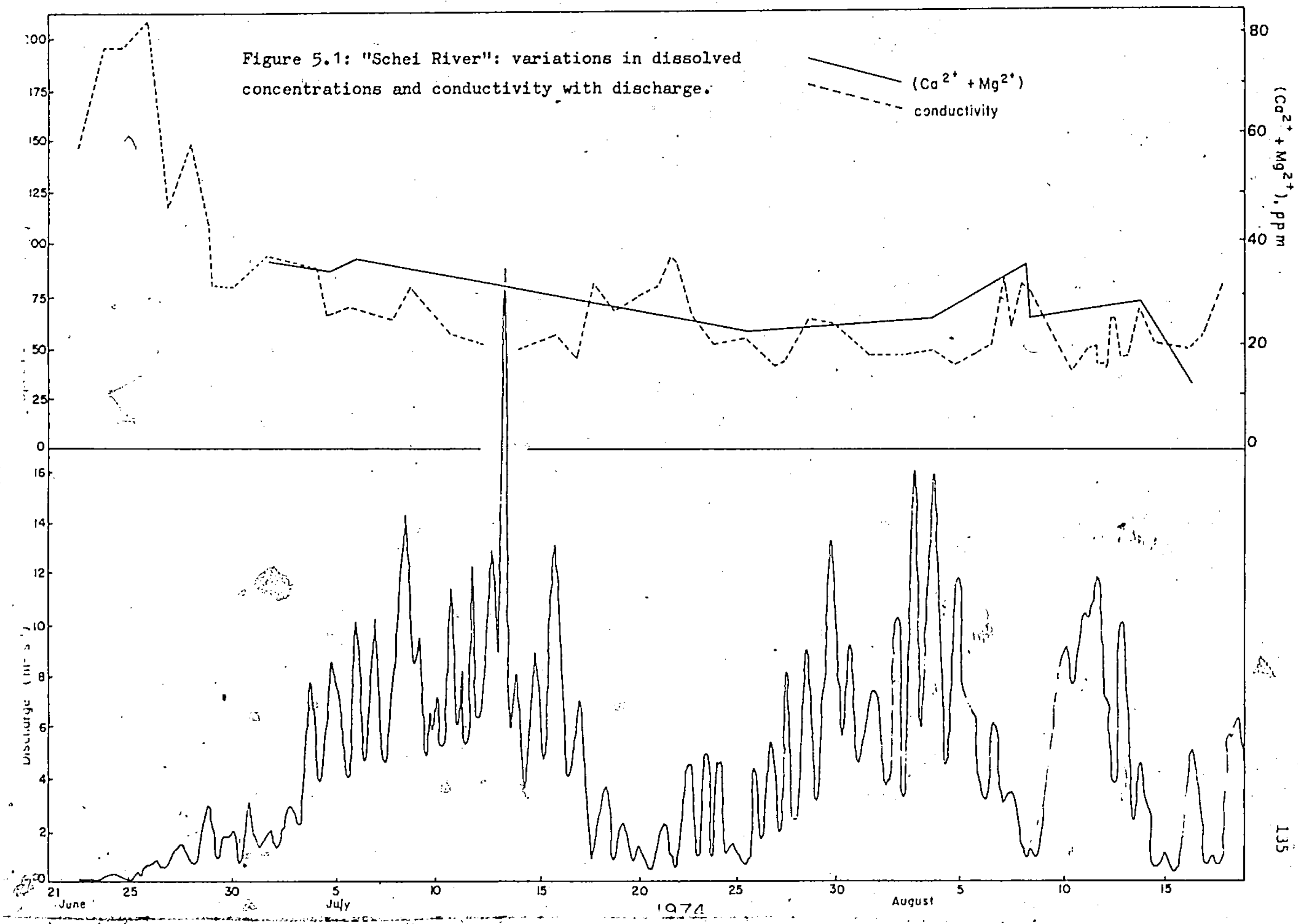


Figure 5.1: "Schei River": variations in dissolved concentrations and conductivity with discharge.

equivalent, as against 65–135 ppm for the Mecham in 1971.

Conductivity exhibits an inverse nonlinear relationship with water discharge (figure 5.1):

$$K = 74.508\, Q_s^{-0.206} \qquad (5.3)$$

$$(r = -0.761)$$

Combining equations (5.2) and (5.3) gives the relationship between total hardness (≈ total dissolved concentration) and water discharge as

$$H_t = 34.870\, Q_s^{-0.206} - 1.145 \qquad (5.4)$$

implying that discharge increases more rapidly than concentration decreases. This well-documented phenomenon has been attributed (for example, by Cogley, 1972) to the operation of a "dilution effect" : as discharge increases, a smaller proportion of the water comes into contact with the channel perimeter and opportunities for solutional reaction are reduced. This explanation may be inadequate, however, for streams as turbulent as the "Schei River". In the present instance two alternative explanations of the inverse hardness – discharge relationship may be invoked, concerning relative inputs to streamflow and increase in turbulence with discharge. At high discharges, glacial and nival meltwaters inevitably constitute by far the most important input to streamflow. For reasons discussed in sections 5.3.2 and 5.3.3 such meltwater is unlikely to contain significant concentrations of dissolved solids. At lower discharges, however, the proportion of total input contributed by groundwater will be greater. As a result of their relatively long travel time through the normally calcareous active layer, groundwaters are probably saturated with respect to

$CaCo_3$ and $MgCO_3$ and therefore contain much higher levels of dissolved material than do surface meltwaters. Hence as discharge increases, increasing amounts of meltwater tend to "dilute" the overall solute concentration. Moreover, at high discharges the turbulence of the river increases markedly, particularly in the steep "Schei Gorge" section. Increasing turbulence will result in diminution of the level of dissolved CO_2, a lowering of the saturation level and the precipitation of some dissolved carbonates as particles in suspension.

To calculate the total solute load passing the "Schei River" gauging station during the summer of 1974, the mean concentration of dissolved solids $(\overline{H_t})$ was calculated using the expression

$$\overline{H}_t = \sum_{i=1}^{n} (H_{t_i} f_i) \tag{5.5}$$

where f_i is the probability of the i *th* class of discharge (measured at intervals of 1.0 $m^3 s^{-1}$) and H_{t_i} is obtained through substituting for Q_s according to equation (5.4). In this way a value of 28.814 ppm was obtained for $\overline{H}_t$. Multiplied by total seasonal discharge (19.867 x 10^6 m^3), this gives a total volume of 572.4 m^3 removed in solution from the "Schei Basin" during the study period; assuming a density of 2.65 g cm^{-3}, this is equivalent to 1517 metric tons. Total load per unit area is therefore 16.6 t km^{-2}; total load per unit ice-free area is 25.8 t km^{-2}.

5.3.2 Tributary Samples

Throughout the field season occasional water samples were taken from the "Schei River" tributaries to enable the hardness of waters from different parts of the basin to be compared. In total, 13 hardness and conductivity values were obtained for the "Upper Schei River", 14 for "Lendal Creek" and 12 for "Endrick Creek". The total hardness - conductivity relationships exhibited by these samples are similar in form to that obtained for the "Schei River" samples (equation 5.2):

$$H_{t_u} = 0.508\ K_u - 1.518 \qquad (r = 0.983) \tag{5.6}$$

$$H_{t_l} = 0.583\ K_l - 1.996 \qquad (r = 0.926) \tag{5.7}$$

$$H_{t_e} = 0.482\ K_e + 6.211 \qquad (r = 0.936) \tag{5.8}$$

The coefficients of K are similar for all these equations and equation (5.2), and the constant term is low for all but equation (5.8), indicating that ions other than Ca^{2+} and Mg^{2+} may be significant for the "Endrick" samples, but not for the "Upper Schei" or "Lendal" samples.

Samples taken from the "Upper Schei River" and "Lendal Creek" after the end of June exhibited relatively little change in total hardness (figure 5.2) and no significant relationship with discharge (figure 5.3). The "Endrick Creek" measurements however varied inversely with discharge according to the regression

$$K_c = 85.313 - 23.103\ Q_e \qquad (r = -0.909) \tag{5.9}$$

From figure 5.2 it is evident that concentrations of dissolved solids are much greater for "Endrick Creek" than for either of its glacier-nourished counterparts. This difference may merely reflect the consistently low

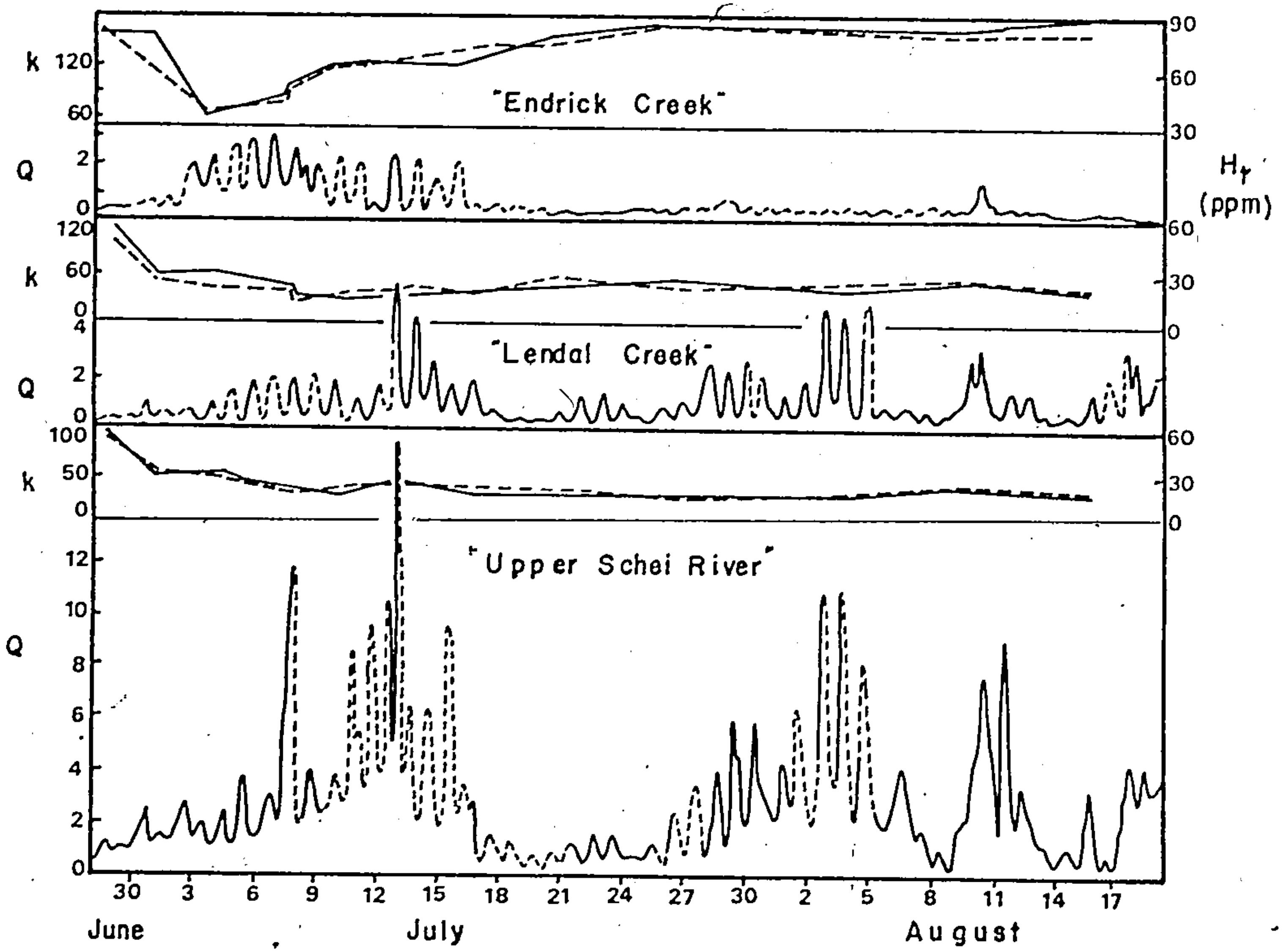


Figure 5.2: "Schei River" tributaries: variations in hardness and conductivity.

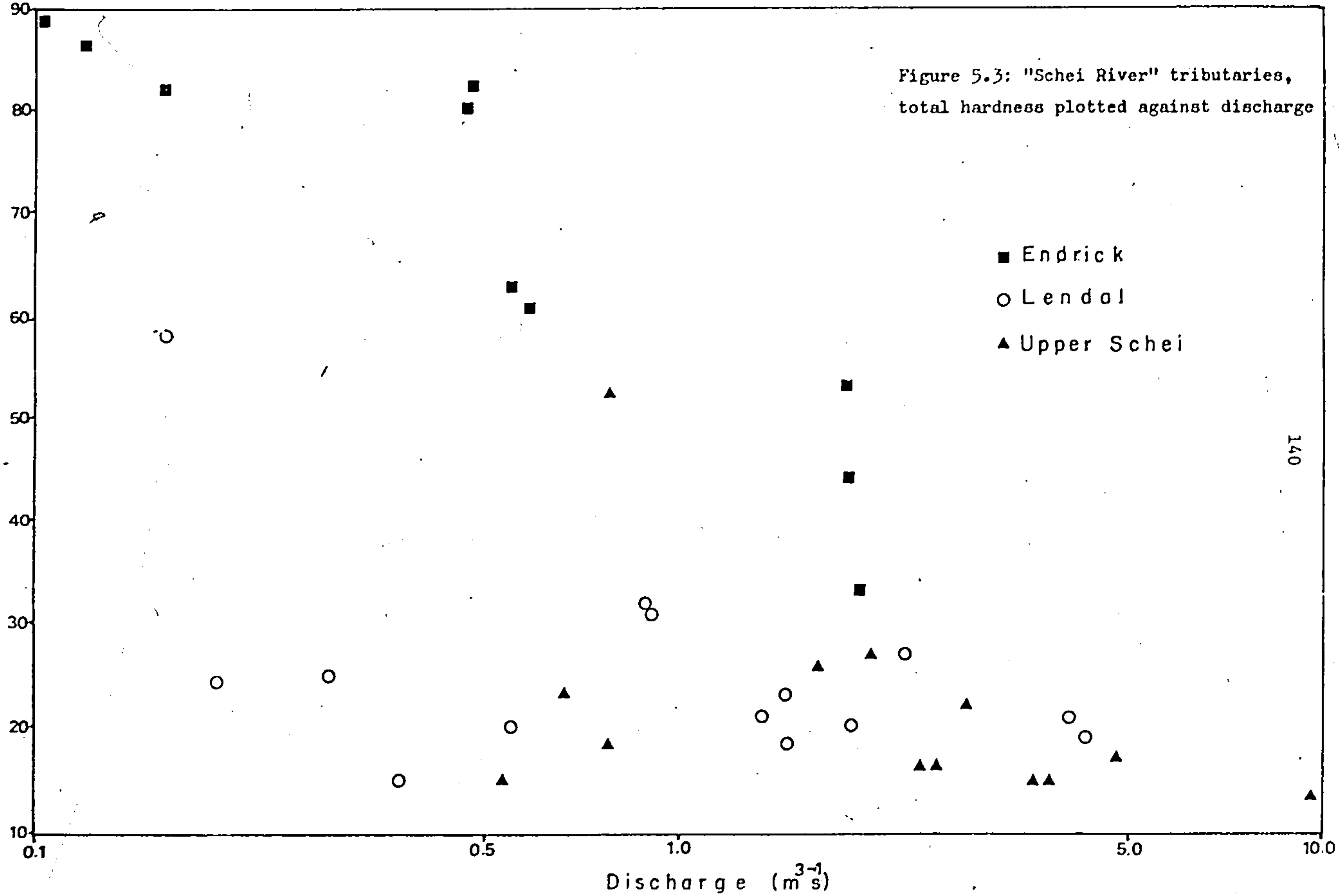


Figure 5.3: "Schei River" tributaries, total hardness plotted against discharge

discharge of "Endrick Creek" after the snowmelt flood, but some features of figure 5.3 suggest a more complex picture. The highest values for "Lendal Creek" (H_t = 58 ppm) and the "Upper Schei River" (H_t = 52 ppm) were both obtained on June 28, when the main input to streamflow was probably snowmelt rather than glacier melt. The studies of Ek (1964, 1966) demonstrated that the CO_2 content of glacier ice is depleted through compaction and densification, so that glacial meltwaters are much less aggressive than those resulting from the melting of new snow, which is apparently not subject to CO_2 depletion (Clement and Vandour, 1968). It is thus possible to account for the low hardness values of the "Upper Schei River" and "Lendal Creek" obtained after June 28 in terms of the increased dominance of glacial meltwater as a streamflow input after that date. Alternatively, it is conceivable that the solute concentration of glacial meltwaters sampled at the "Lendal" and "Upper Schei" gauging sites had not reached saturation before sampling took place.

Figure 5.4 shows the variation in dissolved load per unit basin area for the "Schei River" and its tributaries, based on the limited number of samples taken at each gauge. The high values measured for "Lendal Creek" are striking. Some sources of the dissolved load carried by "Lendal Creek" were investigated through sampling waters near the glacier front, reported in the following section.

5.3.3 Ice-front samples

To further define the role of glacial meltwater in transporting salts in solution, occasional samples were taken at two ice-front locations in the "Lendal" basin. The area around the submarginally-draining pond

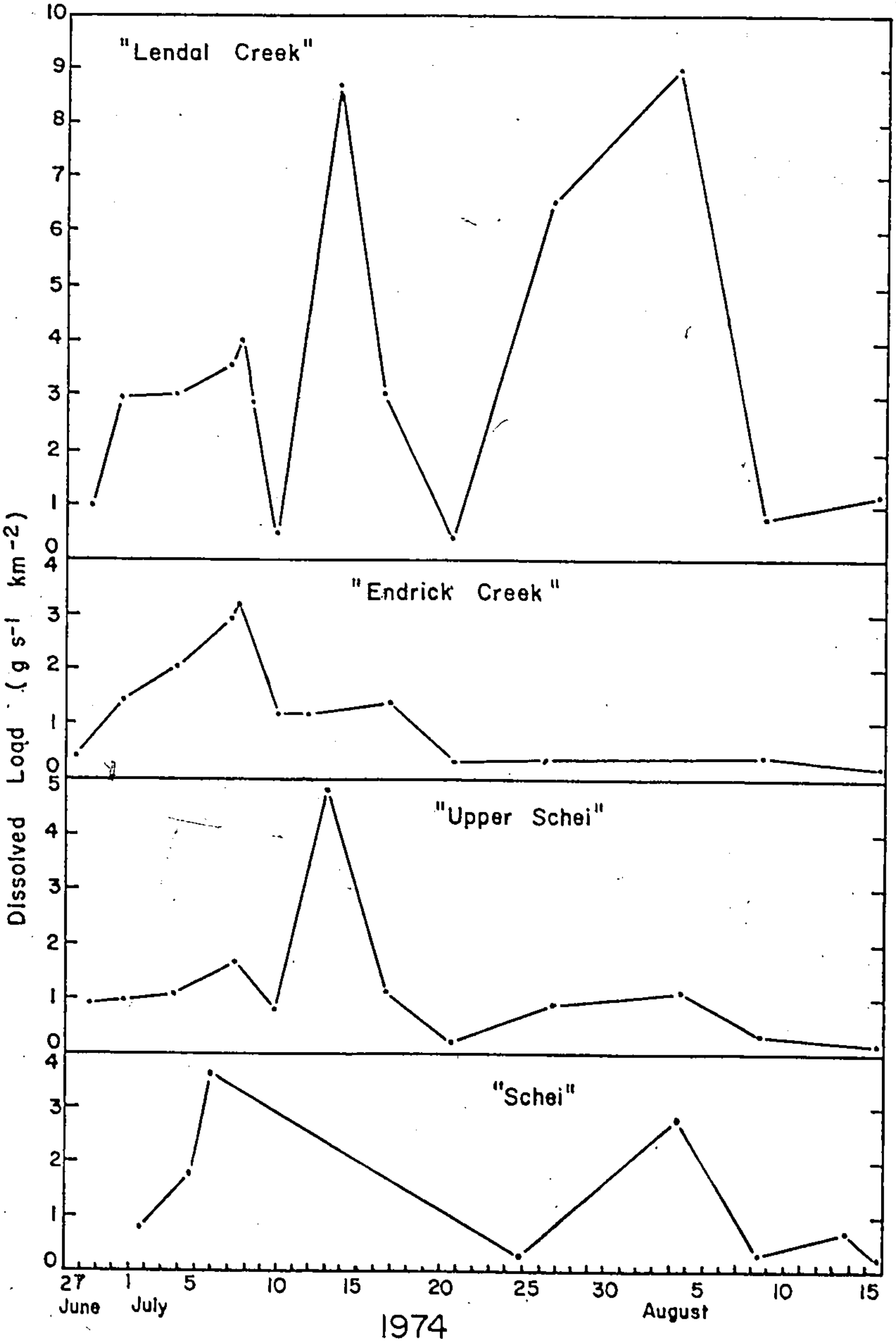


Figure 5.4: Dissolved load (sum of Ca^{2+} and Mg^{2+} expressed as $CaCO_3$) of the "Schei River" and its tributaries.

described in section 4.8 was sampled on six occasions (figure 5.5), and a small proglacial tributary of the "Lendal" christened "Thistle Creek" was sampled ten times between June 26 and August 3 (figure 5.6; table 5.1).

The results of sampling at the proglacial pond were somewhat inconclusive. All of the values obtained were low *(H_t < 25 ppm;* figure 5.5) but significantly those obtained for glacial meltwater were lowest of all *(5, 7 and 13 ppm)*. A single value obtained for glacier ice *(H_t = 12 ppm)* was also very low. These figures suggest that solute concentrations in meltwater rise within a few metres of the glacier front.

The "Thistle Creek" samples, taken 20 m from the ice-front, all displayed low *(12 - 25 ppm)* total hardness values except for the June 28 sample *(56 ppm)*, when flow was probably nourished primarily by snowmelt rather than glacier melt. Some insight was gained into the factors influencing the hardness of "Thistle Creek" by sampling individual water inputs (figure 5.6). Samples taken from supraglacial meltwaters and gravel rills had low *(10 ppm)* concentrations; those flowing over fine deposits (till and mudslump rills) had much higher concentrations *(34 and 33 ppm* respectively). The concentrations of waters flowing in "Thistle Creek" and along the glacier margin were intermediate *(15 and 20 ppm* respectively), although the hardness of "Thistle Creek" increased downstream to *27 ppm*. These results support the inference made for the ice-marginal pond samples, that solution takes place rapidly as meltwater leaves the glacier, particularly, it seems, where flow is over fine sediments.

