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Glacial Geomorphology In A Sub-polar Proglacial Lake Basin: A Process-response Model

David Martin Barnett

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GLACIAL GEOMORPHOLOGY IN A SUB-POLAR PROGLACIAL LAKE BASIN: A PROCESS-RESPONSE MODEL

by

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Submitted in partial fulfillment
of the requirements for the degree of
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ABSTRACT

At the margin of the Barnes Ice Cap in Baffin Island a sub-polar glacial regime prevails adjacent to proglacial Generator Lake. Fluctuations of both ice margin and lake level provide an excellent opportunity to assess the influence of sub-polar ice on geomorphic processes. Interpretation of paleoenvironments would be enhanced with an appreciation of responses (landforms) which may be diagnostic of sub-polar ice influence. This study develops such a process-response model for the Generator Lake basin.

Methods include radiocarbon and lichenometric dating to develop a chronology, echo sounding to determine sub-lacustrine morphology, bottom sampling by grab sampler and gravity corer plus bathythermograph and Knudsen bottle for water sampling and temperature profiles. Fluvial processes are examined using sediment sampler, current meter, thermometer and dye plumes. Morphology (geometry) is established by transit, hand level, photogrammetry and direct measurement. Internal structures are investigated by digging, till fabric and natural exposures.

The four elements - process, geometry, response and time are integrated as a qualitative model of circular form symbolizing the lake basin. A radiocarbon chronology of 4000 to 4500 years is based on 11 dates from detrital organic matter derived from deltaic sediments and supercedes a

tentative lichen chronology. The lengthened chronology is still compatible with regional deglaciation chronology. Relative dating by lichenometry is still adequate.

No single landform is peculiar to the sub-polar ice environment but the degree of development is bold for sub-lacustrine moraines, anchor ice sediment rafts, ice pushed ridges and proglacial lake terraces. The form of the ice ramp and ice-cliff are, in part, attributable to the thermal regime of the ice.

Fluvial and lacustrine processes are highly sensitive to small changes of mean summer temperatures, with geomorphologically active and inactive summers recorded. Asymmetry of intensity of these processes is attributable to the Barnes Ice Cap, and the efficacy of lacustrine processes on pebble morphology greatly exceeds that of frost shattering.

Ice pushed ridges result from high magnitude-low frequency events but are not amenable to high resolution prediction being dependent upon wind direction during the critical minutes of ice impinging on a 2-5° shore slope.

Sublacustrine moraines are predictable in size, form and spacing given adequate information on ice flow, retreat rates and annual lake level fluctuations.

Current processes are of similar magnitude to those of the past 4000 years. Derived process rates from a raised delta sequence are compatible with present fluvial conditions.

A spatial association, as a suite of landforms, diagnostic of the sub-polar ice environment, is recognized which individually the landforms fail to demonstrate; sublacustrine moraines are not annual and require a ramped ice margin which is a function of lake level rise, which can be approximated by lake area. Paleogeographic studies of landforms in the northeastern quadrant of the Laurentide Ice Sheet could give additional insight into the glaciology of the former enormous ice mass.

ACKNOWLEDGEMENTS

Many persons and organizations have contributed to the fieldwork and subsequent production of this dissertation. Its origin is now obscure but began within the Baffin Island Program of the former Geographical Branch of the Department of Energy, Mines and Resources. Drs. J. D. Ives' and J.T. Andrews' considerable enthusiasm for Baffin Island played no small part in the beginning.

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The Terrain Sciences Division of the Geological Survey of Canada merits special mention for kindly formalizing the opportunity for educational leave at the University of

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Dr. Barrie McDonald supervised, and Ron Kelly processed, the sedimentological samples with great care which is hereby acknowledged. Fred Barber of Marine Sciences Branch and Dr. C. F. Mike Lewis (while at Canada Centre for Inland Waters) together helped considerably with the limnological aspects of the project, both in the form of advice and arranging the loan of equipment.

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The collaboration and discussions with Dr. Gerry Holdsworth in writing the paper on sublacustrine moraines was an excellent opportunity to clarify ideas on the processes operating and their tempi. Each contributed data which at first seemed unpalatable to the other's ideas but

with time each was able to modify and refine his own thoughts and a better model was the result."

My colleague, Dr. Lynda Dredge, was most helpful in discussing some of the material and elaborating on the finer points and pitfalls of dissertation presentation.