The limited pH measurements made at the ice-front sampling sites (figure 5.5; table 5.1) range from pH 7.5 to pH 9.2. This range resembles

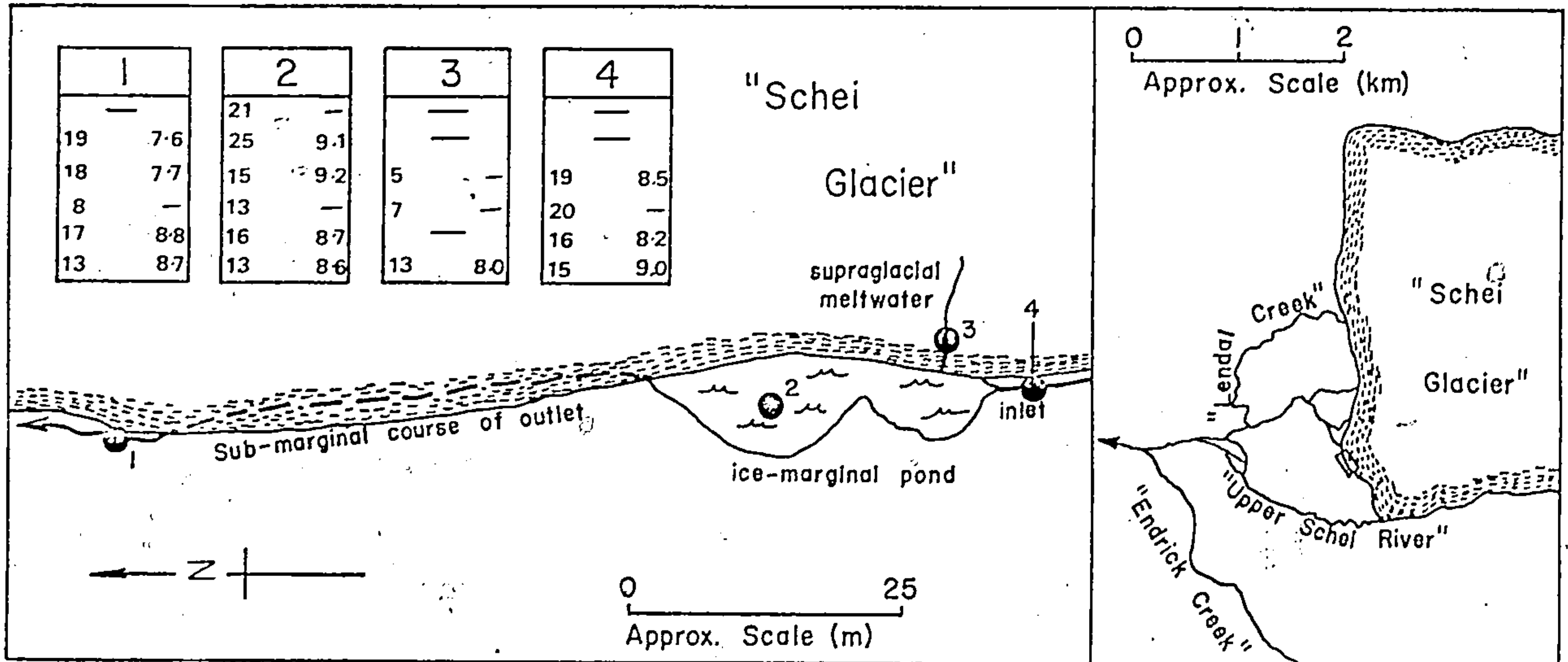


Figure 5.5: Sketch of ice-marginal pond with total (Ca+Mg) hardness and pH values measured on six occasions in the summer of 1974.

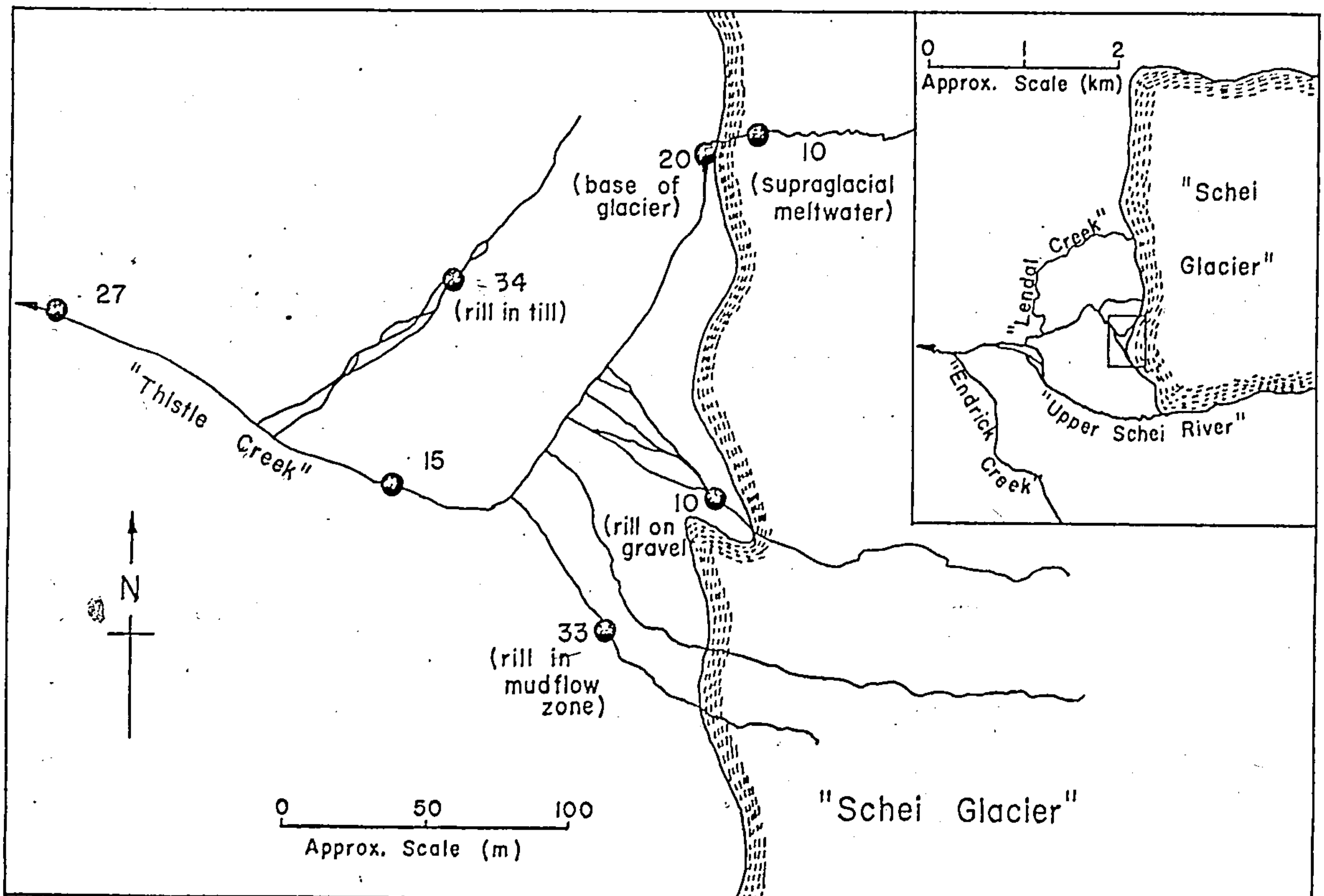


Figure 5.6: Total hardness values in p.p.m. for individual water inputs to "Thistle Creek" on August 3, 1974.

TABLE 5.1;

CONCENTRATION OF SUSPENDED AND DISSOLVED SEDIMENT IN SAMPLES TAKEN AT THE ICE FRONT AT "THISTLE CREEK"

Date	Sample	Suspended Conc. (ppm)	Dissolved Calcium (ppm)	Total Dissolved (ppm)	Conduct- ivity (μ mho cm^{-1})	pH	Stage (cm)
260674	Thistle Creek	911					1.4
280674	Thistle Creek	1634	30	56	106		0.2
300674	Supraglacial stream	4	4	8	22		
	Snowbank at glacier	6	8	14	32		
	Thistle Creek	121	12	20	31		2.6
030774	Thistle Creek	418	17	25	35		3.8
050774	Thistle Creek		11	15	31		6.5
090774	Thistle Creek	5	11	15	31		6.0
120774	Thistle Creek	95	8	12	40		13.0
160774	Thistle Creek	48	11	15	36		11.0
260774	L.H. tributary above mudflow	33	4	10	19	7.8	
	L.H. tributary below mudflow	3971	16	20	47	7.5	
	R.H. tributary	104	7	13	35	8.1	
	Thistle Creek	966	10	19	40		10.0
030874	Supraglacial stream	19	4	10	14		
	Base of glacier	53	16	20	50	8.9	
	Rills in gravel	22	4	10	18	8.1	
	Mudflow rills	8529	26	33	69		
	Rills in till	11	17	34	70		
	Rills in till after mudslump	2275					
	Thistle Creek	537	8	15	33		
	Thistle Creek above confluence	611	15	27	67		

that of pH 7.65 to pH 8.9 measured by Ek (1964, 1966) at glacier snouts in the Savoy Alps and the Italian Dolomites. If, as Ek suggests, these high pH values are attributable to the explusion of CO_2 in glacier ice upon densification, then the low hardness values obtained at the ice front may be explained in terms of a reduction in saturation level consequent on low levels of dissolved CO_2. None of the hardness values obtained at the glacier snout approach the calcite saturation level of 95 ppm under conditions of atmospheric p CO_2 at a temperature of 0°C.

5.4 Suspended Sediment Transport and Supply

5.4.1 "Schei River" samples

Measured concentrations of suspended sediment in the "Schei River" in 1974 ranged from 5 mgl^{-1} to 2,205 mgl^{-1} and measured sediment discharge (load) varied from 2 gs^{-1} to 23,956 gs^{-1}. Nine of the 67 water samples exceeded 1,000 mgl^{-1} in concentration, and loads in excess of 10,000 gs^{-1} were measured on eight occasions. It must be stressed, however, that as sampling generally took place around the time of peak diurnal discharge, these values represent maxima.

As might be expected on the basis of previous studies (for example, Arnborg, Walker and Peippo, 1967; Østrem, Bridge and Rannie, 1967; Church 1972) concentration of suspended inorganics displayed considerable covariance with discharge (figure 5.7). However, intensive sampling on August 11 and 12 demonstrated that order of magnitude changes in suspended sediment concentration may occur in short time intervals (figure 5.8). Obviously, then, the concentration of any sample may be unrepresentative of the discharge at which it was taken and of the mean concentration for the day on which it was taken, particularly as availability of suitable sediment

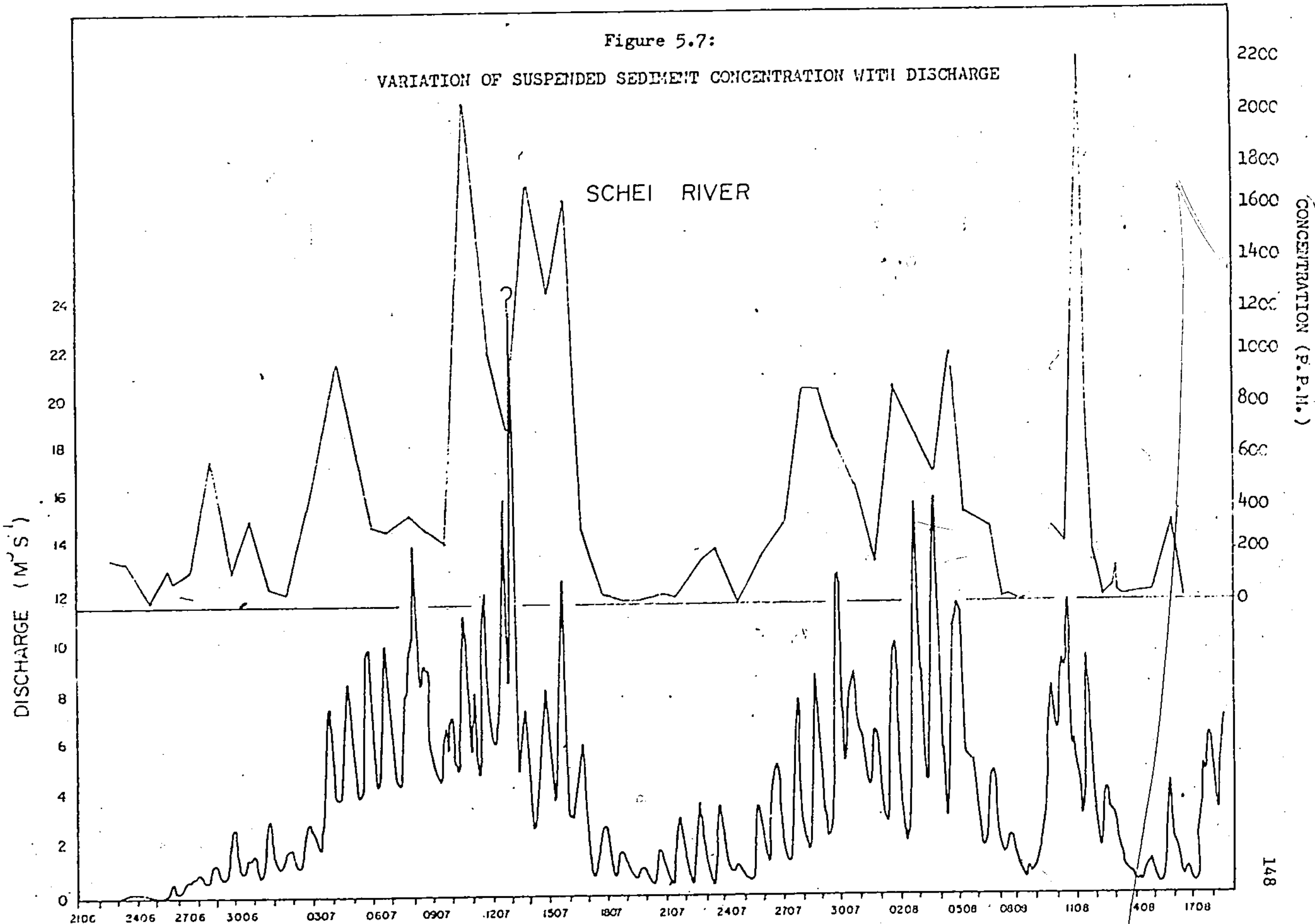


Figure 5.7:

VARIATION OF SUSPENDED SEDIMENT CONCENTRATION WITH DISCHARGE

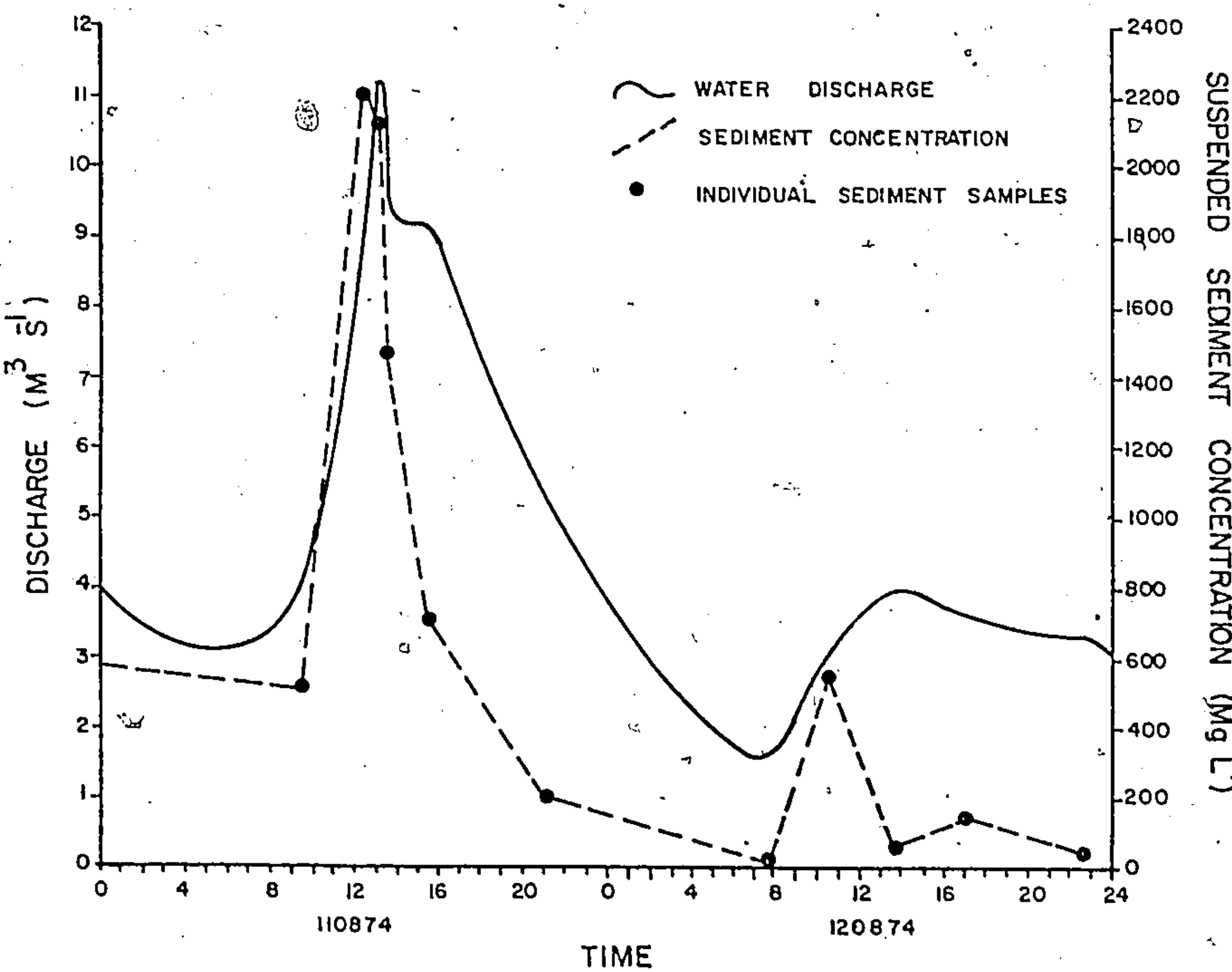


Figure 5.8: "Schei River discharge and suspended sediment concentration, August 11-12, 1974.

exercises, independent of discharge, a major control on concentration. An identical problem has been described by Østrem, Bridge and Rannie (1967), who made measurements under similar circumstances in the glacierized Decade River basin, on Baffin Island.

The relationship between suspended sediment concentration *(S)* and discharge (figure 5.9) is expressed by

$$S_s = 45.710 \, Q_s^{1.22} \qquad (5.10)$$

$$(r = 0.774)$$

confident at the 0.01% level. The wide scatter of points on figure 5.8 is however a phenomenon of some interest. Measurement error is only likely to significantly effect samples with low concentrations, and can be discounted as a major effect. Dumping of silt-laden snow into the river at the very beginning of the 1974 runoff season, giving relatively high concentrations during periods of low discharge, is important for only four samples. More marked is the influence of a hysteresis effect, the lag of peak water discharge behind peak sediment concentration (figure 5.8). In consequence of this, a sample taken on falling stage will exhibit lower concentration at a given discharge than a sample taken on rising stage (figure 5.9). Surprisingly, comparison of the correlation coefficients of regressions of rising and falling stage reveals less variance in the falling stage distribution than in the rising stage distribution, with a falling stage coefficient of 0.875 and a rising stage coefficient of 0.704. For the 1973 "Schei River" suspended sediment data (McCann et al., 1975) the equivalent values were 0.669 and 0.436. The reason for the better "organisation" of the falling stage samples is not clear.

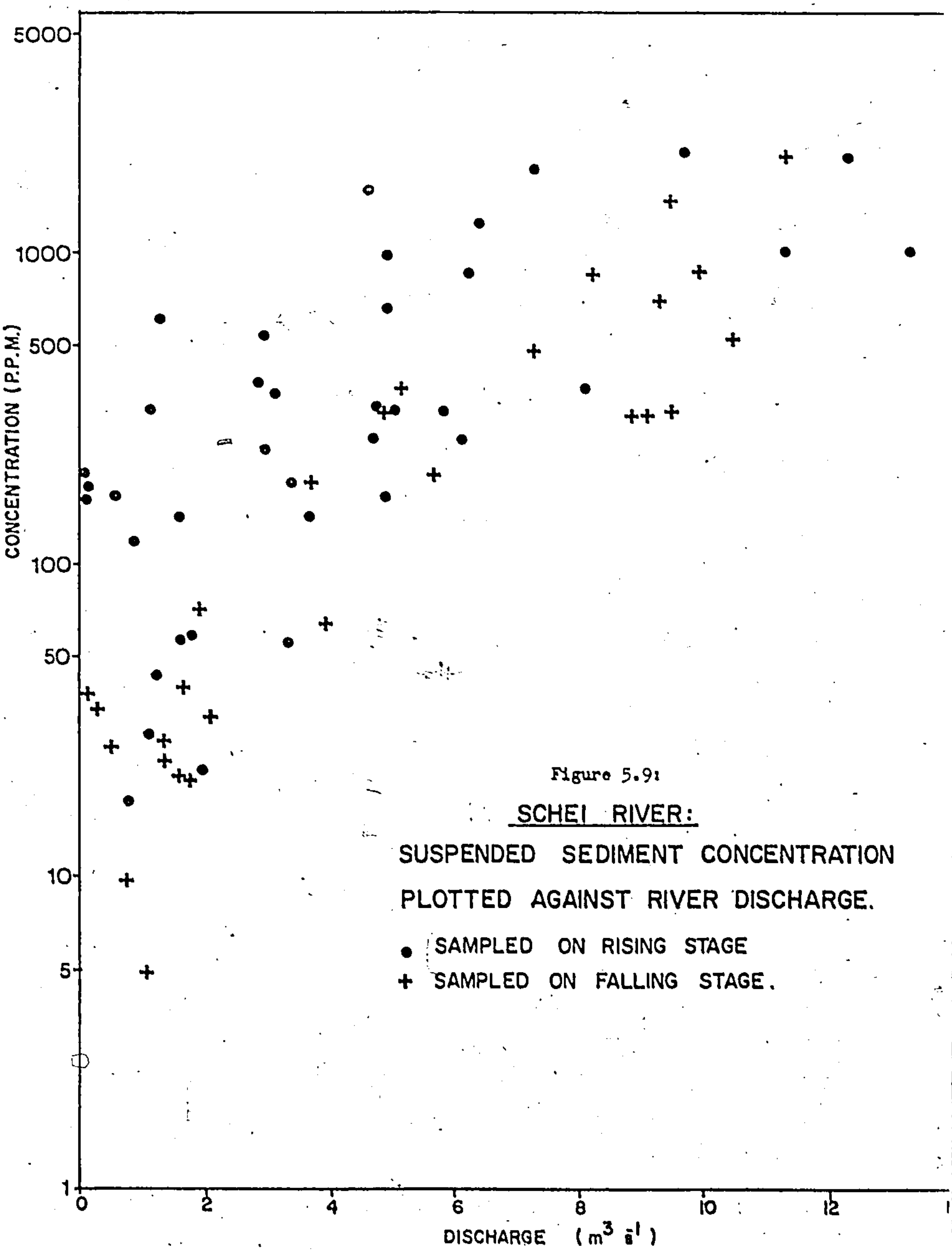


Figure 5.9: SCHEI RIVER: SUSPENDED SEDIMENT CONCENTRATION PLOTTED AGAINST RIVER DISCHARGE.

Probably the most important factor behind the scatter of points in figure 5.8 is variation in the supply of fine sediment. In common with most streams, the "Schei River" normally carries a suspended concentration that is less than its potential capacity. How much less is determined by the availability of suitable material. Some important sources of sediment are described in section 5.4.2.

The exponent of the independent variable Q_s in equation (5.10) is sufficiently close to unity to support Church's (1972) findings of a near-linear relationship

$$S \approx K\, Q \tag{5.11}$$

where K is a constant, and implying that the relationship of suspended load (L) to water discharge is

$$L \approx K\, Q^2 \tag{5.12}$$

This indicates the importance of extreme runoff events in transporting suspended sediment from the basin.

Sediment load is the product of concentration and discharge

$$L = S\, Q \tag{5.13}$$

When S is given in mgl^{-1} and Q is in $m^3\ s^{-1}$, then load will be in gs^{-1}. As L is a function of Q, the relationship between the two is liable to show a spuriously high correlation, essentially the result of regressing Q against itself. For this reason no such analysis was carried out.

The seasonal suspended load for the period June 22 - August 18, 1974 was calculated in a similar manner to total dissolved load (equation 5.5) using flow-frequency data and the suspended sediment rating curve (equation 5.10). In this way a figure of 10,251 metric tonnes was obtained for total load during the study period, corresponding to 112.4 t km^{-2} for the entire basin or 174 t km^{-2} for the ice-free portion of the basin.

5.4.2 Sources of suspended sediment

Measurement of suspended sediment concentration at each of the three tributary gauging sites throws some light on the nature of major sources. Figure 5.10 shows the variations in suspended sediment concentrations obtained on the basis of limited (13 - 15 samples) sampling on each tributary. Some trends are discernible from this data:

1. Sediment concentrations were high relative to discharge at the onset of the flow season. This probably results from the "flushing out" of sediment accumulated on snowbanks by progressively higher meltwater runoff.

2. There is a general correspondence between discharge and suspended sediment concentration in the glacial tributaries after June 5.

3. Suspended sediment concentration in "Endrick Creek" was depleted after peak seasonal discharge was reached.

The pattern of seasonal change in concentration in "Endrick Creek" is largely explained by the occurrence of extensive bank erosion up to the period of peak flow (figure 6.4) and the abrupt cessation of such erosion thereafter (figure 6.5). Little fine material was available from the rock and boulder bed of this stream, so that the waters of "Endrick Creek" were observed to be consistently clear after July 6. Amongst the sources of material supplying the "Upper Schei River", bank collapse

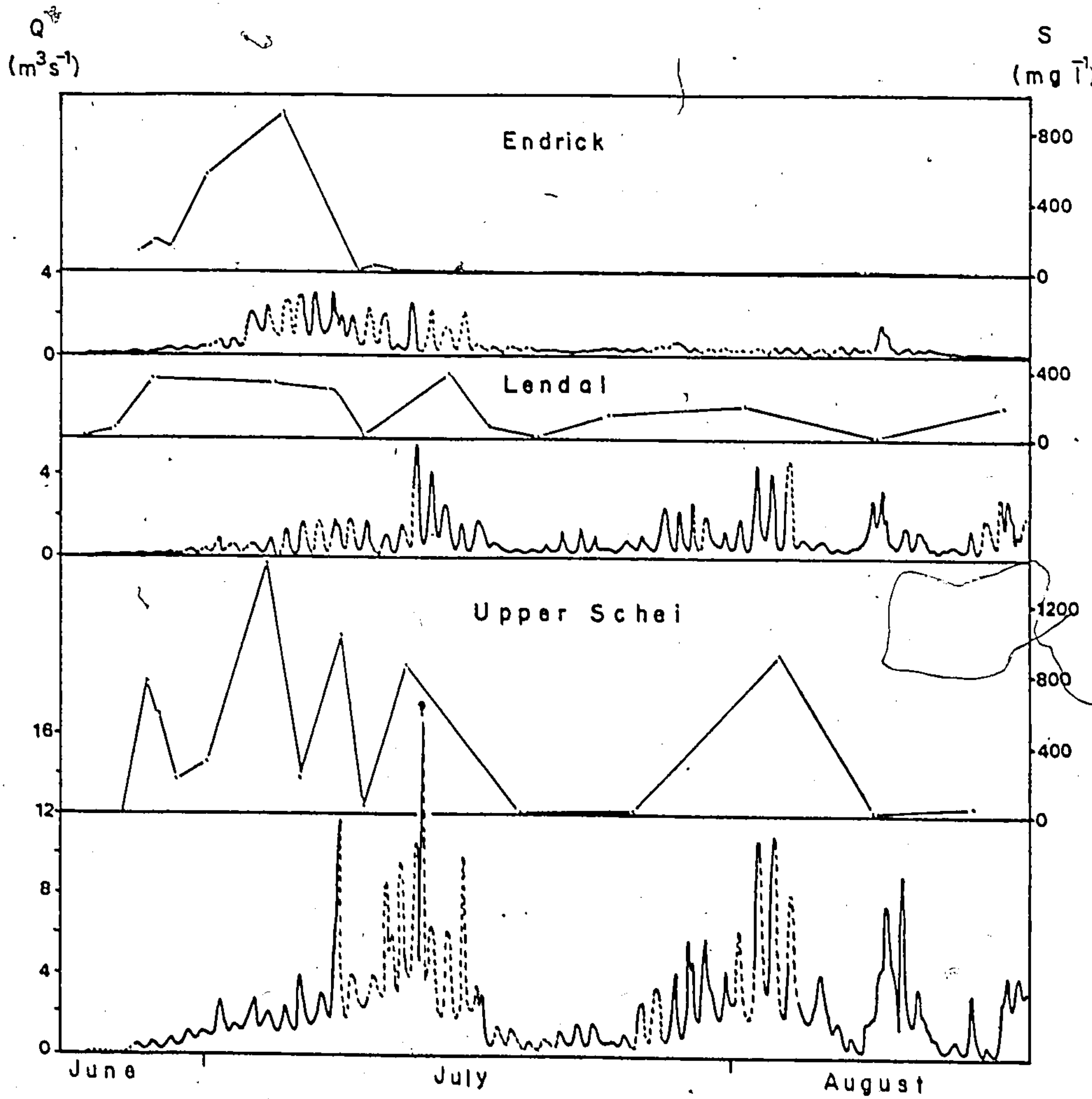


Figure 5.10: "Schei River" tributaries: variations in suspended sediment concentration.

through the development of thermo-erosional niches under ground ice (figure 6.3), flow past dirty thrust planes at the "Schei Glacier" margin and the collapse of sediment-laden glacier ice into the river are thought to be significant.

The most detailed analysis of sediment sources was carried out on "Lendal Creek". It was apparent (figure 5.11) that large amounts of suspended sediment were being supplied to tributaries of this stream at the glacier margin. Samples taken at the "Thistle Creek" and ice-marginal pond sampling sites indicated some major sediment sources. Figure 5.12 depicts these two sites diagrammatically showing the levels of suspended sediment concentration of various sources on August 3. Throughout the season, concentrations in supraglacial meltwater were low, not exceeding 30 mgl^{-1}, but within a few metres of the glacier front extremely high concentrations were measured. Two main sources of sediment were identified:

1. A glacial advance of 2-6 m since the summer of 1973 was locally apparent from photographic and other evidence. This advance has forced streams flowing along the proglacial margin to cut new channels in the silt- and clay-rich till that covers much of the proglacial area. Concentrations in such streams exceeded 2,000 mgl^{-1} on several occasions
2. At the glacier margin, large supraglacial streams fall 30 m vertically to the ground below. Spray from these waterfalls often soaked areas up to 15 m from the ice-margin, so that the shallow active layer of the immediate proglacial area became saturated. The result on accidented terrain was the frequent occurrence of mudslumps over the permafrost table (figure 5.13).

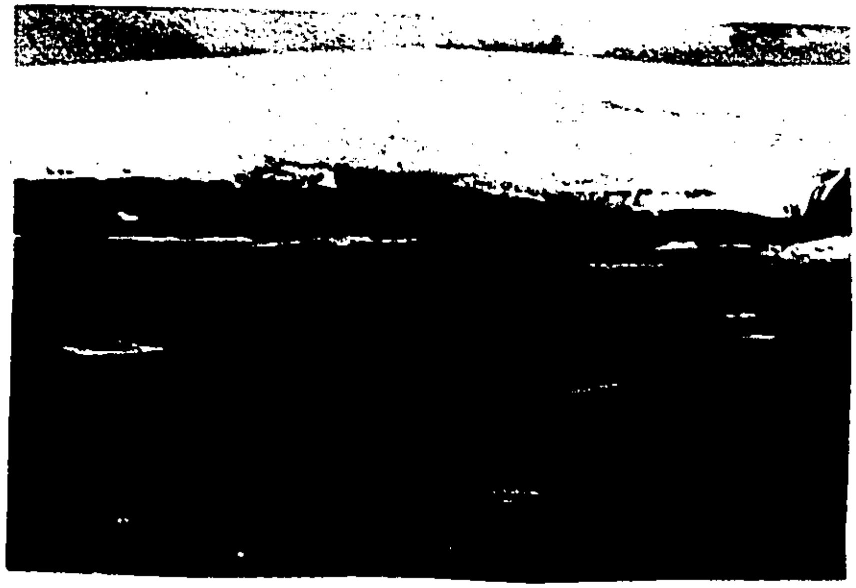


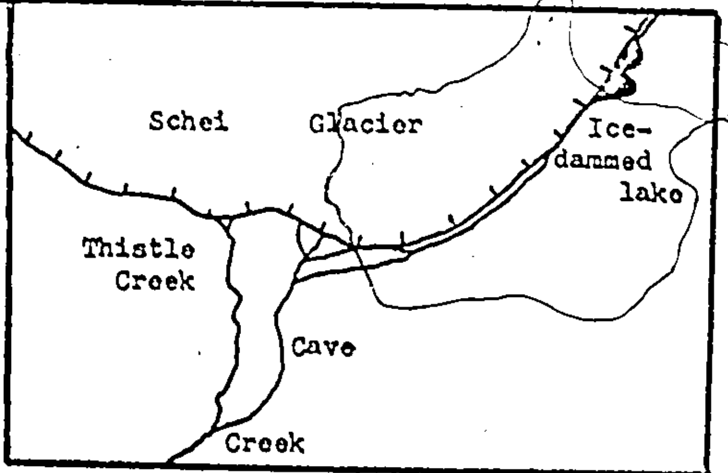
Figure 5.11 : Confluence of the "North Lendal" (left and "South Lendal" Creeks. The turbid "South Lendal" drains accidented terrain at the ice-margin, a source area of abundant fine sediment.

Figure 5.13 : Mudslump at "Thistle Creek", August 3, 1974.

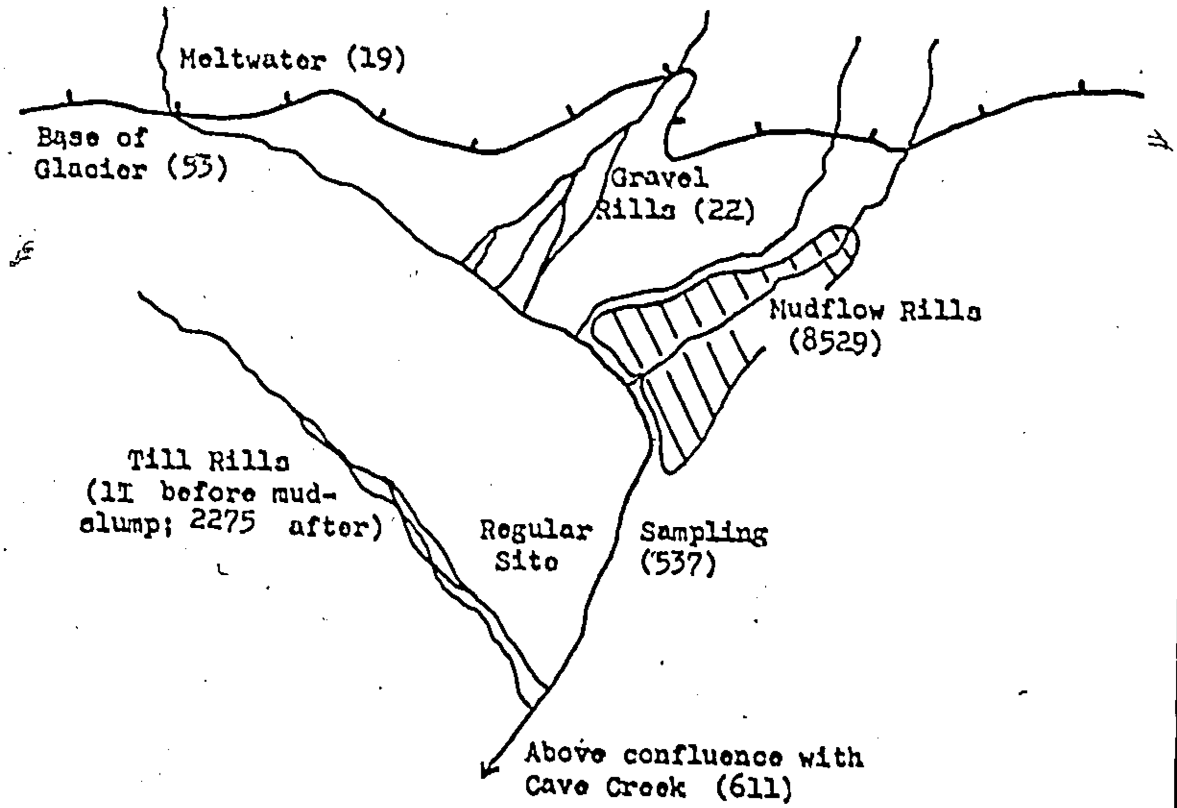
Figure 5.12:

LENDAL BASIN ICE-FRONT SEDIMENT SAMPLING SITES

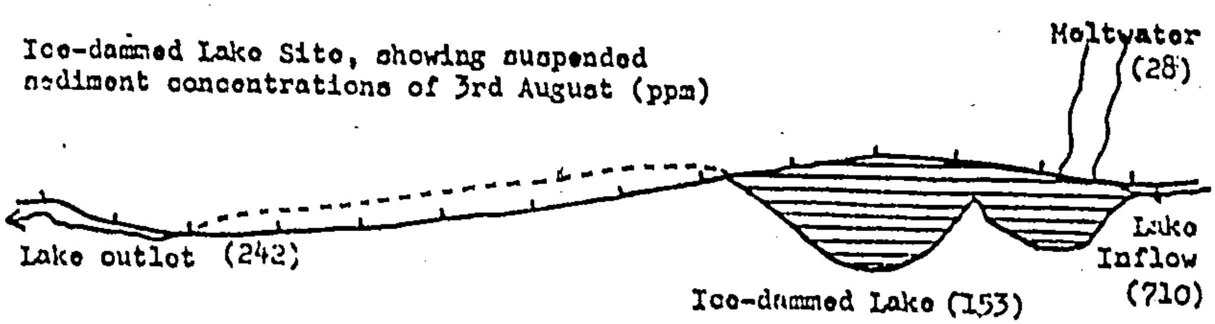
(Diagrammatic; not to scale)



Thistle Creek Site, showing suspended sediment concentrations of 3rd August (ppm)



Ice-dammed Lake Site, showing suspended sediment concentrations of 3rd August (ppm)



From the observations it emerges that the instability of the immediate proglacial area of the advancing, sub-polar "Schei Glacier" is a significant factor in the supply of fine sediment to the "Schei River" system.

5.5. Conclusions

In this chapter, measurements of the concentration of dissolved and suspended inorganics carried by the "Schei River", its three tributaries and ice-marginal meltwater streams in the summer of 1974 have been used to investigate levels of dissolved and suspended load, the relationships between concentration and discharge, the factors affecting these relationships and some of the principal sediment sources. In this section the main conclusions are restated and comparisons are made with the results of similar investigations in high arctic areas.

For the "Schei River", values of specific conductivity ranged from 36 - 206 μmho cm^{-1}, from which a range of total (Ca^{2+} + Mg^{2+}) hardness of 15 - 97 ppm was derived using a hardness - conductivity rating equation. This same equation indicated that the concentration of ions other than Ca^{2+} and Mg^{2+} was negligible, so that total hardness is virtually equivalent to concentration of total dissolved solids. Both conductivity and total hardness displayed an inverse nonlinear relationship with "Schei River" discharge, such that increase in discharge was more rapid than concentration decrease. A calculated total of 1517 metric tons (= 16.6 t km^{-2} or 25.8 t km^{-2} per unit ice-free area) was removed in solution during the study period.

By comparison, the range in suspended sediment concentration was

5 - 2,205 mgl^{-1}. Concentration of suspended sediment varied with discharge in a near-linear fashion, and the main sources of variance in this relationship were hypothesized to be the operation of a hysteresis effect and vicissitudes in the rate of supply of fine sediment. Total seasonal suspended load was calculated to be 10,251 metric tons (=112.4 t km^{-2} or 174 t km^{-2} per unit ice-free area). The amount of material removed in suspension was thus 6.7 times greater than the amount removed in solution.

Limited sampling at the three tributary gauging stations indicated that although concentrations of dissolved load were highest for nival "Endrick Creek", probably on account of consistently low discharge, the highest dissolved load per unit area was that of glacial "Lendal Creek". The tributary samples also revealed relatively high suspended sediment concentrations at the onset of the flow season, a general covariance of suspended concentration with discharge for the glacial tributaries, and a sharp decline in the concentration of suspended inorganics in "Endrick Creek" after peak snowmelt discharge was attained. Sampling at the glacier front showed that supraglacial meltwaters exhibited low hardness values, although concentration increased markedly as the meltwaters flowed over proglacial sediments. Hardness values remained relatively low, however; high pH readings (pH 7.5 - pH 9.2) obtained for glacial meltwaters suggested that CO_2 depletion may be responsible for a lowering in saturation level, and thus for the low hardness values measured. Supraglacial meltwaters were also found to contain low concentrations of suspended sediment, but considerable amounts of fine sediment were eroded from the immediate proglacial zone as new meltwater channels were cut and as saturated till slumped over the permafrost table into proglacial streams.

To facilitate comparison of the measurements reported above with information on the range of dissolved and suspended concentrations in other high Arctic streams, the available data was tabulated (table 5.2) together with data obtained by the writer for the "Schei Basin" in the summer of 1975. Only very general trends emerge. Conductivity, total hardness and calcium (H_{Ca}) hardness values are generally higher for nival streams such as "Jason's Creek", "Endrick Creek" and the Mecham River, but for large rivers this pattern disappears : the high conductivity values measured for the glacial "Sverdrup River" in 1973 and 1974 indicate concentrations in excess of those measured by Smith (1969, 1972) in large nival streams on Somerset Island. Direct comparison is complicated, however, by differences in the proportion of carbonate terrain in these basins. The lower concentrations measured in glacial rivers such as the "Lendal", "Upper Schei" and "Schei" may reflect higher discharges, or may result from lower saturation levels, as discussed above.

In contrast, suspended sediment concentrations are greater for glacial streams (Decade, "Upper Schei", "Schei" and "Sverdrup" Rivers) than for nival streams of similar basin size. Also evident is an increase in concentration with basin size, probably effected by increased discharge. The crude nature of this comparison must be stressed, however, as the availability of suitable material and the occurrence of exceptional hydrologic events may result in unrepresentative ranges of concentration. Although some of the suspended concentrations measured in glacial streams are high, they fall short of values recorded in the proglacial zones of "temperate" glaciers, for example the 17,200 ppm obtained by Fahnestock (1963)

TABLE 5.2

RANGE OF DISSOLVED AND SUSPENDED SEDIMENT CONCENTRATIONS OBTAINED FOR SELECTED HIGH ARCTIC STREAMS

(Listed by increasing basin area)

River	Location	Basin Area (km^2)	%Glaci-erized	Ter-rain[1]	Year	K (μmho cm^{-1})	H_t (ppm)	H_{ca} (ppm)	S ($mg\,l^{-1}$)	Source
"Jason's Creek"	Devon Is.	2.3	0.0	C	1970	-	29-102	6-79	0-291	Cogley, 1971
"Lendal Creek"[2]	Ellesmere Is.	9.4	62.8	P	1974	32-98	15-58	10-32	2-381	This Study
"Lendal Creek"[2]	Ellesmere Is.	9.4	62.8	P	1975	33-64	16-26	11-23	-	Unpub. Data
Decade River	Baffin Is.	12.8	68.0	?	1965	-	-	-	0-1183	Østrem, Bridge and Rannie, 1967
"Endrick Creek"[2]	Ellesmere Is.	29.9	0.0	P	1974	71-171	33-89	20-60	0-877	This Study
"Endrick Creek"[2]	Ellesmere Is.	29.9	0.0	P	1975	139-178	75-43	60-79	-	Unpub. Data.
"Upper Schei R."[2]	Ellesmere Is.	49.3	54.3	P	1974	29-104	15-52	6-32	2-1410	This Study
"Upper Schei R."[2]	Ellesmere Is.	49.3	54.1	P	1975	24-77	10-34	6-22	-	Unpub. Data.
"Schei River"	Ellesmere Is.	91.2	35.5	P	1973	39-96	-	-	0-4900	McCann et al., 1975
"Schei River"	Ellesmere Is.	91.2	35.5	P	1974	36-206	15-97	8-64	5-2204	This Study
"Schei River"	Ellesmere Is.	91.2	35.5	P	1975	32-122	14-56	10-47	-	Unpub. Data
Mecham River	Cornwallis Is.	97.7	0.0	C	1970	-	33-90	-	0-571	McCann, Howarth and Cogley, 1972
Mecham River	Cornwallis Is.	97.7	0.0	C	1971	88-210	65-135	-	-	McCann et al., 1975
"Sverdrup River"	Ellesmere Is.	1630	77.0	P	1973	53-233	-	-	12-8100	McCann et al., 1975
"Sverdrup River"	Ellesmere Is.	1630	77.0	P	1974	51-230	-	-	20-2900	McCann et al., 1975
"Sverdrup River"	Ellesmere Is.	1630	77.0	P	1975	50-168	21-49	14-43	-	Unpub. Data
Various Major Rs.[2]	Somerset Is.	-	0.0	P	1964	-	25-66	1-29	-	Smith, 1969
Colville River	Alaska	50,000	Very	?	1962	-	-	-	19-1650	Arnborg, Walker and Peippo, 1967.

Notes: 1. C = carbonate, P = partly carbonate, ? = unknown

2. Based on less than 20 samples.

in the glacial White River at Mt. Rainier, Washington.

The relationship between dissolved concentration and discharge obtained for the "Schei River" (equation 5.4) is essentially of the same form as that presented by Cogley (1972) for data from "Jason's Creek" and the Mecham River, and by Church (1972) for four Baffin Island streams on non-carbonate terrain. It seems likely that the general form of the solute-discharge relationship for arctic streams is

$$H_t = \nu \; Q^{-\omega} \qquad (5.14)$$

where ν and ω are constants and $0 < \omega < 1$. Similarly, the form of the suspended concentration-discharge relationship resembles that found by Church:

$$S = x \; Q^{y} \qquad (5.15)$$

where x and y are constants, and $y \simeq 1.0$.

The erosion of till at or near the glacier front, as reported above, was also considered important in supplying fine sediment to arctic rivers by Maag (1969) and Church (1972). The present study has indicated that solution in the same zone is also of some importance, but less so than solution on ice-free arctic carbonate terrain (McCann and Cogley, 1971; Cogley, 1972; Wilkinson, 1972; Smith, 1969, 1972). Knowledge of the sediment contribution of different terrain types remains very scant, however, and deserves a great deal more study before the sediment transport and supply of arctic stream basins can be adequately described.

CHAPTER 6

FLUVIAL AND GLACIOFLUVIAL LANDFORMS

6.1 EROSIONAL FORMS

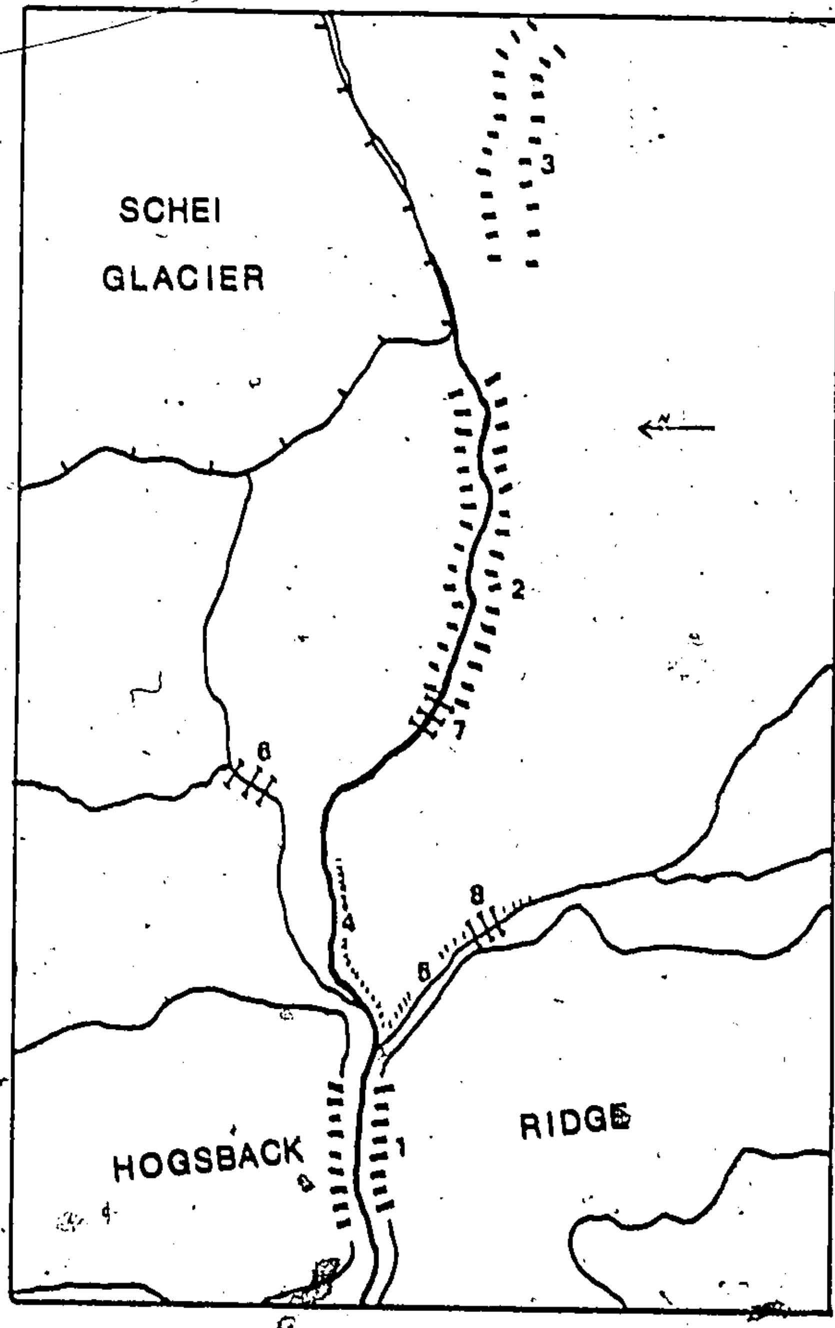
River action in the high arctic has produced a number of distinctive landforms that are rare or absent in lower latitudes. One such group of feature consists of the deep gorges, with vertical or near-vertical walls, which have been described in many parts of the high arctic. Jenness (1952), working in the western arctic archipelago, described

> "deep trenches in the underlying bedrock ... ravines as much as 200 or 300 feet deep..."
>
> (p. 240)

and similar features variously termed "box canyons", "narrow V-shaped valleys" and "steep-sided ravines" have been described from elsewhere in the Canadian Arctic by Robitaille (1958, 1960), St-Onge (1965) and McCann and Cogley (1974).