My wife, Karin, has cheerfully endured several different versions and I thank her for her unfailing support throughout the sometimes painful process of preparation.

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CHAPTER 1

PHILOSOPHY, PURPOSE AND METHODS

THE RESEARCH PROBLEM

The land-based margin of a polar or sub-polar ice mass has a thermal regime normally below pressure melting point. Where such an ice cap margin impedes the drainage of ice-free terrain a lake forms, which, if deep enough not to freeze to the bottom provides a winter heat source and so changes the thermal regime of the ice margin significantly. Drainage from such a lake is by either an ice marginal or other topographic low across the watershed. The position of the glacial ice margin together with that of the watershed governs many of the hydrological characteristics of the lake basin and consequently landforms developed within the basin are affected by any change in the position of the ice margin.

A fluctuating ice margin often creates more than one set of landforms as lake levels adjust to newly exposed or freshly dammed outlets thus varying the relative areas of glacial, lacustrine and sub-aerial geomorphic activity within the basin. Such landscapes in the sub-polar environment have received very little detailed attention primarily due to inaccessibility and the related factor of cost. Nevertheless some of the existing Canadian proglacial lakes are the closest extant equivalents to the now extinct glacial

Great Lakes - - a theme successfully developed by Goldthwait (1959) who used photographs from Baffin Island to illustrate scenes in Ohio during the ice age. Boulton (1975, p. 39) suggested that ice masses with a cold margin make good analogue models for study of Quaternary ice sheets. Proglacial lake systems formerly covered enormous areas of Canada and their deposits have had profound effects on the present landscape particularly in settled areas. The use of the word system is significant as the ice-lake-land relationships are interdependent.

The Scope of the Dissertation

The dissertation deals with the geomorphology of a Canadian example of a proglacial lake environment system, the Generator Lake basin in central Baffin Island, N.W.T. (Figure 1-1), the detailed introduction to which is deferred to Chapter 2. Field work was undertaken in 1966, 1968 and 1969. In particular the aim is to develop a model to portray the geomorphic system of process-geometry-response and the interaction between the various contiguous environments (lacustrine, littoral, sub-aerial). The landscape of the basin is dominated by landforms created by several phases of a proglacial lake dammed by a sub-polar ice cap, the land based margin of which is theoretically below pressure melting point and shown to be so by Hooke (1976, Figure 6, p. 57). As it is frozen to the underlying material, meltwater does not normally penetrate to the base of the marginal ice but tends to remain as surface drainage.

Figure 1-1: Map showing regional location of Generator Lake (G), Bieler Lake (B), Conn Lake (C) and the Barnes Ice Cap (shading). Principal settlements of Baffin Island are shown.

The intent is to use the data collected to develop and examine a conceptual approach to this segment of glacial geomorphology. The study includes new data on the glacial history of the area but this material is used primarily as a chronological framework on which to assess process and response.

Basic questions posed are, therefore, what are the processes operating in such an environment? How has the landscape geometry influenced them and are the responses distinctive at any or all of the interfaces? These are the ice-water interface, the ice-land-water interface and the land-water interface. In the latter case are the responses notably different from those in a non-proglacial lake? Is it possible to identify a geomorphic unity in the drainage basin other than a hydrological one?

The ice-land interface has already been investigated by Goldthwait (1951a) and was re-examined by Hooke (1973a, 1973b) in this locality and was considered theoretically by Weertman (1961). Therefore this aspect of the proglacial environment will be given only a cursory review.

An attempt is made in this study to establish the rate at which processes are taking place, with the ancillary question of whether uniform or variable rates are indicated from the available data. As the framework for assessing process intensity is based on chronologies derived by different techniques of variable accuracy, an appraisal and comparison of the techniques is included as a basis for judging reliability of the data. It is considered unreasonable to anticipate elegant or even completely satisfactory answers to all questions asked. Progress can be measured by the number of new and perhaps more incisive questions it is possible to ask as a result of the investigations.

Prospect for the contribution to knowledge: Although this is not the time to asses the contribution to knowledge made by this dissertation, it is appropriate to consider such a contribution under the heading 'purpose'. Three areas were recognized where contributions might be made; unranked they are: collection of new local and regional data with concomitant interpretation, innovative or improved field techniques, and philosophical or conceptual development of the geomorphology of the proglacial environment. The melding of the three areas to include significant parts of the existing literature constitutes the dissertation.