It was argued in chapter 3 that the "Upper Schei Gorge" (figures 3.14 and 6.1) is an entirely postglacial features, downcut as a consequence of falling sea-level. Some mass-wastage scars on the ground above the gorge constitute the only evidence of recent slope development, even in the weak sedimentary rocks of the Eureka Sound Formation. One is led to conclude, with Jenness (1952), that

> "... the powerfully eroding streams that deeply incise their own valleys have little effect on the denudation of the area



1 "Schei Gorge"
2 "Upper Schei Gorge"
3 Meltwater gorge
4 Thermo-erosional niches
5 Terrace erosion
6 Channel cross-section
7 Channel cross-section
8 Channel cross-section

Figure 6.1: Map of erosional landforms mentioned in

through which they pass"

(p 248)

the central question raised by arctic gorges does not concern the rapidity of their erosion, but the stability of their walls.

Unlike the "Upper Schei Gorge", the "Schei Gorge" is almost certainly polygenetic. It is incised into a structural syncline that probably formed a routeway for ice movement, although glacial modification is not evident. Even though up to 9 m of fluvial downcutting has taken place since deglaciation, it would be misleading to describe the "Schei Gorge" as the product of exclusively fluvial erosion.

Of greater interest is a ravine, cut 200 m into dolomite, parallel to the south margin of the "Schei Glacier" (figure 6.2). The channel at the base of this impressive feature, presently occupied by a very small stream, is apparently "graded" to the lateglacial surface at the top of the "Upper Schei", implying formation in the period immediately following the withdrawal of ice. If this feature is entirely of postglacial development it would therefore appear to have been cut, presumably by glacial meltwater, in the relatively short interval between ice retreat and the abandonment of the lateglacial surface, testimony to the considerable erosive power of glacial meltwater in the high arctic.

The lateral erosion of unconsolidated deposits operates on a very different time-scale to the downcutting of bedrock and is readily observable in the field. In many permafrost areas lateral erosion occurs through the development of a thermo-erosional niche which occurs when the river melts a horizontal slot in the adjacent permafrost or ground ice (Walker and Arnborg, 1966; Maag, 1969). In 1974 the "Upper Schei River"

Figure 6.2 : Ravine cut by glacial meltwater near the south margin of the "Schei Glacier"

Figure 6.3 : Bank collapse following the development of a thermo-erosional niche at the distal end of the "Endrick" relict sandur, August 4, 1974.

developed a niche up to 9.0 m deep under the deposits of the relict "Endrick" sandur. Bank collapse occurred (figure 6.3) along the lines of ice-wedge cracks, and large blocks of frozen sediment were dumped in the river. Bank retreat, measured with reference to a stake placed in the relict sandur surface in August 1973, had exceeded 4.5 m by August 15, 1974. This figure is probably a maximum for the bank as a whole.

Bank retreat over the 1974 runoff season was also measured at the western margin of the "Endrick" relict sandur, where "Endrick Creek" is actively eroding banks of coarse gravel (figures 6.4 and 6.5). No niche was developed. The average distance of retreat was only 0.13 m (maximum 0.45 m). This slower rate of erosion is probably attributable to the greater discharge of the "Upper Schei River" and the coarser nature of the deposits eroded by the "Endrick" rather than the non-development of a thermo-erosional niche.

Net channel change during the snowmelt flood period was measured for each of the "Schei River" tributaries by levelling accurate sections across straight reaches immediately upstream or downstream of the tributary gauging sites. The results (figure 6.6) show that net change was negligible and absolute change minimal, except where appreciable bank erosion took place as in "Upper Schei" profile 2. The channel form of all three tributaries is apparently stable, at least at discharges less than those experienced at the peak of the 1974 snowmelt flood.

6.2. DEPOSITIONAL LANDFORMS

6.2.1 Introduction

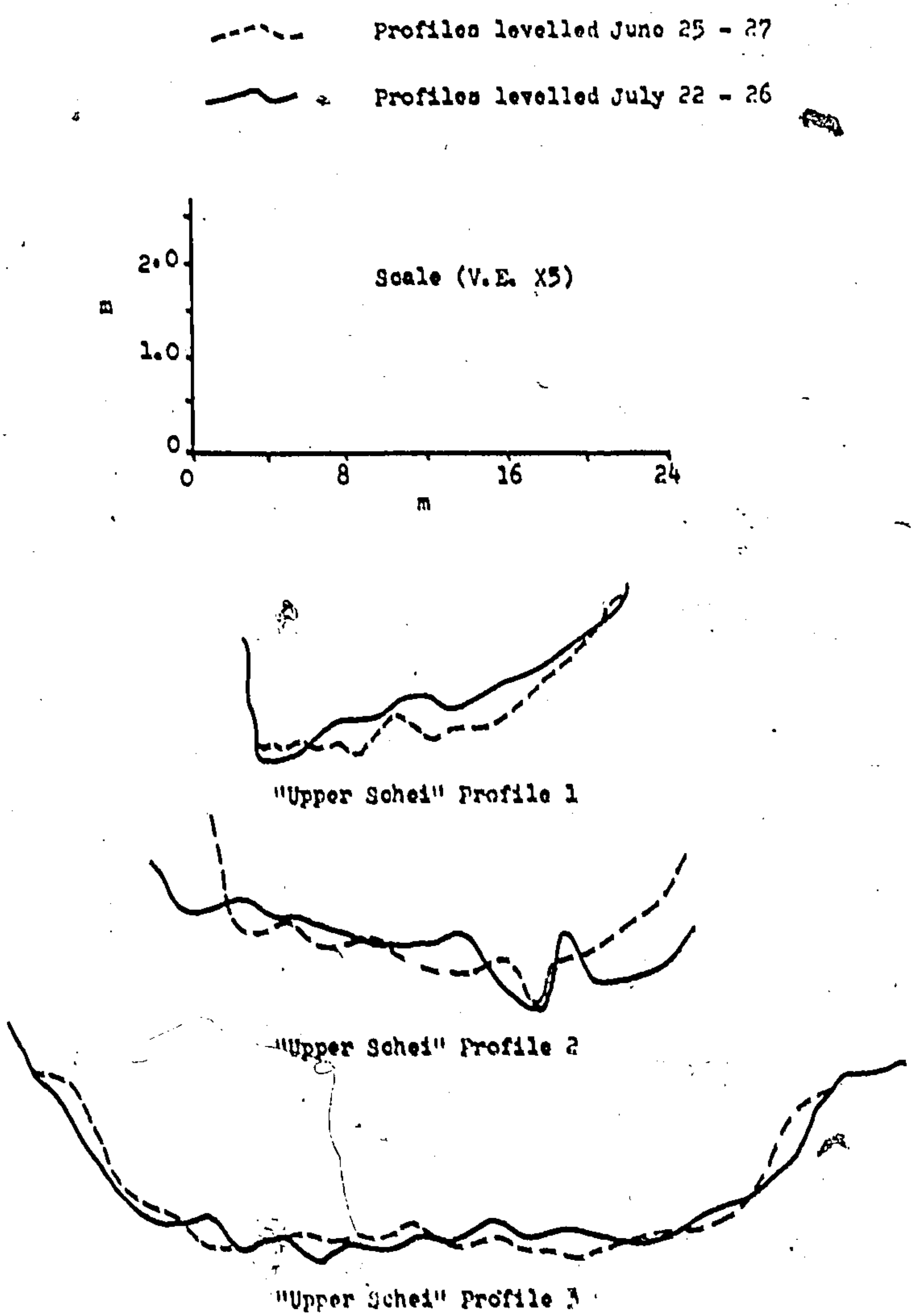
The two major landforms of fluvial on glaciofluvial deposition

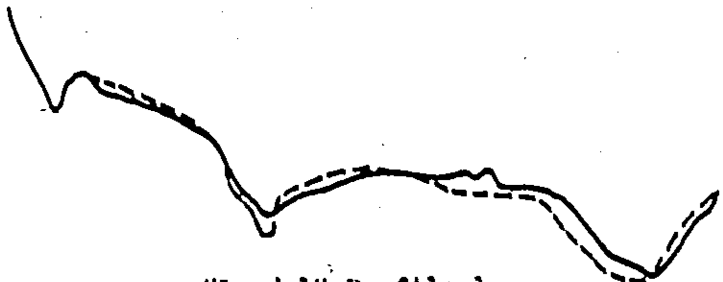
Figure 6.4 : "Endrick Creek", downstream from the gauging site, 1430 h, July 7. Q_e = $2.6\ m^3\ s^{-1}$
Extensive recent bank erosion.

Figure 6.5 : "Endrick Creek" downstream from the gauging site, 1700 h, July 19. Q_e = $0.4\ m^3\ s^{-1}$
Note readjustment of river bank after cessation of bank erosion.

Figure 6.6:

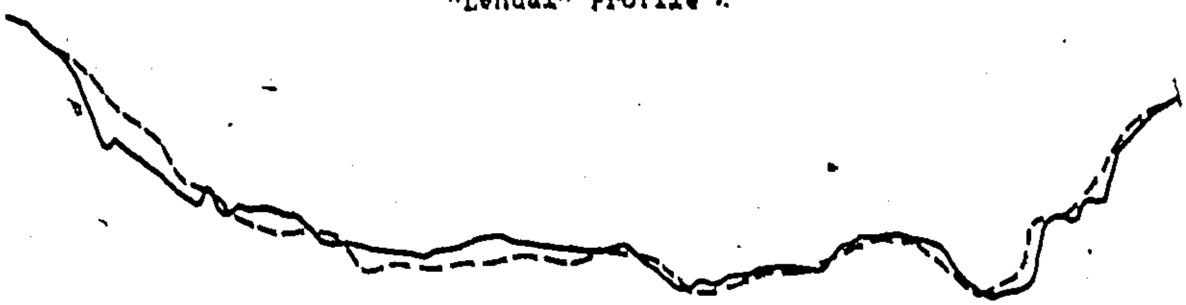
LEVELLED CROSS-SECTIONS ON "SCHEI RIVER" TRIBUTARIES



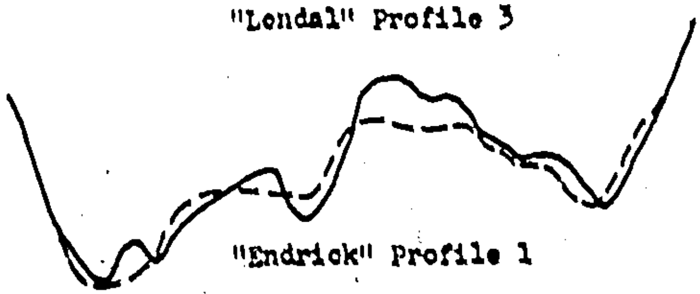


"Lendal" Profile 1

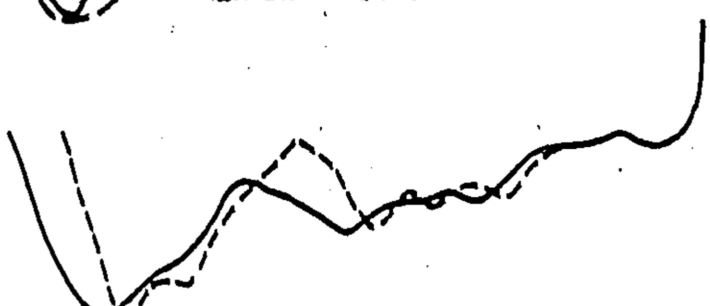
"Lendal" Profile 2



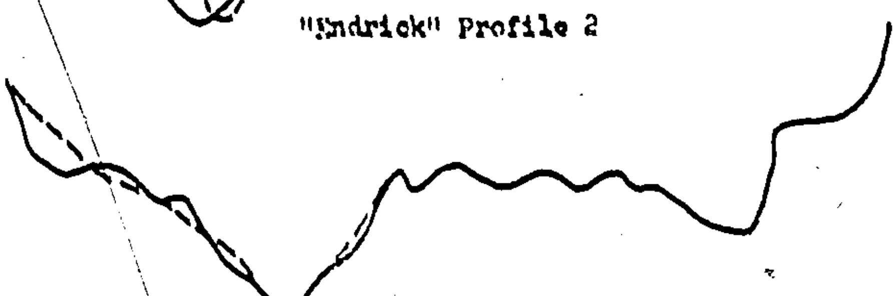
"Lendal" Profile 3



"Endrick" Profile 1



"Endrick" Profile 2



"Endrick" Profile 3

in arctic areas are the delta and sandur our outwash plain. Two small sandar constitute the main features of fluvial deposition in the "Schei River" basin. Both have been intensively investigated : the characteristics of the "Schei Sandur" have been studied by Bennett (1975, McCann et al., 1975) and various aspects of the smaller "Upper Schei" sandur (figures 6.7 and 6.8) were investigated as part of the present study.

The dominance of sandar in the hierarchy of arctic fluvial landforms has been stressed in chapter 1 and requires no further elaboration. Considerable information is now available on the characteristics of arctic sandar (Church, 1972). As a broad, single-season study of the "Upper Schei" sandur would be unlikely to yield more than confirmation of previous findings, it was decided to investigate two outwash characteristics which have received only limited attention in the literature, namely

1. the nature of mean cobble size distribution in different sedimentological environments on the surface of the active sandur; and
2. the comparison of cobble size distribution on the active surface with that on steeper terraced fragments.

6.2.2 The Form of the "Upper Schei" Sandur

The "Upper Schei River" sandur is 1,000 m long and has a maximum width of 180 m (figure 6.9). It is essentially a "valley train" or "dalsandur" (Krigström, 1962) although the lateral boundaries consist of banks of terraced till and outwash deposits rather than bedrock walls. The crescentic plan of the sahdur is largely due to the bedrock constriction of the "Schei Gorge" at its distal end. This constriction has

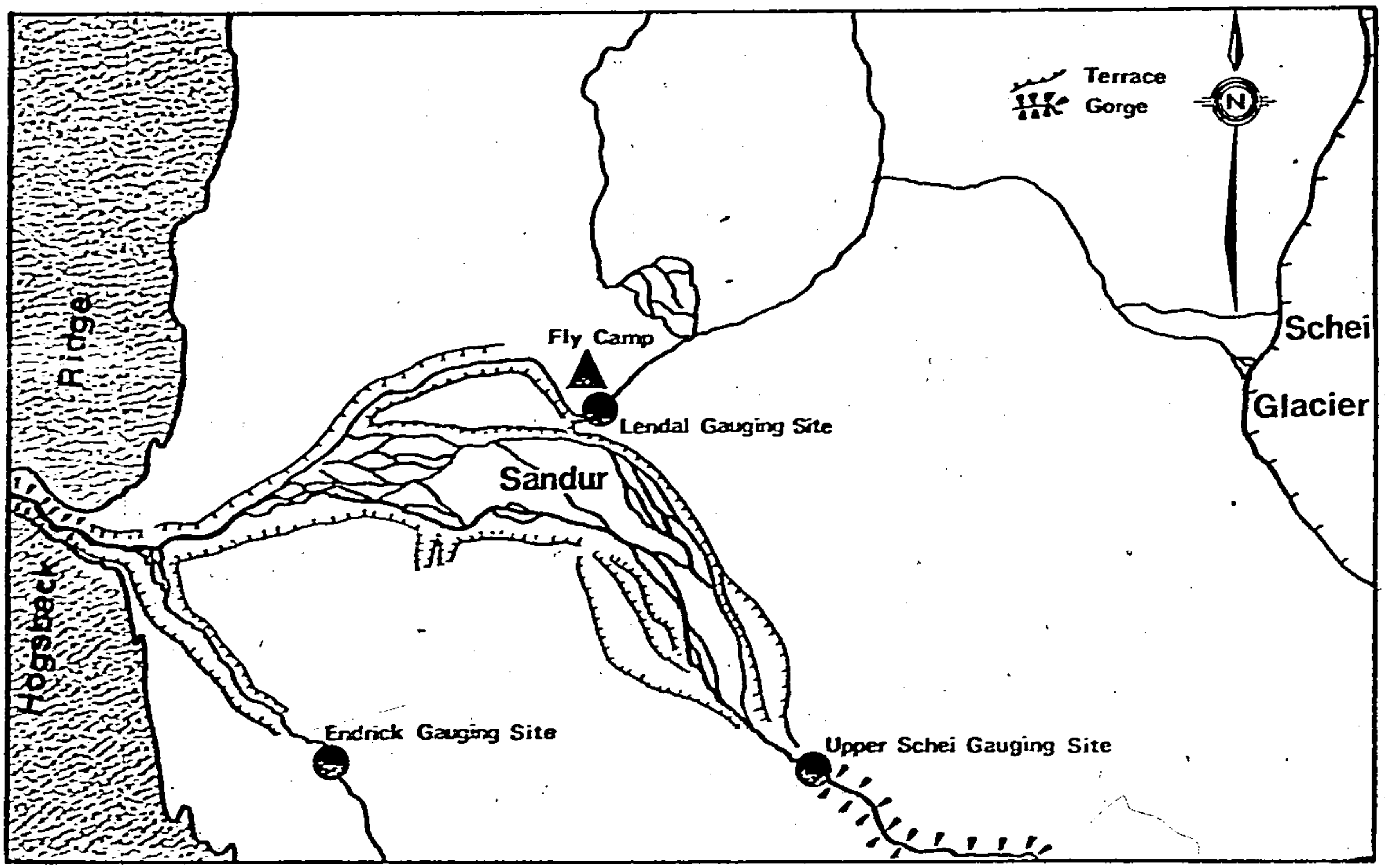


Figure 6.7: Location of the "Upper Schei" sandur.

Figure 6.8 : The "Upper Schei" sandur from the "Hogsback Ridge", June 25, 1974.

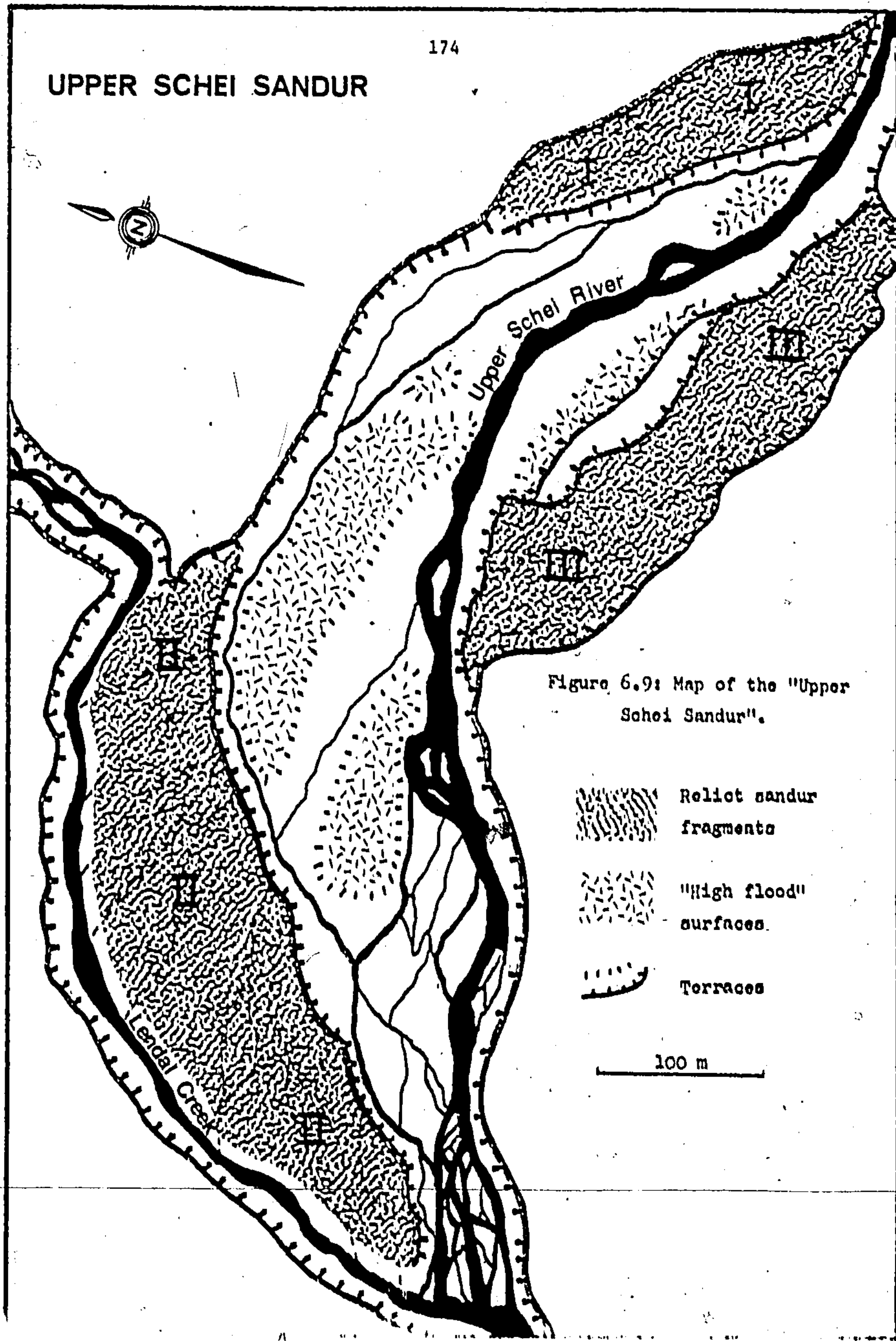


Figure 6.9: Map of the "Upper Schei Sandur".

the effect of drawing together all the sandur distributaries to a single channel, inhibiting the formation of a more typical fan-shaped feature. The bedrock constriction also acts as a local base level to which the sandur surface is graded. Downcutting of this base level in phase with postglacial drops in sea-level has caused a progressive lowering of the active sandur surface, leaving a number of terraced remnants.

The long profile of the main channel on the "Upper Schei" sandur was levelled and found to approximate

$$H = 0.000486\, D^{1.5856} \qquad (r = 0.996) \tag{6.1}$$

where H is height in metres above an arbitrary datum at the distal end of the sandur, and D is distance in metres from the same point. Bennett (1975) obtained the approximation

$$H = 0.996\, e^{0.0004D} \qquad (r = 0.97) \tag{6.2}$$

for the main channel profile on the "Schei River" sandur, which is less steep and of a different form. The general form of the equation and used by Bennett

$$H = K_1\, e^{K_2 D} \tag{6.3}$$

has been used by several workers to approximate the long profiles of main sandur channels (Thorarinsson, 1939; Fahnestock, 1963, 1969; Church, 1972), but when fitted to the "Upper Schei River" data, as

$$H = 0.939\ e^{0.0041D} \tag{6.4}$$

$$(r = 0.93)$$

gives a poorer fit than equation 6.1.

A three-dimensional approximation of the form of the sandur surface was obtained through the use of trend surface analysis (Krumbein and Graybill, 1965). A first order surface, based on 54 data points, was fitted to most of the sandur area (figure 6.10). This surface, with the equation

$$E = 87.47 - 0.63 \times 10^{-1}\ x - 0.86 \times 10^{-8} Y \tag{6.5}$$

explains 98.9% of the height variation. Slight increases in explanation were obtained by fitting second-order (99.2%) and third-order (99.6%) trends. The third order surface (figure 6.11) is interesting in that it indicates the presence, in the north-central part of the sandur, of a "plateau" corresponding with the "high flood" surfaces depicted on figure 6.9. The significance of these "high flood" surfaces is discussed below.

Krigstrom (1962) devised a three-part zonation of Icelandic dalsandar, later endorsed by Church (1972) and Gustavson (1974). The proximal zone is typically crossed by a few main channels, often incised; in the intermediate zone there is extensive braiding, with wide, shallow channels; and the distal zone is essentially transitional to a delta, being covered by a single sheet of water. On the "Upper Schei" sandur, areas corresponding to Krigstrom's proximal and intermediate zones can be recognised, and it was possible to subdivide the intermediate zone into areas entirely active during high discharge (approximately > 8.0 $m^3\ s^{-1}$) and areas only partly

Figure 6.10: First order trend surface of sandur elevation.

(Heights in metres)

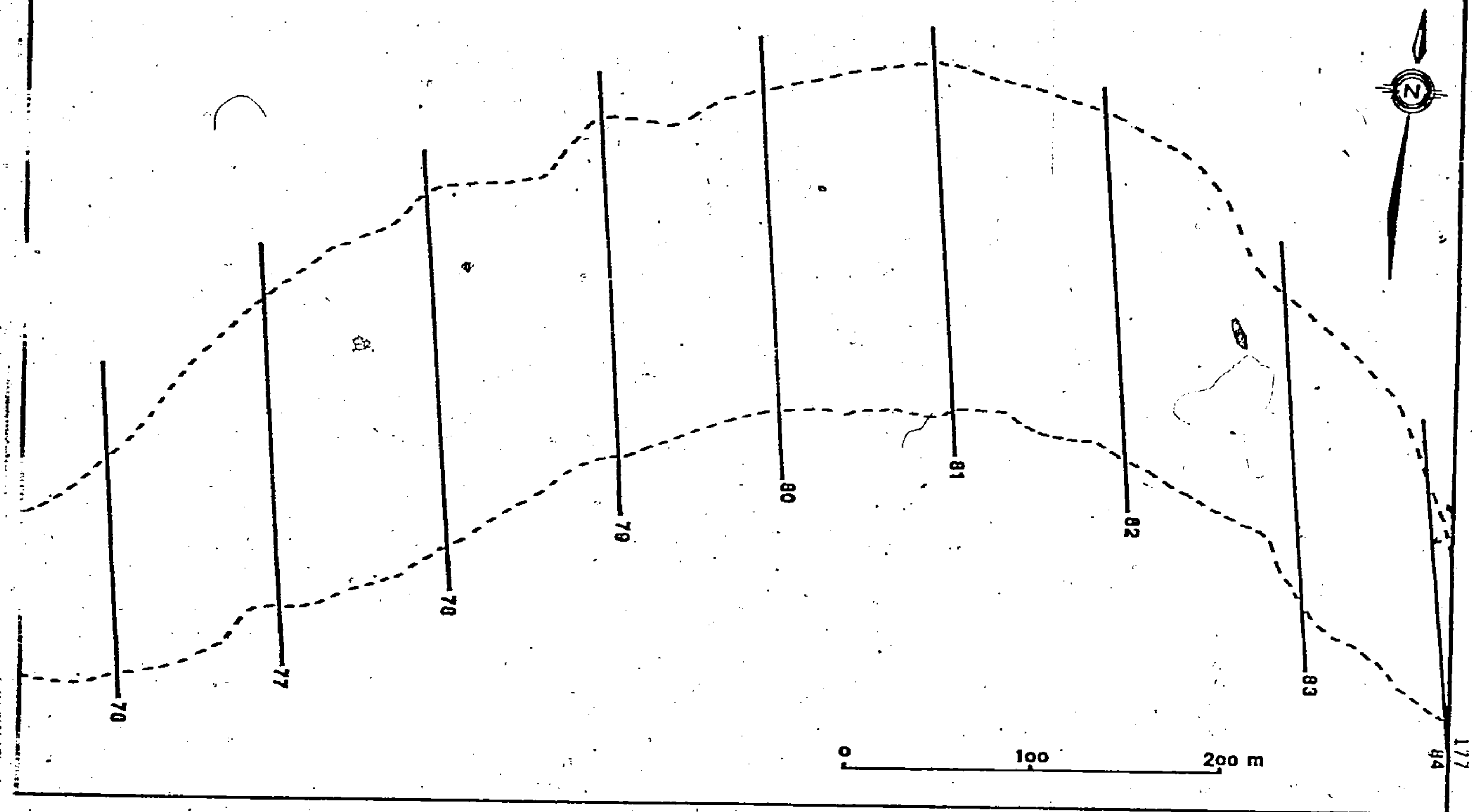
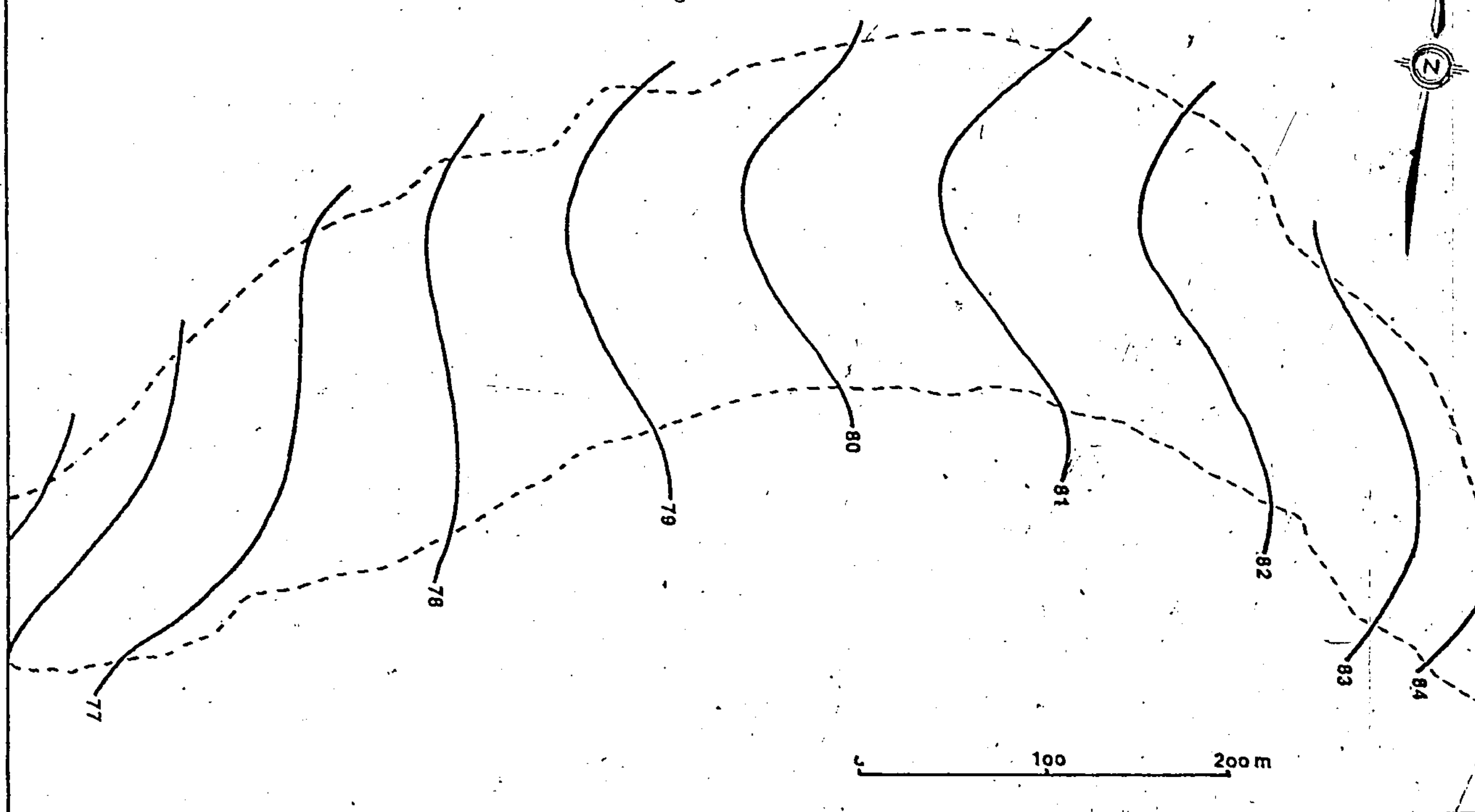


Figure 6.11: Third order trend surface of sandur elevation.

(Heights in metres)



active (figure 6.12). A true distal zone was absent.

6.2.3 Techniques employed in the analysis of cobble-size distribution

In order to investigate downsandur variation in cobble size for different sedimentological environments, it was necessary to :

1. Establish a sampling pattern that would permit down-sandur variation in mean clast size to be distinguished from variations between different depositional "environments".
2. Establish a simple classification of "environments", allocating each sample to a particular class.
3. Employ analytical techniques designed to separate down-sandur trends in clast-size variation from variations attributable to deposition in different "environments".

A suitable sampling scheme was established by levelling five cross-sandur transects across the "Upper Schei" sandur establishing sampling sites at irregular intervals along each transect (figure 6.13). The planimetric position and height of each site was determined using a Kern GK1a level and Wild T.2 theodolite. Each site was then assigned to one of the depositional classes depicted on figure 6.14. The main channel (C_1) and main channel bars (C_2) areas are active even at low (< 6.0 m^3 s^{-1}) discharges. The flood channel (F_1) and flood channel bar (F_2) areas carry water at higher discharges. The "high flood" classes (H_1 and H_2) refer to areas not observed to be active during the 1974 season. These are located on the "plateau" revealed by the third-order elevation trend surface (figures 6.9 and 6.11). The "high flood" area is probably only active during extreme rainstorm or jokuthlaup floods, or may represent a recently-terraced remnant, abandoned through avulsion of the main channel.

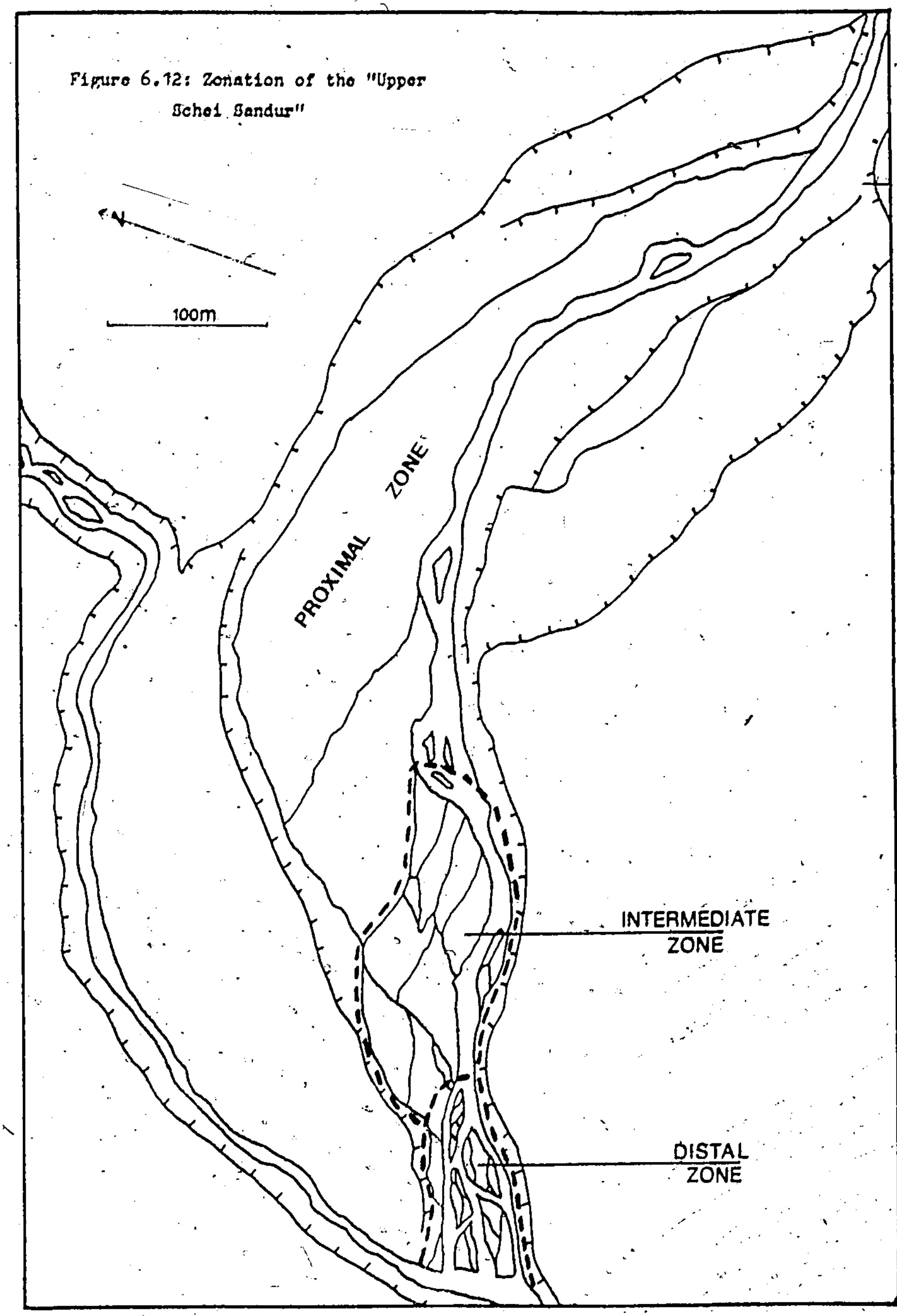
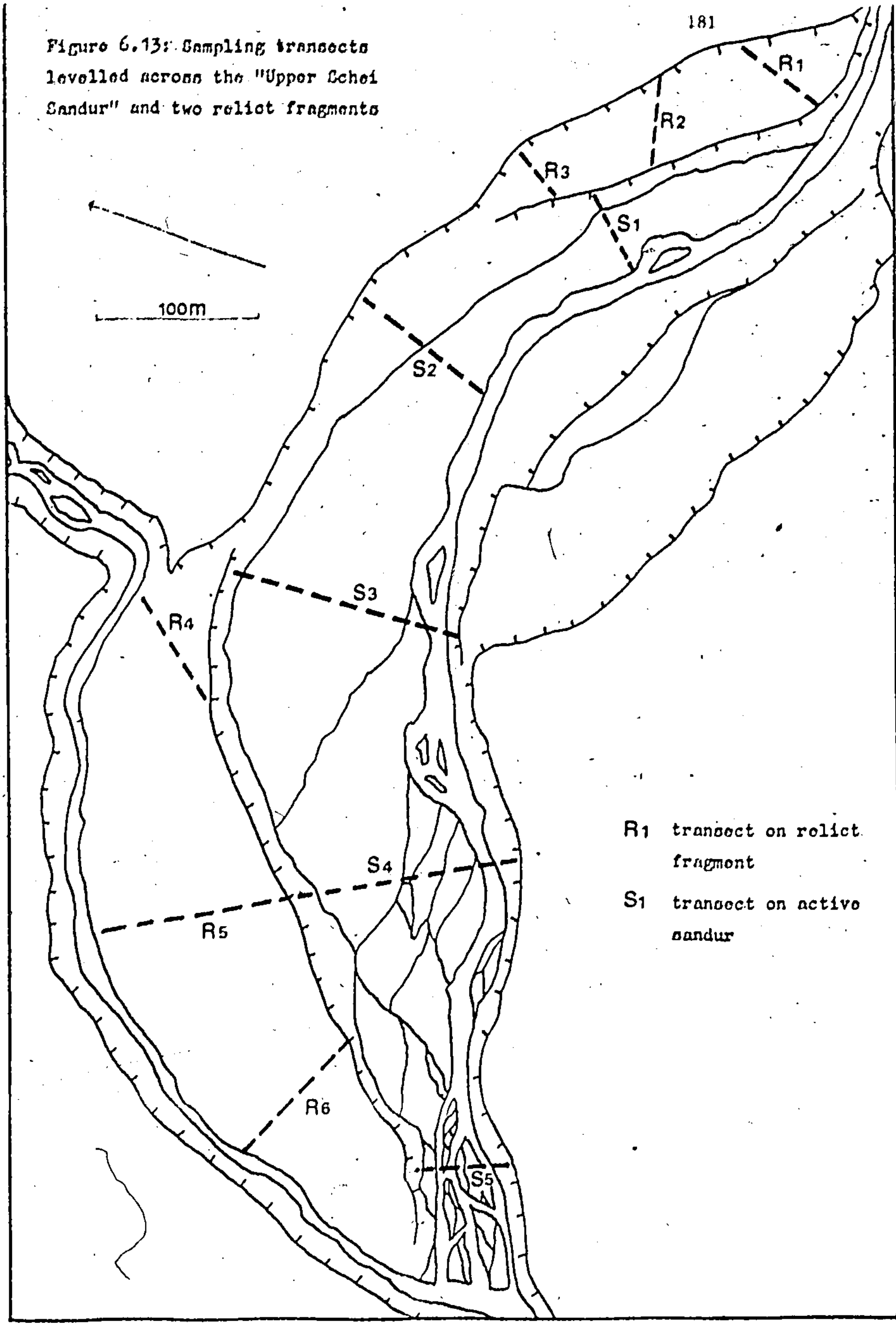


Figure 6.12: Zonation of the "Upper Schei Sandur"
N
100m
PROXIMAL ZONE
INTERMEDIATE ZONE
DISTAL ZONE

Figure 6.13: Sampling transects levelled across the "Upper Schei Sandur" and two relict fragments



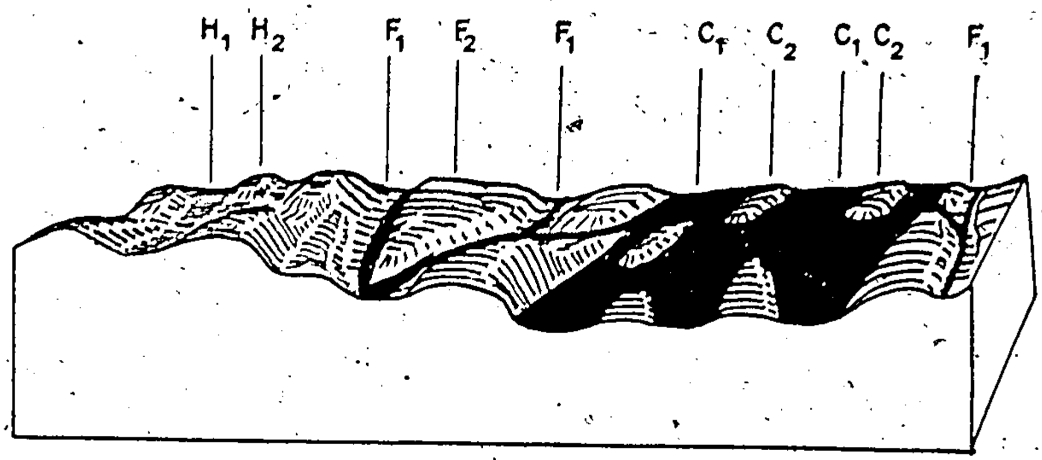


Figure 6.14: Classification of depositional "environments" (diagrammatic).

H_1 High flood channels
H_2 High flood channel bars
F_1 Flood channels
F_2 Flood channel bars
C_1 Main channels
C_2 Main channel bars

Similar "high flood" surfaces have been reported elsewhere (Church, 1972, p 109).

Allocation of each site to a particular class was based on height relationships with adjacent parts of the sandur. Channels and bars were distinguished on the grounds of morphology and only well-defined sites were sampled. The drawbacks of the classification procedure are acknowledged : the classification itself simplifies reality and the allocation of sites is necessarily subjective, but an apparently successful precedent exists in the work of Williams and Rust (1969).

As traditional methods for obtaining mean cobble size data (Wolman, 1954; Leopold, 1970) are excessively time-consuming in the field, sampling was carried out using a photographic procedure devised by Church (1967). This method is summarised in appendix 2. In the present study, coarse clasts are defined as having length of intermediate (B) axis greater than 8 mm (-3 ϕ).

Once the locational and mean cobble size data had been reduced, trend surface techniques were employed to model clast size variation over the sandur surface, and trend residuals (Chorley and Haggett, 1965) were investigated for each class of depositional "environment". The question arises as to whether the clustering of data along five transects will affect the validity of the trend surface technique (Gray, 1972). However, as the "clusters" in the present instance are aligned normal to the anticipated (downsandur) trend, the trend is unlikely to be distorted. Also, a spuriously high level of explanation is unlikely to result from this alignment as the within-transect variations in clast size are large.

6.2.4 Analysis and Results

Both the first order surface of cobble size variation (figure 6.15), with the equations

$$\bar{B} = 70.68 - 0.10\,X + 0.84 \times 10^{-1}\,Y \qquad (6.6)$$

$$(r = 0.594)$$

where $\bar{B}$ is mean cobble size, and X and Y are orthogonal map co-ordinates, and the second order surface (figure 6.16), given by

$$\bar{B} = 68.96 - 0.10\,X + 0.90 \times 10^{-1}\,Y + 0.35 \times 10^{-3}\,X^2$$

$$- 0.23 \times 10^{-2}\,XY + 0.20 \times 10^{-2}\,Y^2 \qquad (6.7)$$

$$(r = 0.624)$$

show a fairly regular down sandur decrease in $\bar{B}$. However, these surfaces explain only 35.1% and 38.9% respectively of the variation in mean clast size. The third order surface (figure 6.17) with the equation

$$\bar{B} = 127.60 - 1.70\,X + 0.48\,Y + 0.11 \times 10^{-1}\,X^2$$

$$+ 0.87 \times 10^{-2}\,XY - 0.35 \times 10^{-1}\,Y^2 - 0.23 \times 10^{-4}\,X^3$$

$$- 0.69 \times 10^{-5}\,X^2Y - 0.44 \times 10^{-4}\,XY^2 + 0.27 \times 10^{-3}\,Y^3 \qquad (6.8)$$

$$(r = 0.657)$$

shows a "plateau" corresponding to the "high flood" areas of the sandur. Even this complex surface explains only 43.2% of mean cobble size variation. A further increase in explanation through the employment of higher order surfaces will involve a trend pattern of utmost complexity, probably defying interpretation.

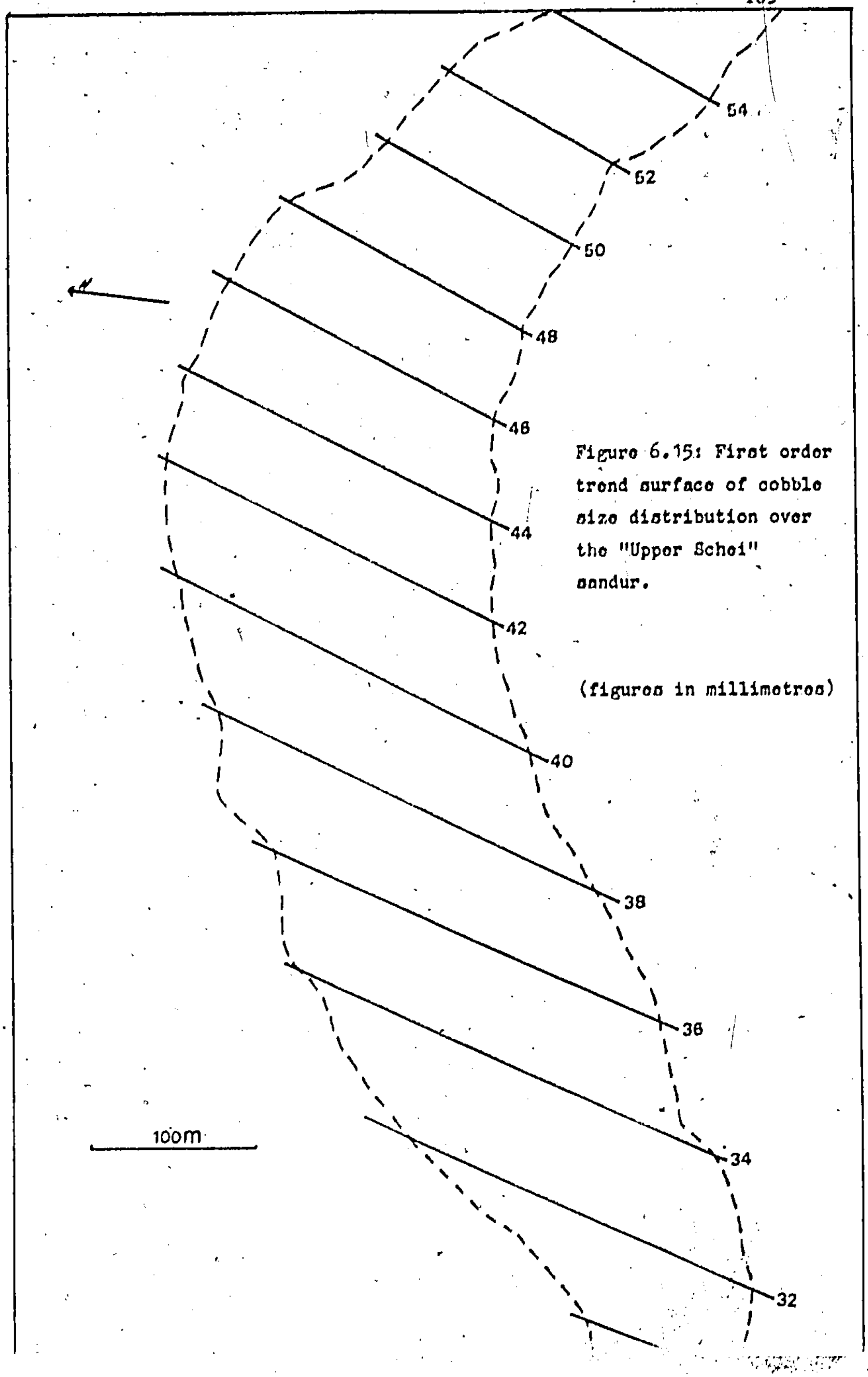


Figure 6.15: First order trend surface of cobble size distribution over the "Upper Schei" sandur.

(figures in millimetres)

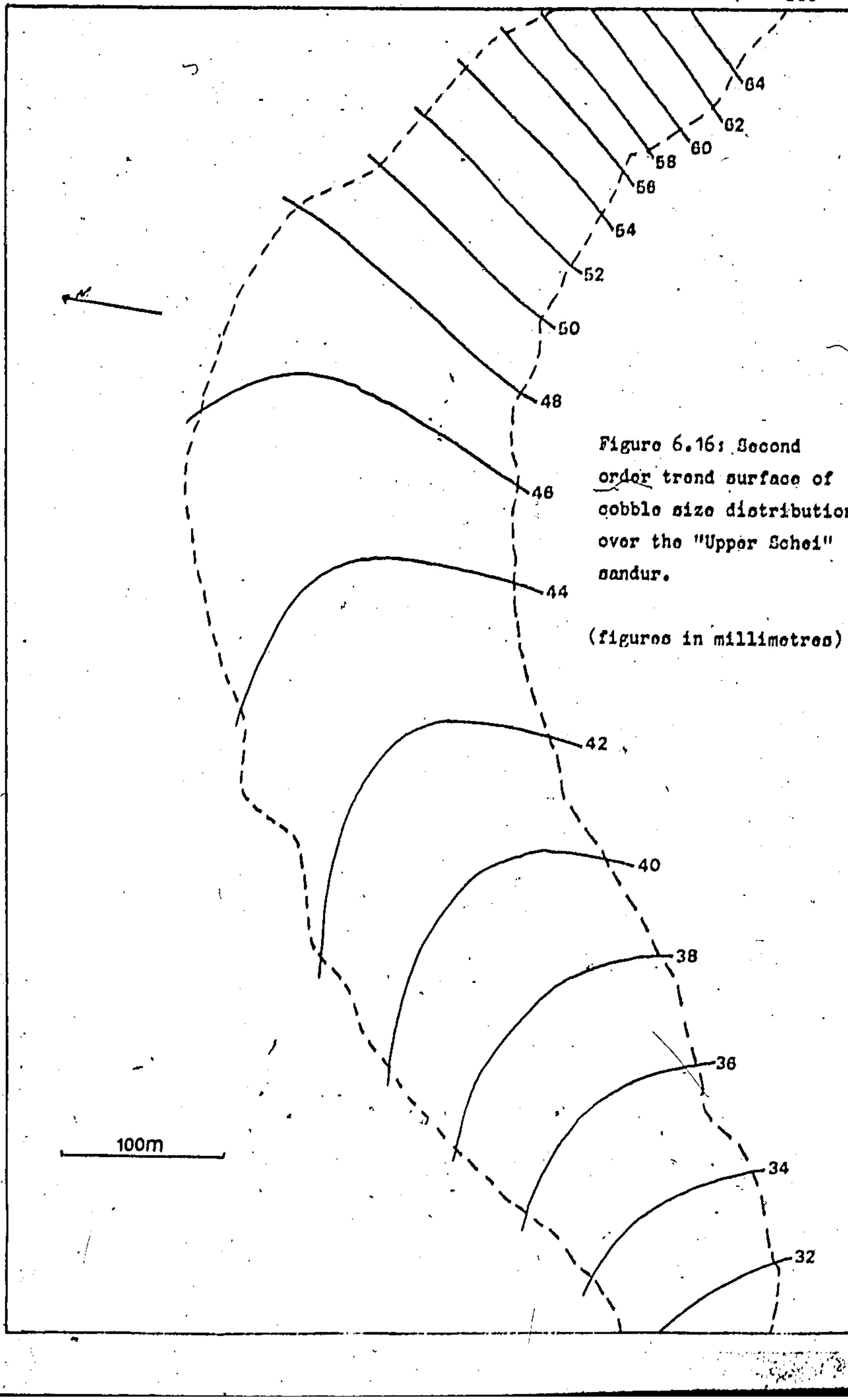


Figure 6.16: Second order trend surface of cobble size distribution over the "Upper Schei" sandur.

(figures in millimetres)

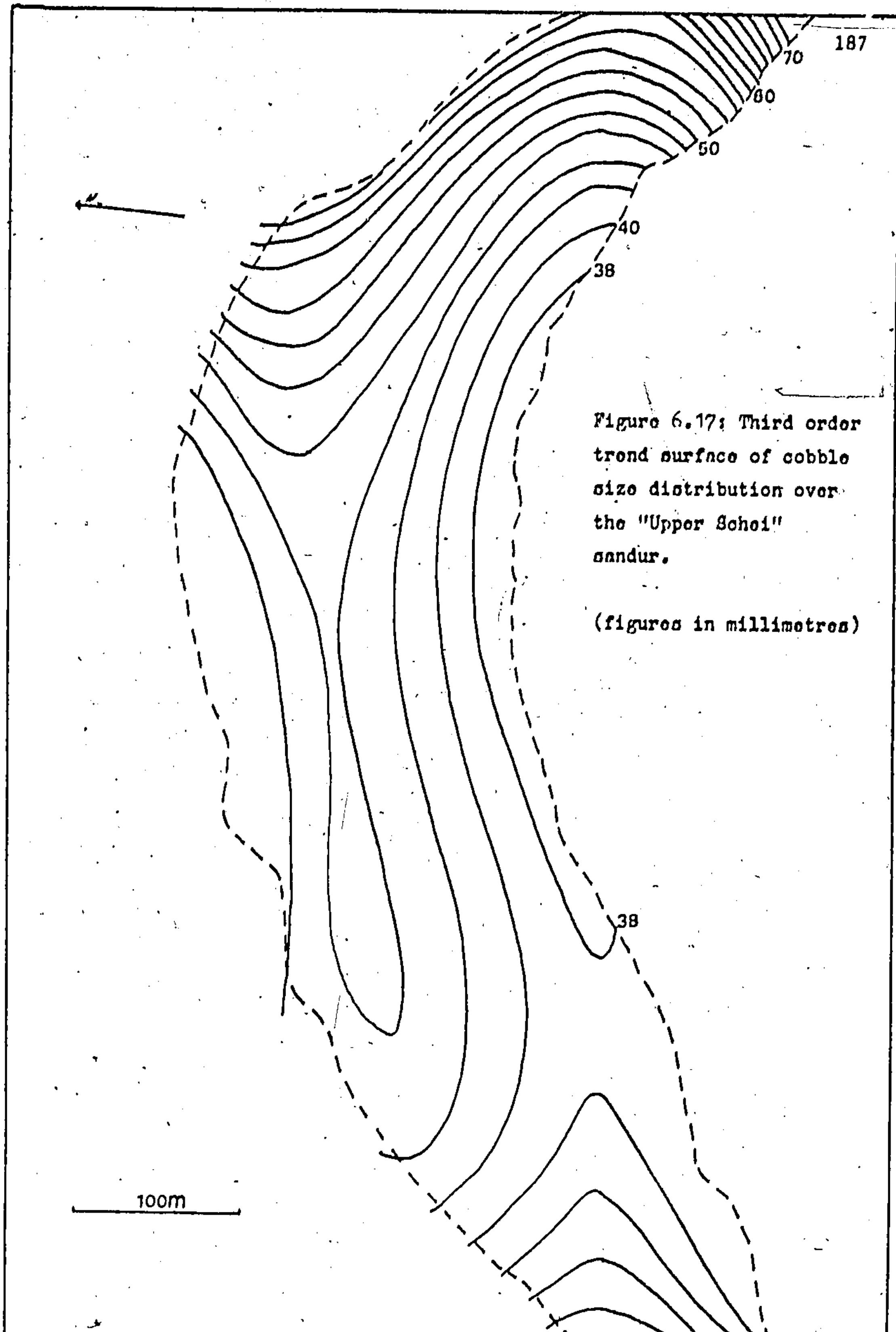


Figure 6.17: Third order trend surface of cobble size distribution over the "Upper Schei" sandur.

(figures in millimetres)

Analysis of residuals is normally accomplished by isarithmic mapping, but this approach would have little meaning in the present instance as isarithms plotted in the between-transect areas would have no significance. As an alternative, the mean first - second - and third-order residuals for each depositional "environment" have been tabulated for each of the five transects (table 6.1). A number of trends emerge:

1. C_2 samples almost invariably display negative residuals, whereas C_1 samples have consistently positive residuals. This implies that main channel deposits are coarser than those contained in main channel bars.
2. The relationship between H_1 and H_2 samples is apparently similar, although insufficient data is available to confirm this.
3. No systematic relationship is evident between F_1 and F_2 samples.
4. The absolute residual values on classes C_1, C_2 and H_1 show a general downsandur decrease, indicating a decrease in variation between classes. This may imply downsandur increase in the overall degree of sorting. No downsandur decrease in absolute residual values is evident for classes F_1 and F_2.

The reality of these trends is confirmed by plots of $\overline{B}$ against elevation above sea-level (figure 6.18) and against distance measured along the curve of the sandur from its apex (figure 6.19). Both diagrams indicate a downsandur decrease in overall clast size variation, possibly reflecting increased overall sorting, and the generally coarser nature of

TABLE 6.1 MEAN 1st, 2nd and 3rd ORDER CLAST SIZE RESIDUALS FOR EACH "DEPOSITIONAL ENVIRONMENT" TABULATED FOR FIVE CROSS SANDUR TRANSECTS

DOWNSANDUR DIRECTION →

	Transect 1			Transect 2			Transect 3			Transect 4			Transect 5		
Order of Surface	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3
Main Channel C_1	+97	+97	+96				+58	+ 55	+62	0	+10	+6	0	+4	+9
	n = 1						n = 1			n = 1			n = 2		
Main Channel Bars C_2	−20	−20	−20	−25	−18	−15	−2	0	+1	−5	−5	−7	−3	−2	−2
	n = 1			n = 2			n = 3			n = 6			n = 7		
Flood Channels F_1	−5	−5	−9	−16	−7	−10	+4	+5	+1	+24	+17	+11			
	n = 3			n = 3			n = 3			n = 1					
Flood Channels Bars F_2	−15	−18		−6	+4	+4	−12	−14	−9	+12	+6	−1			
	n = 4			n = 4			n = 3			n = 2					
High Flood Channels H_1	+79	+69	+58	+32	+38	+29	+28	+24	+23						
	n = 1			n = 1			n = 1								
High Flood Channel Bars H_2							−9	−6	−8						
							n = 3								

channel deposits vis à vis bar deposits. Regression of all bar samples against distance from the sandur apex yields

$$\bar{H} = 88.37\, e^{-0.00111D} \qquad (6.9)$$

$$(r = -0.788)$$

A steeper regression curve

$$\bar{H} = 110.17\, e^{-0.00138D} \qquad (6.10)$$

$$(r = -0.867)$$

is obtained for channel samples. The higher coefficient and exponent of e in equation (6.10) indicates that channel deposits display more pronounced downsandur fining. The elimimination of all deposits except types C_1 and C_2 at the distal end of the sandur (figures 6.18 and 6.19) is one apparent reason for the downsandur decrease in inter-sample variation.

Three general conclusions emerge from this analysis :

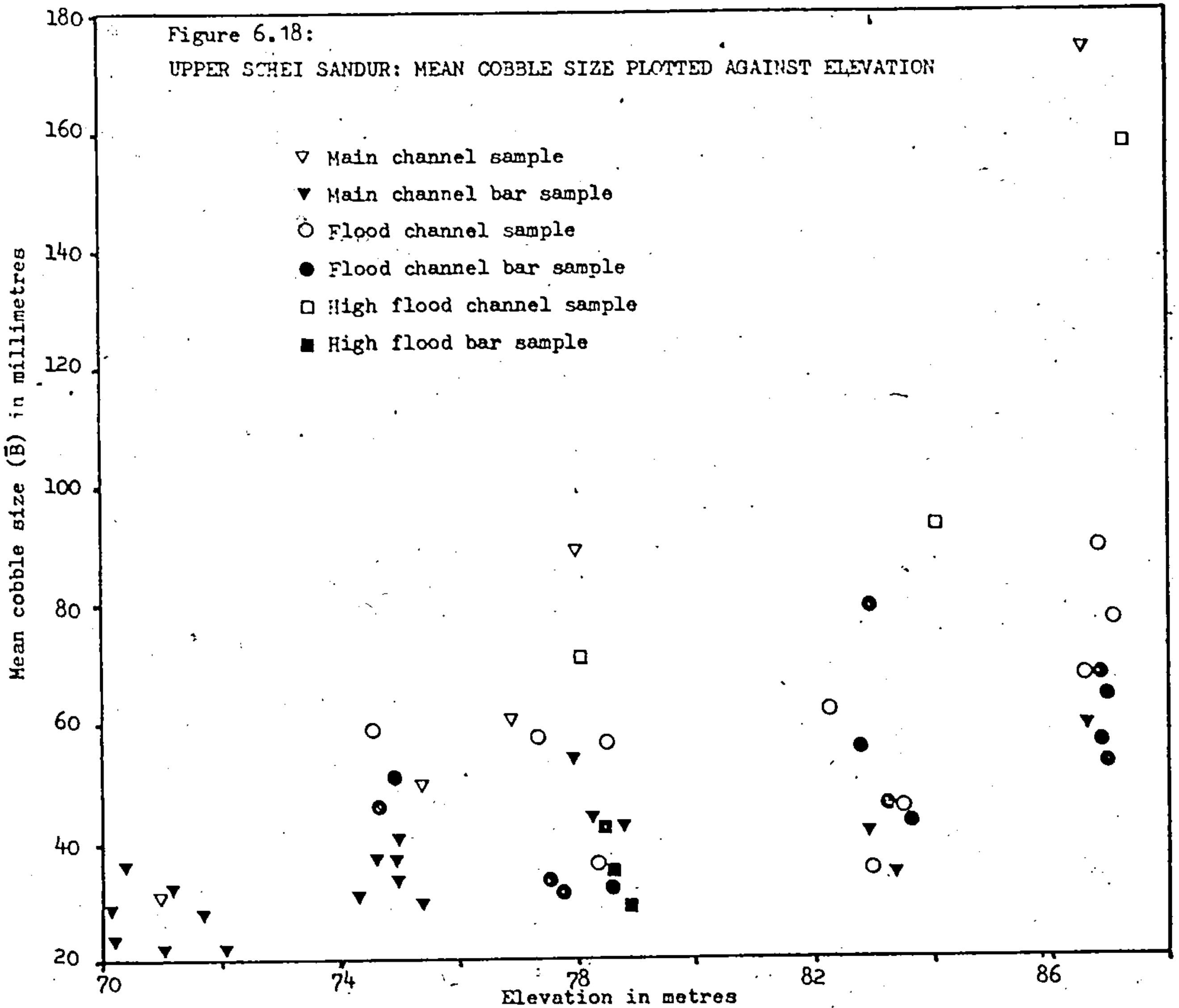
1. Coarse clast sizes exhibit a general downsandur decrease, but less than 40% of total cobble size variation is explained by linear or quadratic trend surfaces of downsandur clast size decrease. Tabulation of residuals indicates that a significant part of the remaining variation is explained by differences in clast size under differing depositional circumstances irrespective of distance from the sandur apex.

2. In general, channel deposits are coarser than bar deposits. This tendency is marked for main channel and "high flood" deposits, but is insignificant for flood deposits.

3. Variation from the mean cobble size decreases downsandur. This may reflect improved sorting or the elimination of all depositional classes

Figure 6.18:

UPPER SCHEI SANDUR: MEAN COBBLE SIZE PLOTTED AGAINST ELEVATION



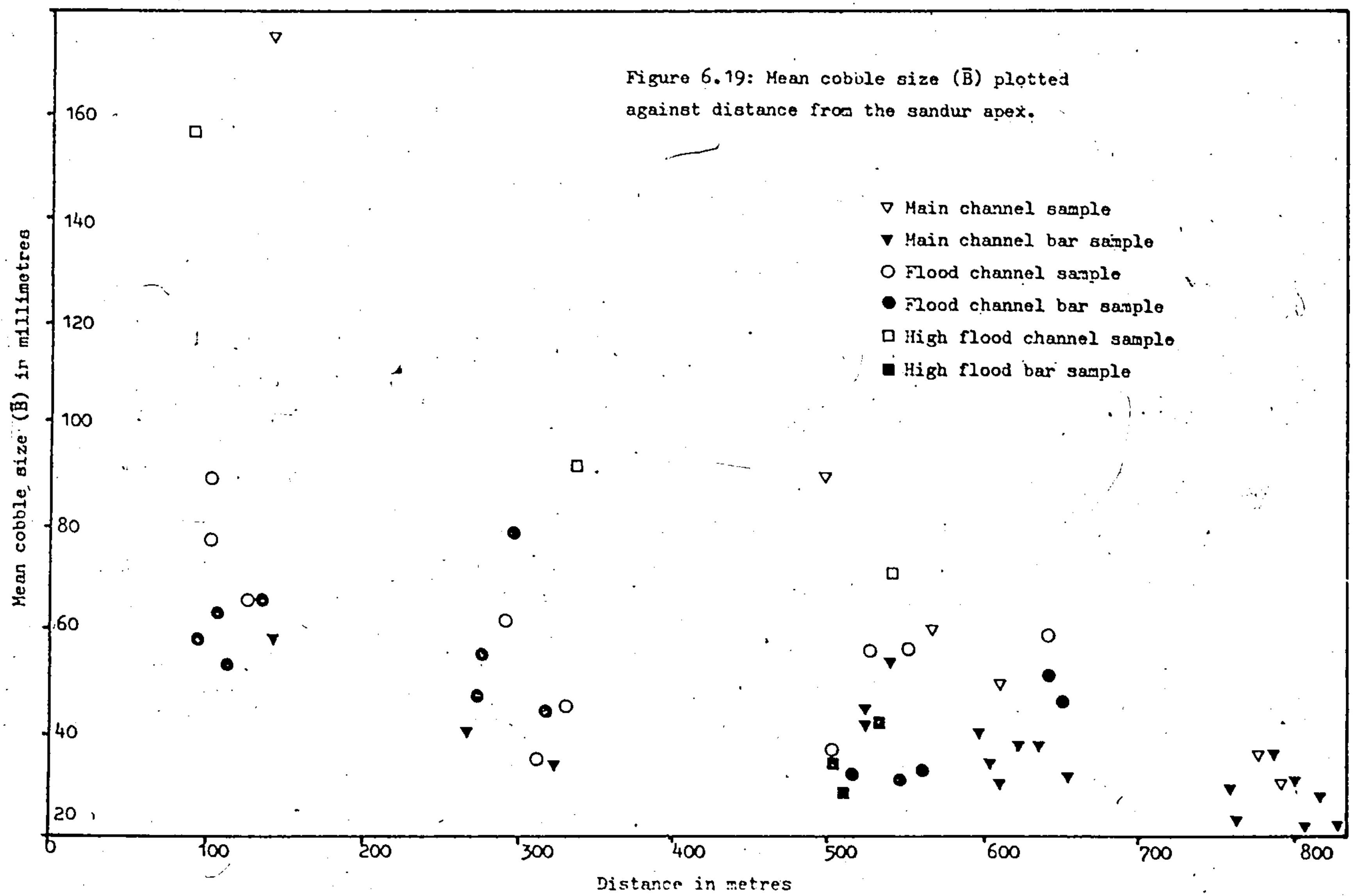


Figure 6.19: Mean cobble size ($\bar{B}$) plotted against distance from the sandur apex.

except main channel types.

6.2.5 Discussion

Downstream decrease of cobble size is a well-documented feature of coarse clastic alluvial deposits, and appears to be prominent on sandar (Church, 1972; Gustavson, 1974). McDonald and Banerjee (1971) relate downsandur decrease in clast size to

> "... the cumulative effect of selective sorting processes resulting from local variations in such hydraulic factors as water velocity and depth, and their relationship to stream competence ..."
>
> (p 1288)

In particular, decline in competence may be related to a parabolic decrease in slope (Church, 1972) and to decreasing depth as water becomes progressively distributed over the sandur surface (Krigstrom, 1962). However the results of the present study indicate that a simple "downsandur fining" model only offers a partial explanation of the total observed cobble-size variation, and that differing depositional circumstances, irrespective of distance downsandur, also account for a substantial part of total variation. In simple terms, cross-sandur cobble-size variations may equal or exceed downsandur variations.

Two sources of cross-sandur cobble-size variation have been identified : differences in the coarseness of bar and channel deposits; and differences in mean cobble size between main channel and "high flood" deposits on the one hand and flood deposits on the other. A difference in the coarseness of bar and channel deposits is to be expected as a result of Smith's (1974) findings, that individual bars fine downstream but adjacent channels do not. This pattern he attributed to the lateral migration of bars and deposition of smaller clasts during waning discharge. For the entire downsandur length of a main channel, this pattern will be reflected in a systematic decrease in mean cobble size in the channel itself (the result of decreasing competence, which in turn reflects decreasing slope and increased downsandur distribution of flow) whereas no such systematic decrease will be evident from sampling several discrete main channel bars. The lack of a distinct downsandur fining trend amongst flood channel deposits probably reflects relatively uniform competence amongst the streams occupying such channels during periods of high discharge, in contrast to the declining downsandur competence of main and "high flood" channels. The abundance of sand and silt veneers in flood channels suggests deposition under conditions of rapidly declining discharge.

The three-zone subdivision of dalsandar proposed by Krigström (1962) acquires new meaning in terms of the three channel types classified above. In the proximal zone the "high flood" surface is best developed, and main channels are incised. Flood channels predominate in the intermediate zone. As the distal zone is neared, "high flood" areas disappear and the main channel becomes widely distributed. The successive downsandur elimination of "high flood" and flood deposits offers an explanation for the marked decrease in inter-sample variation at the distal end of the sandur.

In summary, this study constitutes a first attempt at describing and explaining the size distribution of coarse clastic sediments over an entire sandur surface in terms of different depositional "environments". The results indicate that cross-sandur variations may exceed downsandur variations, and that much of the former may be explained by consideration of differences between channel and bar deposition in channels active at extremely high flow, high flow, and low flow. The successive downsandur dominance of each of these three types may provide the physical basis for the three-part sandur subdivision first proposed by Krigström (1962). An observed downsandur decrease in cobble size variation may reflect progressive elimination of all but low flow channels, and possibly implies improved downsandur sorting. The utility of such results in the study of ancient or Pleistocene glaciofluvial sediments requires no emphasis, but first they require refinement and confirmation.

6.2.6 Studies of relict sandar : introduction

The principal aim of the following section is the comparison of cobble size distribution over the surface of two relict sandur fragments (I and II on figure 6.9) with that already described for the active sandur surface. Such comparison may reveal something of the conditions under which the relict surfaces were deposited. The development of earth hummocks and vegetation precluded similar investigations on the surfaces of the large relict sandar of the Innuitian retreat phase (chapter 3), but some sedimentological observations on one of these features are presented in section 6.2.8.

Beyond the margin of the active "Upper Schei Sandur" are three relict sandur fragments which lie appreciably above the present active surface yet are clearly incised into the lateglacial relict sandar. In chapter 3 these fragments were assigned a neoglacial age. This is borne out by their fresh unvegetated appearance.

As a prelude to the study of clast size distribution on these features, if is necessary to ascertain their contemporaneity. This was done by levelling spot heights at the terrace margins, then plotting these with reference to their positions relative to the present main channel. The result (figure 6.20) is a profile showing conformity to the power curve

$$H = 0.00184D^{1.399} \qquad (6.11)$$

$$(r = 0.999)$$

The equation for the long profile of the present main channel is

$$H = 0.00049D^{1.586} \qquad (6.1)$$

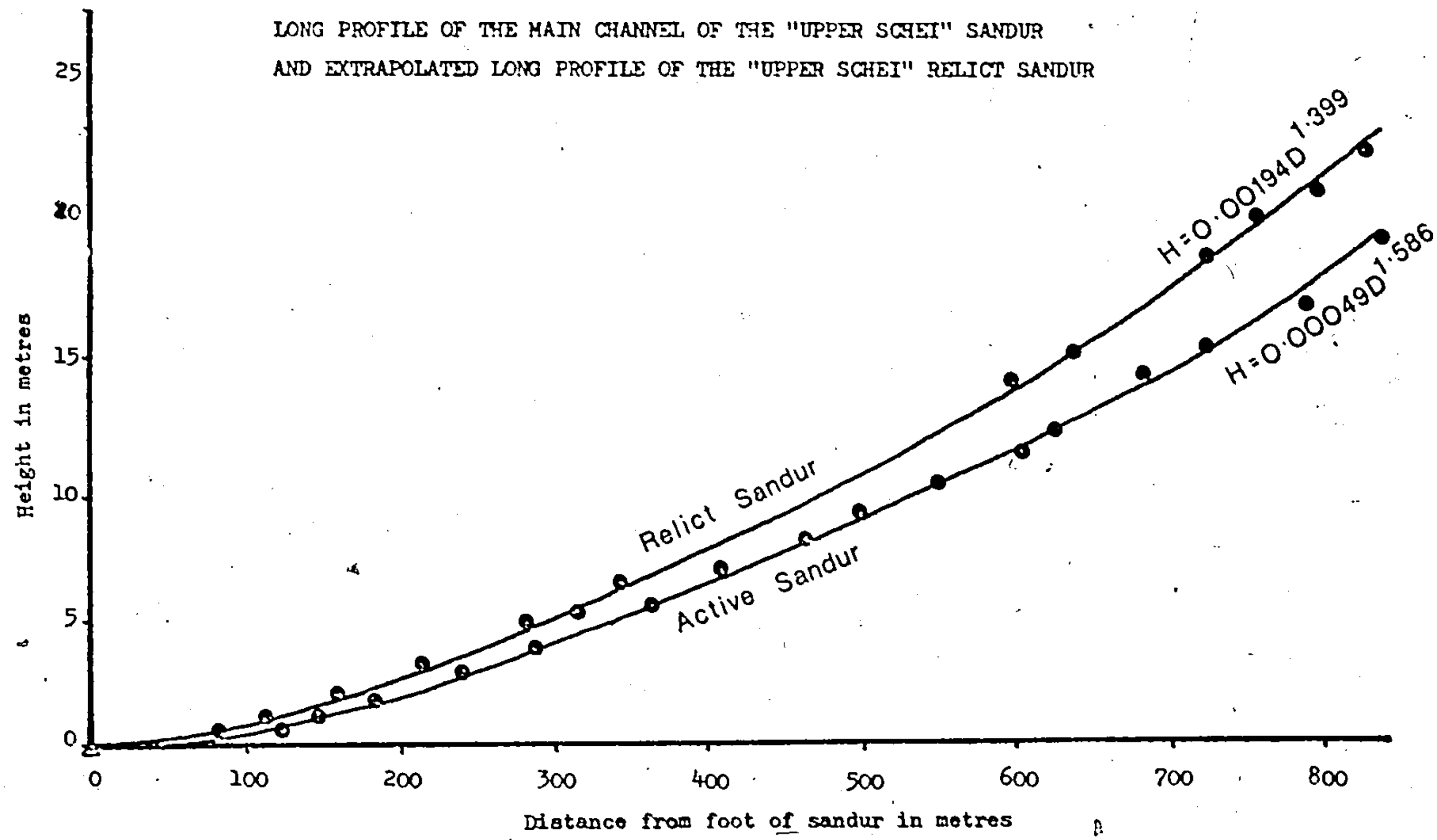
$$(r = 0.996)$$

The higher coefficient of D in equation (6.11) indicates that the relict sandur was steeper than its present-day equivalent. This is confirmed by elevation trend surfaces generated independently for two sandur fragments and for the active sandur. The second order surfaces (figure 6.21) show closer trend lines on the relict sandur than on the active surface, except in the extreme distal area. The second-order elevation explanation level exceeded 99% for all these surfaces.

Little downcutting of the "Schei Gorge" base level has occurred since the deposition of the neoglacial sandur, which is graded to almost the same level as the present feature. It therefore seems likely that changes in hydraulic regime rather than changes in base level account for

Figure 6.20:

LONG PROFILE OF THE MAIN CHANNEL OF THE "UPPER SCHEI" SANDUR
AND EXTRAPOLATED LONG PROFILE OF THE "UPPER SCHEI" RELICT SANDUR



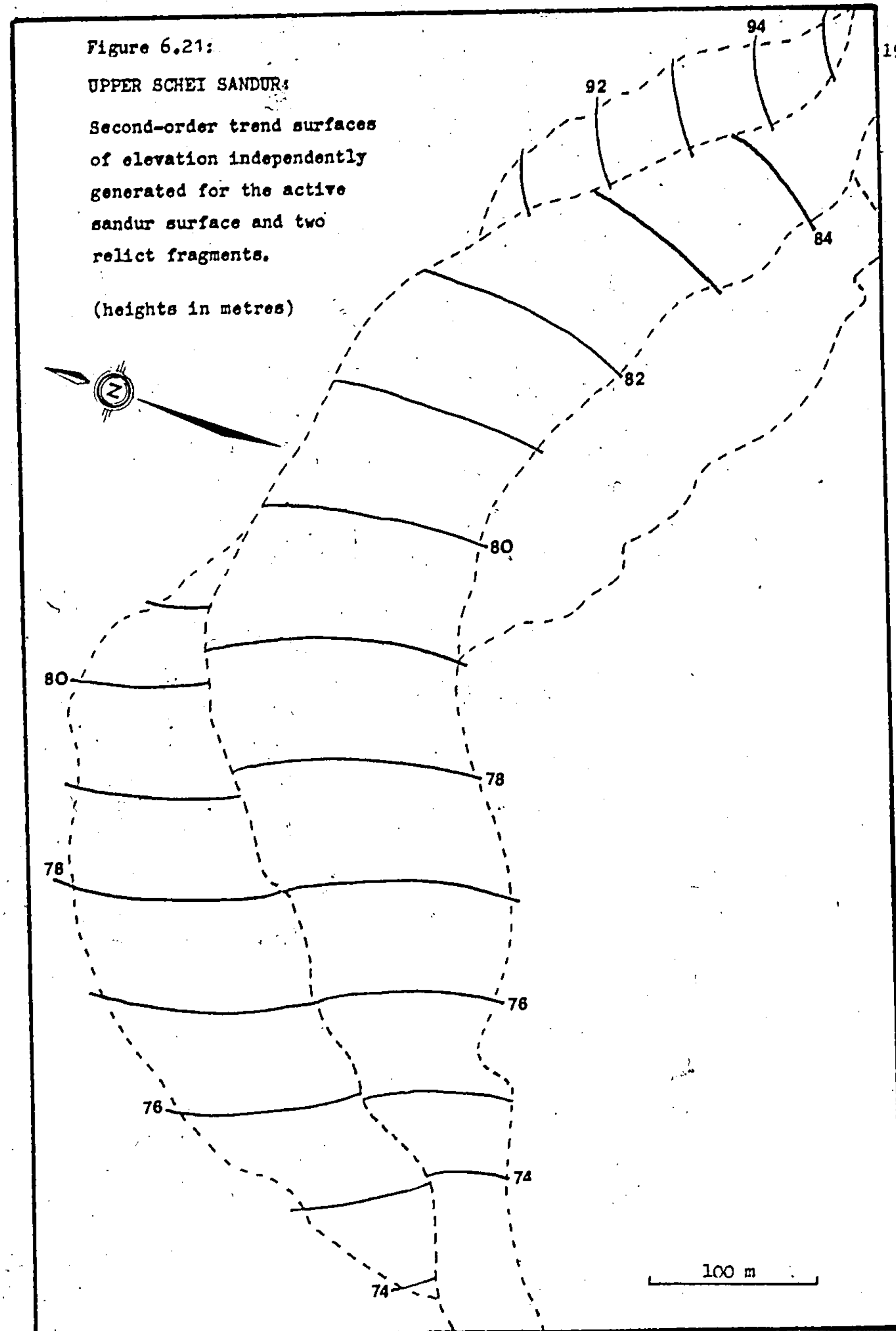


Figure 6.21:

UPPER SCHEI SANDUR:

Second-order trend surfaces of elevation independently generated for the active sandur surface and two relict fragments.

(heights in metres)

the present sandur being less steep than its neoglacial predecessor.

6.2.7 Results and analysis

The field and analytical procedures employed in the study of cobble size distribution over the surface of the relict sandur fragments were similar to those used on the active sandur. Three transects with 15 sampling sites were established for fragment I (figure 6.13), and a further three with 16 sampling sites, for fragment II. Bar and channel deposits were distinguished, and sampling was carried out using the photographic technique described in appendix 2. First - and second - order trend surfaces of clast size variation were independently generated for each fragment (figures 6.22 and 6.23).

The first-order surface for the upsandur fragment

$$\overline{B} = 76.49 - 0.511\,X + 1.484\,Y \qquad (6.12)$$

$$(r = 0.789)$$

explains 62.31% of clast size variation, but the trend is to the sandur margin rather than downsandur. This is less apparent for the second-order surface

$$\overline{B} = 118.23 - 1.367\,X + 0.424\,Y + 0.503 \times 10^{-2}\,X^2 - 0.22 \times 10^{-2}\,XY + 0.488 \times 10^{-2}\,Y^2 \qquad (6.13)$$

$$(r = 0.840)$$

which explains 70.55% of the variation. The first-order surface for fragment II

Figure 6.22: First order trend surfaces of cobble size distribution for two relict sandur fragments.

(figures in millimetres)

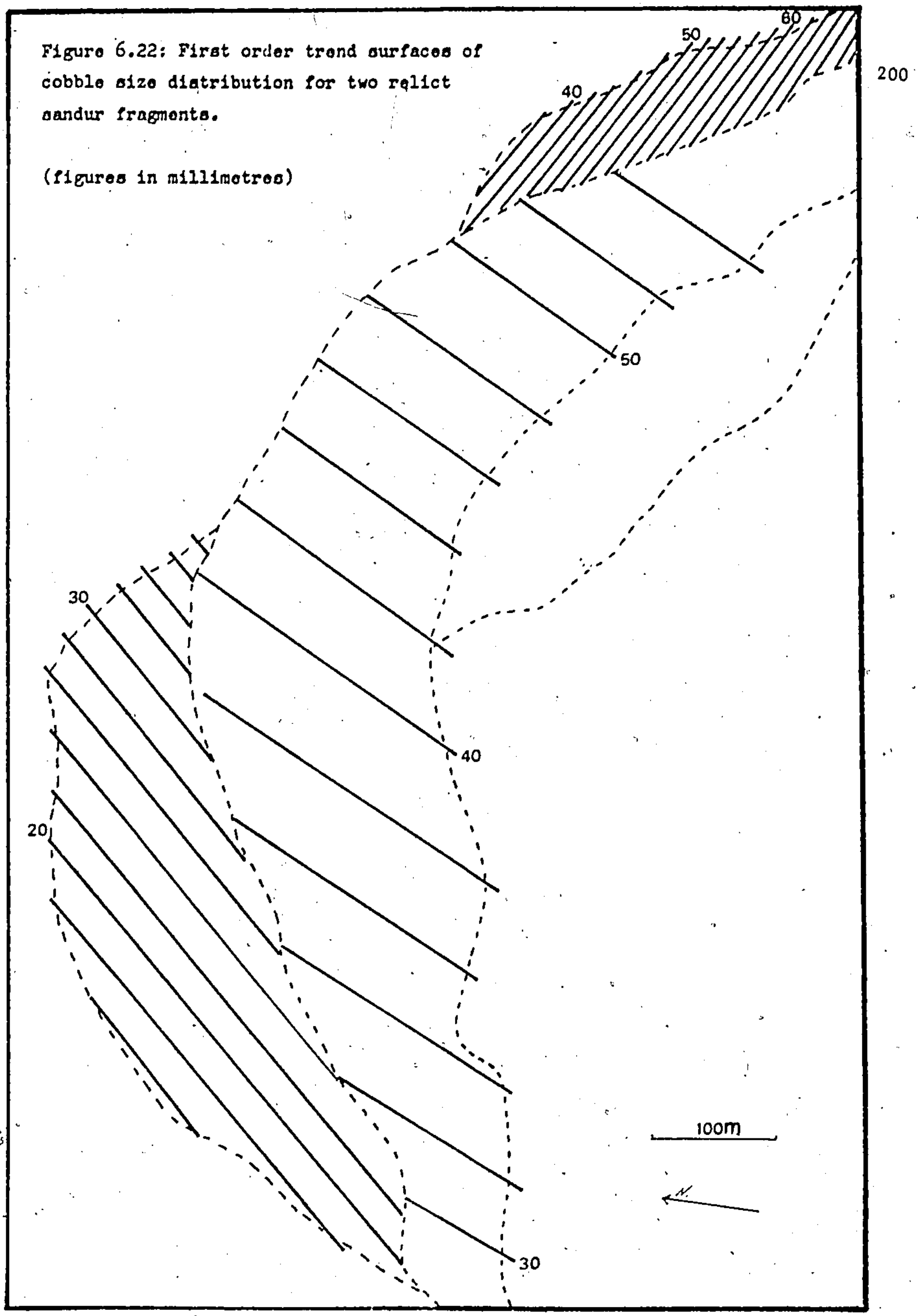
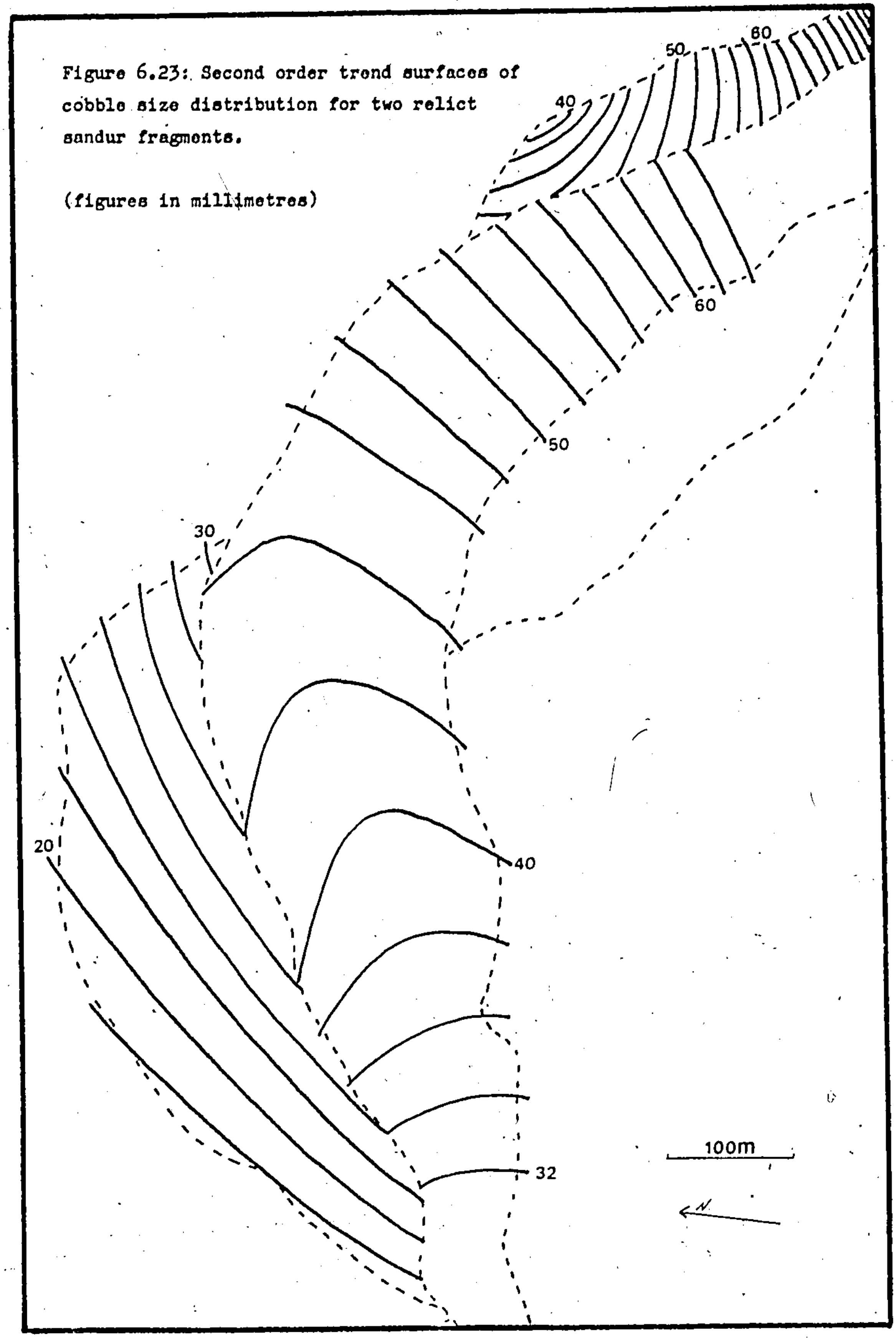


Figure 6.23: Second order trend surfaces of cobble size distribution for two relict sandur fragments.

(figures in millimetres)



$$\overline{B} = 24.03 - 0.037\, X + 0.096\, Y \qquad (6.14)$$

$$(r = 0.844)$$

explains 71.21% of clast-size variation and shows a marked size decrease away from the sandur apex. The second-order surface, with 77.13% explanation, shows this trend more clearly. It has the equation

$$\overline{B} = 24.378 - 0.023\, X + 0.104\, Y - 0.175 \times 10^{-3}\, X^2$$
$$+ 0.561 \times 10^{-3}\, XY - 0.466 \times 10^{-3}\, Y^2 . \qquad (6.15)$$

The more pronounced downsandur decrease in mean clast size of the relict fragments in comparison with the active surface (figures 6.22 and 6.23) is presumably related to the difference in slope. Much larger ($\overline{B}$ = 600 mm) boulders are found at the apex of the relict sandur than are evident on the active surface. This indicates deposition during a period when competence (hence discharge) exceeded that of the present day. This would be expected at a time of glacial retreat.

Tabulation of the mean first - and second-order residuals of bar and channel deposits for each of the six fragment transects (table 6.2) revealed trends similar to those found for the active sandur. Mean channel deposit residuals are characteristically positive, bar deposit residuals negative, and the absolute size of the mean residuals decreases downsandur. These results indicate that mean cobble size tends to be coarser in channels and that sorting probably improves downsandur.

The positive correlation between slope and cobble size suggested by the comparison of the relict and active sandur surfaces was further tested through sampling mean cobble size in the "Schei Gorge", which has an approximately linear long profile (figures 3.9 and 3.10). Samples

TABLE 6.2

MEAN FIRST- AND SECOND-ORDER CLAST SIZE RESIDUALS FOR SIX CROSS-FRAGMENT TRANSECTS

Downsandur →

Mean first-order residuals

Transect	1	2	3	4	5	6
Bar	-23 (n=3)	-23 (n=4)	+13 (n=3)	-2 (n=5)	0 (n=3)	-1 (n=4)
Channel	+58 (n=2)	-1 (n=2)	+24 (n=1)	+2 (n=2)	+3 (n=2)	

Mean second-order residuals

Transect	1	2	3	4	5	6
Bar	-34 (n=3)	-4 (n=4)	-1 (n=3)	-1 (n=5)	-1 (n=3)	-1 (n=4)
Channel	+51 (n=2	+6 (n=2)	-2 (n=1)	+3 (n=2)	+1 (n=2)	

were measured both from active channel and T_{11} terrace deposits, the T_{11} terrace being correlative with the "Upper Schei" sandur fragments. Although only eleven samples were measured, the results (table 6.3) suggest a lack of downstream fining in the absence of a decrease in gradient. Again, channel deposits were consistently coarser than bar deposits.

In summary, the results of comparing the "Upper Schei" relict sandur fragments with the active surface emphasise the close relationship between sandur gradient and cobble-size distribution : the steeper the slope, the more rapid the downstream fining of material. The coarseness of the deposits at the head of the relict feature is attributable to high water discharge during ice retreat from the neoglacial maximum. As on the active sandur, channel deposits were consistently coarser than bar deposits on the same cross-sandur transect.

6.2.7 Sections through the "Endrick" relict sandur

As the digging of pits in active sandar is virtually impossible, the examination of sections in palaeo-outwash deposits provides the main source of information on sandur stratigraphy (Jewtuchowicz, 1953; McDonald and Banerjee, 1972). Such investigations have shown that the predominant (gravel) facies display massive or crude horizontal bedding, beds being distinguishable through differences in water-retaining qualities (Church, 1972). Sediments finer than -3ϕ (8 mm) may show more complex structures, such as cross- and parallel-stratification, and ripple laminae (McDonald and Banerjee, 1972).

TABLE 6.3

MEAN COBBLE SIZES IN THE "SCHEI GORGE" ($\overline{B}$ *in mm*)

	Active Channel		T_{11} Terrace	
	Channel	Bar	Channel	Bar
Upstream	76.75 85.72	44.25	84.04	66.23
Mid-gorge		32.90 43.24	72.15	66.93
Downstream	75.62 105.10			

Terrace scarps of the lateglacial relict sandar in the "Schei River" basin are normally obscured by slumped material, but at the distal end of the relict "Endrick" sandur rapid erosion through the development of thermo-erosional niches has exposed several fresh stratigraphic sections (figure 6.24). Three predominant types of deposit were exposed :

1. coarse unsorted gravels in a sandy matrix;
2. horizontal or near horizontal laminated sands, silts and clays, often peaty in texture and containing roots; and
3. Interdigitated sands and gravels.

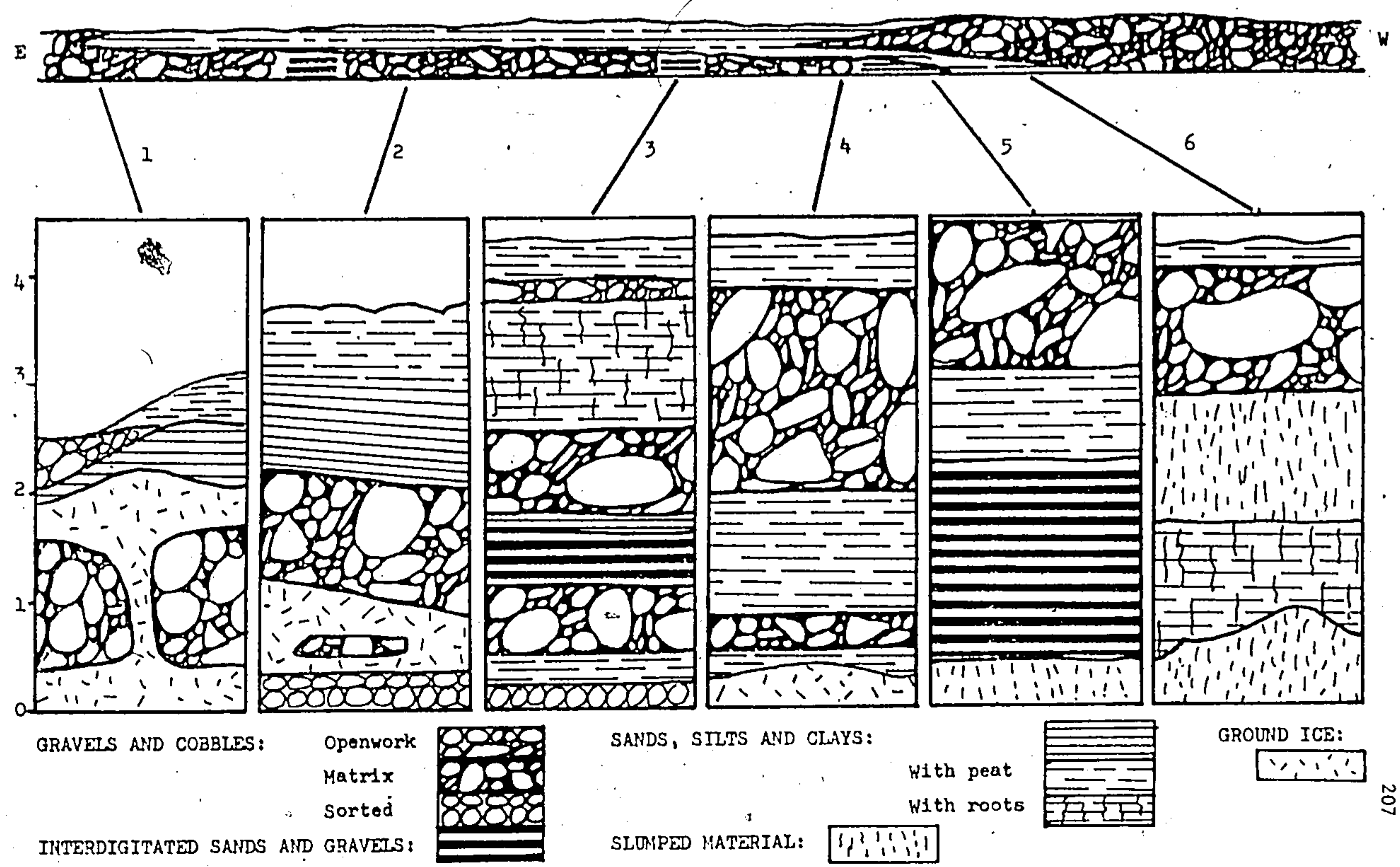
Types 1 and 2 extended uninterrupted over large areas of terrace scarp; type 3 existed in localised lenses.

Despite the complexity of the six sections depicted in figure 6.24, it was possible to resolve these into a simple overall pattern. The western end of the scarp is characterised throughout by massive deposits of coarse unsorted gravels in a sandy mattrix. This deposit tapers out eastwards between two strata of finer material, sands, silts and clays. These fine deposits are further underlain by more coarse gravels, which in places contain lenses of stratified sands and gravels.

The depositional sequence is interpreted as follows. During the main postglacial retreat of ice from the "Endrick" basin, a large sandur predominantly composed of coarse gravels aggraded above the valley floor. Locally, stratified sands and gravels, possibly representing bar deposits (Smith, 1974) were laid down. As the ice masses in the "Endrick" basin attenuated, however, discharge fell and the eastern two-thirds of the sandur were abandoned. The western third continued to

Figure 6.24:

STRATIGRAPHIC SECTIONS IN THE "ENDRICK" RELICT SANDUR



aggrade, however, causing a large shallow pond to develop to the east. In this pond, laminated deposits of sand, silt and clay accumulated, with consequent plant colonization and peat formation. With the post-glacial downcutting of the "Schei River", the "Endrick" finally abandoned the sandur surface and became incised along its western margin.

This interpretation is supported by the nature of the sandur surface. Above the western "gravel" area braided channel forms are well preserved, but on the surface of the fine "ponded" material no such braids are apparent (figure 3.6). It is concluded that ponding over substantial areas may accompany sandur aggradation, such ponding being analogous to the formation of sloughs and backswamps on the floodplains of lower latitudes. In the study basin, evidence of similar pondings was found along the western section of the "Lendal" relict sandur, indicating that damming by sandur aggradation was not unique to the "Endrick" basin.

CHAPTER 7

CONCLUSION

Geomorphologists have traditionally perceived the high arctic landscape as dominated by glacial and periglacial processes, and only recently has serious investigation commenced into the role of post-pleistocene fluvial processes. In the present study, a fourfold approach to the problem of evaluating the geomorphic role of arctic rivers was adopted, through investigations carried out in the "Schei River" basin in the summer of 1974. The form of the lateglacial landscape was interpreted, to allow the assessment of postglacial fluvial modification. The hydrology of the "Schei River" was studied in detail, then linked to measurements of sediment concentration to yield information on the nature and rate of sediment removal. Finally fluvial landforms were described, with particular attention to sedimentary characteristics of sandur deposits.

The study of postglacial landscape modification yielded interesting information on local lateglacial history as well as indicating substantial subsequent fluvial activity. Evidence was found for two glacial readvances, one of lateglacial (7,000 - 8,000 B.P.) age, the other possibly neoglacial. Comparison of the lateglacial and contemporary drainage networks revealed basic similarities, except where the neoglacial readvance had disrupted the earlier pattern. Postglacial modification has consisted primarily of vertical incision concomitant on rapid glacio-isostatic uplift in excess of 60 m. This has resulted in the widespread terracing of earlier alluvial deposits, and the cutting of near-vertical gorges in bedrock. At one point, the river has downcut 30 m through competent sandstones and siltstones in the c.7,000 years since deglaciation. However, as the rate of sea-level change is now minimal, it is unlikely that such rapid vertical erosion persists at the present time.

In 1974, flow on the "Schei River" commenced on June 22, and maximum diurnal discharge increased steadily to a peak of c.17 $m^3 s^{-1}$ at the height of the snowmelt freshet on July 12. Similar discharges were recorded during warm weather at the beginning of August, and high (> 7 $m^3 s^{-1}$) discharges were measured as late as August 18, when recording ceased. Total discharge during the period June 22 - August 18 was calculated to be c.19.5 x $10^6 m^3$, of which 55% was supplied by glacier melt, 35% by snowmelt and 10% by summer precipitation. The main nival tributary, draining 32.8% of the basin, supplied only 14.6% of total discharge; in contrast, a glacial tributary draining 10.3% of the basin area supplied 28.3% of total discharge. The dichotomy arose from the depletion of runoff inputs in the unglacierized portion of the basin after the termination of the snowmelt freshet. Runoff from glacial streams, however, responded to high radiative and temperature inputs throughout the entire runoff period.

In common with other streams nourished primarily by snow and ice melt, the "Schei River" evinced an oscillatory response to diurnal fluctuations in radiative and temperature inputs. The mean lag of peak diurnal discharge behind peak diurnal net radiation was 5.2 h. Time of diurnal peak discharge for the main glacial tributaries advanced progressively throughout the early part of the runoff season, due to the increasing efficiency of the supraglacial drainage network. In contrast the time of maximum diurnal discharge on the principal nival tributary was progressively retarded as sources of meltwater retreated to higher parts of the ice-free basin. The magnitude of river response to rainfall on unglacierized terrain was apparently dependent on antecedent moisture conditions in the active layer.

The main tributary of the "Schei River" flows along and under the glacier margin for much of its course. The ponding of water behind collapsed ice at the glacier margin represented a considerable hydrologic hazard, as sudden floods of unpredictable timing and magnitude resulted from the breaching of such ice dams. One such flood was responsible for the maximum recorded discharge of c.24 m^3s^{-1}.

The concentration of dissolved solids measured in the "Schei River" in 1974 ranged from 15 to 97 ppm, with a mean value of 29 ppm, composed almost entirely of $CaCO_3$ and $MgCO_3$. The range of concentration of suspended inorganics was 5 - 2,204 mgl^{-1}. Total dissolved load over the period June 22 to August 18 was calculated to be 1517 metric tonnes (= 16.6 t km^{-2} or 25.8 t km^{-2} per unit ice-free area). Total suspended load was 6.7 times greater, being 10,251 metric tonnes (= 112.4 t km^{-2} or 174 t km^{-2} per unit ice free area). Dissolved concentration exhibited an inverse nonlinear relationship with discharge, such that discharge increase was more rapid than concentration decrease, but the relationship between discharge and concentration increase for suspended sediment was near-linear.

A very high proportion of total load was supplied by the glacial tributaries, largely reflecting the higher discharge of these streams vis à vis the main nival contributant, especially after the termination of the snowmelt freshet. Much of the dissolved and suspended load carried by the glacial streams was provided by meltwater erosion at the glacier snout.

The most distinctive erosional forms in the "Schei River" basin are deep gorges cut in competent bedrock, the formation of which is attributable to erosion by lateglacial meltwaters at a time of rapid isostatic recovery. The dominant small-scale erosional landform is the thermo-erosional niche, development of which was responsible for terrace erosion of 4.5 m y^{-1} ine one area. Sandar constitute the most important depositional features in the basin. Analysis of cobble-size distribution over the surface of a small sandur revealed that a simple downsandur-fining model is inadequate, and that depositional circumstances are important in accounting for clast-size distribution. Channel deposits were almost invariably coarser than nearby bar deposits, and homogeneity of cobble size increased downsandur. Similar trends emerged from analysis of cobble size distribution on steeper relict sandur fragments. Examination of sections in relict sandar revealed that large pondings may develop during sandar aggradation.

Although the approaches adopted in this study were somewhat selective, a common leitmotif is discernible in the results obtained. Post-glacial fluvial incision into bedrock, a hydrologic regime involving rapid fluctuations in discharge, the erosion of large quantities of material at the glacier margin, the transportation of large amounts of sediments in solution and suspension and the deposition of coarse clastic debris in widespread sandur surfaces all evidence the fundamental role of rivers in reshaping the postglacial arctic landscape. To describe rivers as the most important high-latitude geomorphic agents would be an overgeneralization; but of all the contemporary processes operative in the arctic, those associated with river action are certainly the most dynamic.

APPENDIX I

RATING CURVES OBTAINED FOR THE SCHEI RIVER GAUGING SITES

The rating curves for the Schei River gauging stations (Figure A.1.1) were established by obtaining a series of stage-discharge relationships for each site. This was done in two ways. Low values of discharge (normally less than 10 $m^3\ s^{-1}$) were measured using rotating-cup current meters. This method utilises the relationship

$$Q = \nu A \qquad (A.1.1)$$

where Q is mean stream discharge in $m^3\ s^{-1}$, ν is mean velocity in m s^{-1}, and A is submerged cross-sectional area of the gauging section in m^2. The "six-tenths" methods was employed (Church and Kellerhals, 1970), whereby a number of velocity readings are made across the stream at a depth which is 0.6 of total depth (z) for that point, as measured from the surface. Each velocity (ν) and depth (z) reading is assumed to be representative of mean velocity and depth over a certain width (w) of stream cross-section. Each value of w is bounded by the midpoints between one velocity-depth measurement point and adjacent measurement points, or by the midpoint with one adjacent point and the stream bank in the case of the first and last measurements across a section. When n velocity-depth measurements are made, then

$$Q = \sum_{i=1}^{n} w_i z_i \nu_i \qquad (A.1.2)$$

Velocity was measured over 60 seconds. The number of measurements made per single stage value normally ranged from 12 to 20, although as few as 8 measurements were made during periods of low stage on Endrick Creek and Lendal Creek. Church and Kellerhals (1970) show that discharge values obtained with 8 readings are liable to an outside error of ± 15%; with 12 readings, ± 9%; with 20 readings, ± 7%. With more than 20 measurements, decrease in outside error is minimal.

The stage-discharge relation for high values of discharge was obtained using slug injection dilution techniques employing 20% Rhodamine WT dye, in the manner described by Church and Kellerhals (1970). Four high-discharge points on the Upper Schei River stage-discharge rating curve were obtained using an integrated sampling technique: 50 ml of dye were injected 900 m upstream of the sampling point, and 26 2 dram samples were collected at two minute intervals from the time of injection. The samples were mixed, the integrated simple concentration was measured using a fluorometer, and discharge was calculated using the equation

$$Q = \frac{c_1 S}{c_2 t} \qquad (A.1.3)$$

where S is the volume of dye injected in l, c_1 is the initial concentration of dye (20%), c_2 is the mean concentration value of the sample, and t is the sampling period.

A non-integrated sampling technique was used to obtain some of the points on the Schei River rating curves. In this case dye was slug-injected 400 m above the sampling point and samples were collected in 2 dram vials at intervals of 15–300 seconds for 30 minutes after the time of injection. Each sample was analyzed individually on a fluorometer, and a time versus dye concentration graph was drawn from the results. Discharge was obtained using the equation

$$Q = \frac{Sc_1}{Aab} \qquad (A.1.4)$$

where S and c_1 are as for equation A.1.3 and A is the area under the time concentration surve, given by

$$A = \int_{t_i}^{t_f} c(t)dt \qquad (A.1.5)$$

where t_i is initial sampling time, t_f is final sampling time and c is concentration. The terms a and b in equation A.1.4 are corrections for graph scales. Sources of error for these techniques have been discussed by Church and Kellerhals (1970, pp. 50-55, 68-69, 72).

The derived rating equations, together with the number of rating points and the correlation coefficients for each equation have been given in Table 4.2.

REFERENCE

Church, M. and Kellerhals, R., 1970. Stream Gauging Techniques for remote areas using portable equipment. Inland Waters Branch, Dept. Energy, Mines and Resources, Tech. Bull., 25, 89p.

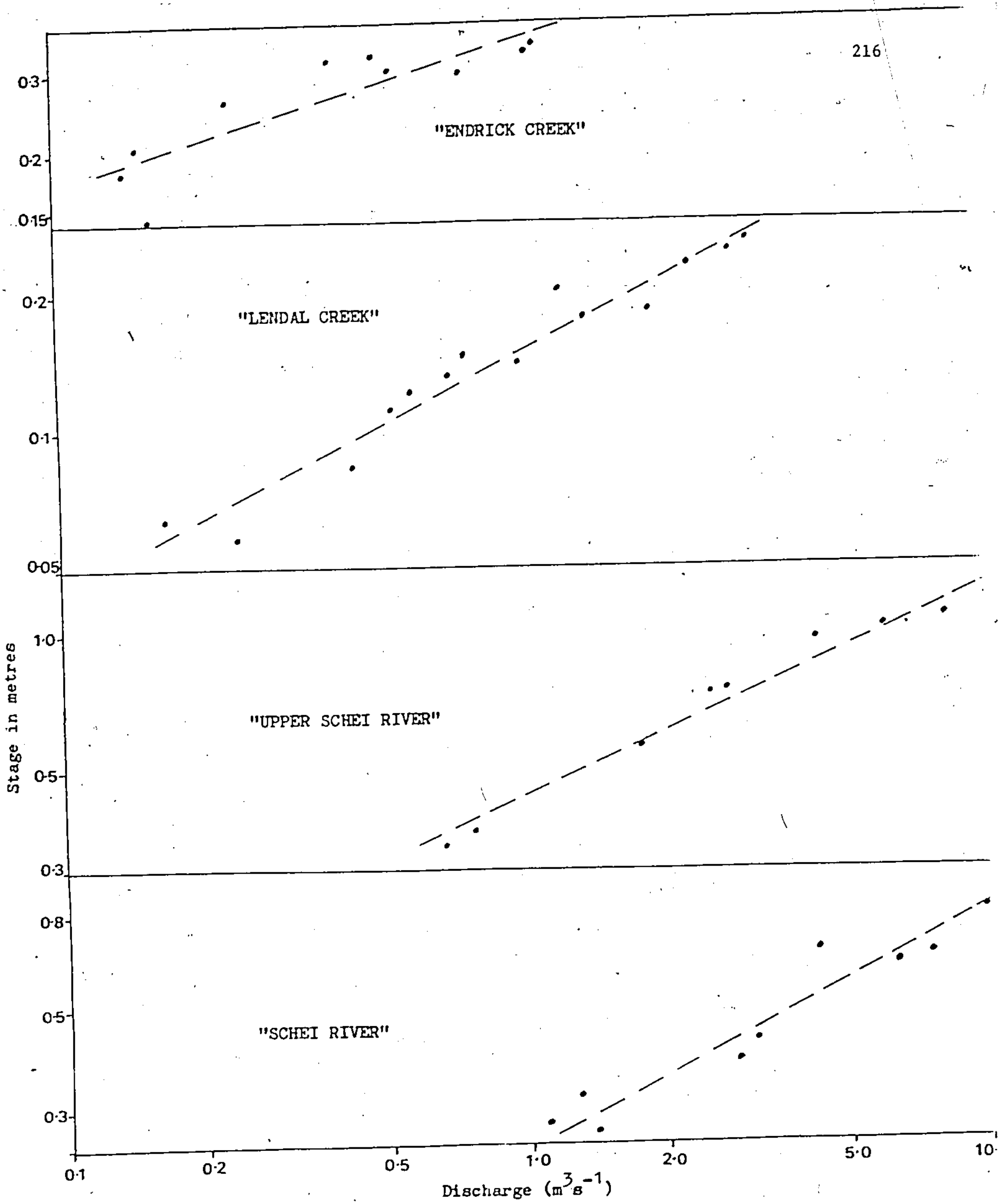


Figure A.1.1: Stage-discharge rating curves for the "Schei River" and tributaries, 1974.

APPENDIX 2

COARSE SEDIMENT SAMPLING PROCEDURE

The established sampling procedure for obtaining mean cobble size data for coarse clastic deposits involves the random selection of a suitable number of clast samples at predetermined sampling sites and measurement of the principal axes of each sample (Wolman, 1954; Leopold, 1970). These measurements are then employed to compute representative mean values for each site. This technique is however excessively time-consuming in the field, and for the present study a more rapid photographic procedure devised by Church (1967) was employed to obtain mean cobble size data for the "Upper Schei" sandur surface.

Church's method involves the random placing of a sampling quadrat (in this case 1.0 m square) on the sandur surface at predetermined sampling points and photographing this grid from (near) vertically overhead. In addition, the intermediate or B-axes of cobbles in the grid at a number of representative "control" samples are measured. For the purpose of analysis, coarse clastic material is considered to have B-axes greater than -3 ϕ (8 mm).

Analysis of this primary data to obtain mean cobble size for each site is based on the assumption that the number of cobbles counted in each quadrat bears some systematic relationship to mean cobble size. This relationship is established using the "control" data (table *A.2.1*) to create a "control" curve (figure*A.2.1*)for mean cobble size against number of cobbles. The establishment of a control curve is however complicated by a number of factors. Many quadrats contain zones of surficial material of less than -3 ϕ, and an area adjustment must be made to account for such zones. This is done by measuring total quadrat area (A_t) and area of cobble cover (A_c) by planimeter, then adjusting the counted number (n) of cobbles, such that

$$n_{adj} = \frac{n\ A_t}{A_c} \qquad (A.2.1)$$

Further difficulty arises from the fact that the number of pebbles counted in the field is unlikely to equal the number measured on a photograph, due to counting clasts which do not appear on the photographs, poor photographic resolution and counting errors. Hence the control curve measured in the field, expressed as the regression

$$Log\ \overline{B} = 3.312\ n_{adj}^{-0.12} \qquad (A.2.2)$$

$$(r = -0.975)$$

differs from that for the same samples measured from photographs, expressed by

$$Log\ \overline{B} = 6.621\ n_{adj}^{-0.28} \qquad (A.2.3)$$

$$(r = -0.987)$$

Mean cobble size $(\overline{B})$ values were obtained for all of the "non-control" samples by counting the number of pebbles (n) in the photographed quadrats, measuring A_T and A_c for each photograph by planimeter, calculating n_{adj} by equation (A.2.1) then substituting this value into equation (A.2.3).

Table A.2.1

Control data for cobble size analysis

Quadrat Number	Area Adjustment	Measured n	Counted n	Measured n_{adj}	Counted n_{adj}	Measured mean($\bar{B}$) mm.	Calculated mean($\bar{B}$) mm.
1.1	1.467	34	33	49.9	48.4	157.48	162.97
1.9	1.067	104	251	110.9	267.8	58.24	59.06
2.4	1.593	188	236	144.9	375.9	44.46	44.29
3.3	1.658	61	145	101.1	240.4	57.70	60.77
3.15	1.000	132	264	132.0	264.0	42.55	44.66
4.7	1.294	127	388	164.3	502.0	36.70	38.63
5.6	1.000	228	773	228.0	773.0	30.35	28.04
R2.3	1.000	117	240	117.0	240.0	56.70	55.61
R4.3	1.352	145	359	196.0	485.3	31.52	32.39
R5.2	3.308	109	1136	360.5	3757.9	18.31	18.77
R6.2	2.070	127	790	262.8	1635.3	23.35	24.62
G1.2	1.027	97	135	99.7	138.6	76.75	66.85
G3.1	1.096	86	68	94.3	74.5	75.62	71.42

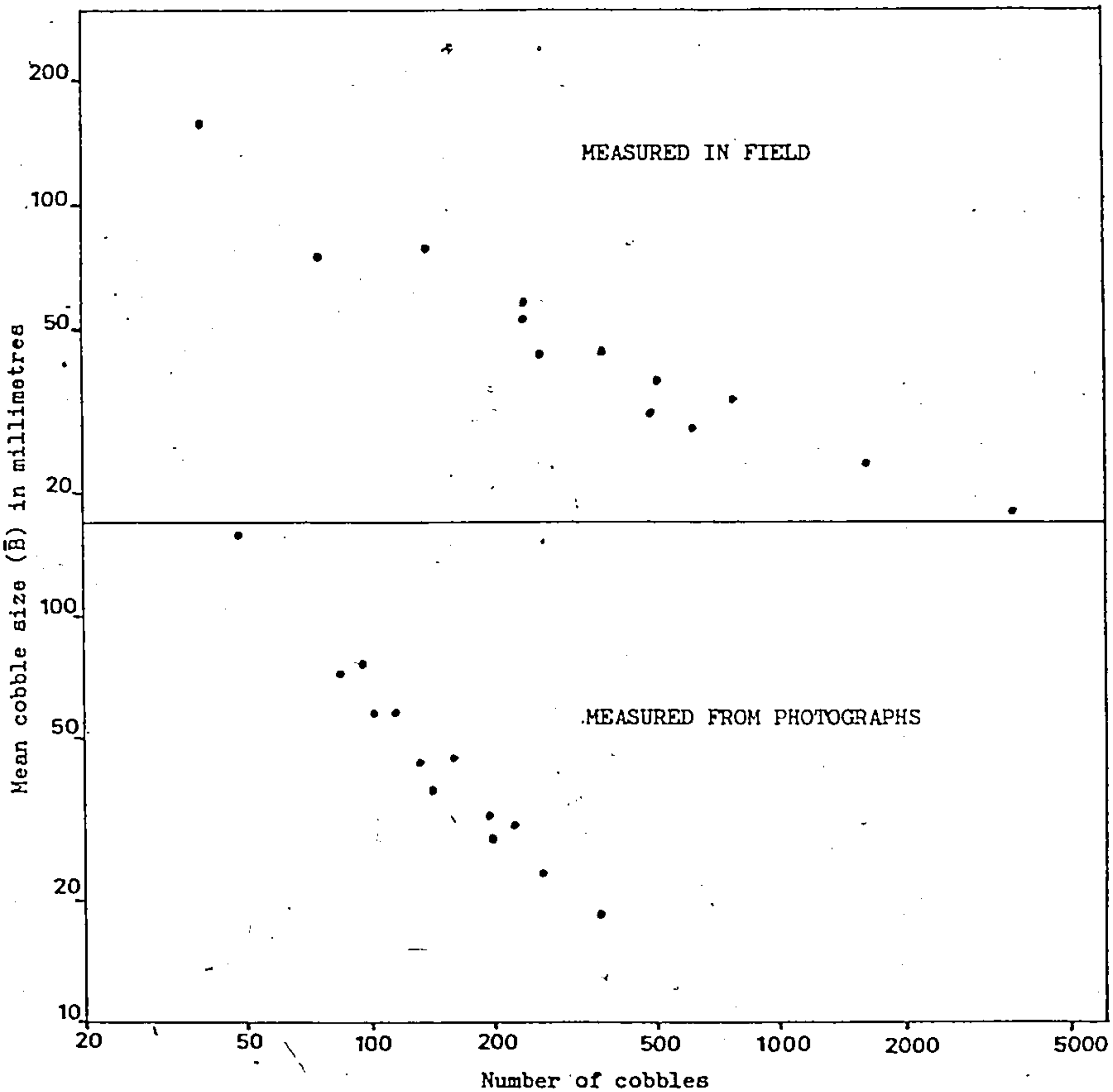


Figure A.2.1: Control curves for the calculation of mean cobble size ($\bar{B}$).

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