STATUS OF PROCESS-ORIENTED GEOMORPHOLOGICAL WORK IN THE PROGLACIAL ENVIRONMENT

I General

This regional review outlines several types of investigations in the proglacial environment each of which has some bearing on the theme being developed, but none has the same objective as the present one, and in some cases the link with proglacial process is tenuous.

II North America

a) Canada: Most studies in the Arctic have concentrated on one particular aspect of the geomorphology such as the classic work of Goldthwait (1951a) on the process of moraine formation along the margin of the Barnes Ice Cap within the field area of the present study. Also on Baffin Island, in the Isortoq Valley, Andrews & Smithson (1966) devoted considerable effort toward identifying processes involved in the creation of cross-valley moraines, distinctive landforms intimately related with the proglacial lake environment. Holdsworth (1973a) studied the calving process of the Barnes Ice Cap at Generator Lake and the evidence for a surge of the same ice (Holdsworth, 1973b). Hooke (1973a, b) has more recently studied the structure and flow of the ice margin and their effects on ~~the~~ formation of ice-cored moraines at the edge of the Barnes Ice Cap. Hooke (1976) also concluded that some of the ice at the base was of Pleistocene age. Maag (1969) made detailed observations on Axel Heiberg Island of seasonal variation of ice dammed lakes and morphological change associated with ice marginal drainage. The detailed and painstaking work of Church (1972) on processes in Arctic fluvial environments included data from the immediate proglacial environment in central Baffin Island.

Other studies have concentrated on lakes themselves (not all proglacial ones) as the central focus: for example the chemical composition, stratification and history of the water on northern Ellesmere Island by Hattersley-Smith & Serson (1964) and Hattersley-Smith et al. (1970), or the study of sedimentation and other physical limnological characteristics in Stanwell-Fletcher Lake on Somerset Island (Coakley & Rust, 1968; Rust & Coakley, 1970) or limnological investigations such as by Oliver (1964) in Nettilling Lake on Baffin Island.

On Devon Island process-oriented research has been conducted by a McMaster University group on the development of present Arctic marine beaches (Owens & McCann, 1970; McCann & Owens, 1969). The same group studied fluvial processes on Cornwallis Island (McCann, Howarth & Cogley, 1972) a periglacial environment, but still characterized by very rapid changes in tempo.

Another focus was on the behaviour of glacial ice both by the Axel Heiberg Island group from McGill University formerly led by Dr. Fritz Muller; by the Defense Research Board of Canada group on Ellesmere Island led by Dr. G. Hattersley-Smith and by the Baffin Island group from the Glaciology Division of Environment Canada.

Papers on the Alpine environment were also examined for methodology and to consider similarities and differences with studies at higher latitude. The detailed work of Mathews (1956) is a good example of a study of the processes of sedimentation and physical limnology in a glacial lake and Marcus (1960) examined the process of drainage of Tulsequah Lake, another glacial dammed lake in British

Columbia.

Studies of Pleistocene ice-marginal lake deposits and landforms have been conducted in several localities and Shaw (1975) discussed what he called proximal glacial deposits from Edmonton and the Okanagan Valley at the same time calling for systematic studies in modern glacial lakes.

All the above studies have some degree of overlap with one or more of the others but none seeks to draw together the variety of processes into a single system which is admittedly a big task. The intent of the present study is to at least begin to examine linkages and relationships between these various geomorphic processes.

b) Alaska: Studies in Alaska have been in some ways similar to those in Canada but the proglacial environment there is more Alpine in character. Stone (1963) concluded that Alaska has the greatest concentration of ice dammed lakes in the world (53 documented) and he sought to generalize on their distribution (in tributary valleys) and draining characteristics which he concluded were by subglacial routes, commonly regularly. Price (1965) examined glaciofluvial processes in front of the Casement Glacier with good historical chronological control. McKenzie (1969) made detailed observations of the process of collapse of a kame terrace in Glacier Bay. The study of outwash sediments of the Mendenhall Glacier by Ehrlich & Davies (1968) concentrated on glaciofluvial processes but also considered process intensity, an aspect not often considered. Gustavson

(1972) documented the sedimentation and physical limnology of Malaspina Lake in a very similar degree of detail as was possible in this study.

c) Washington: The work of Fahnestock (1963) on the processes operating in a glacial stream draining from Mount Rainier is an excellent example of modern quantification of proglacial processes.

III South America

Patagonian studies have been undertaken from time to time though they are primarily of a reconnaissance type by non-residents, for example a historical study of the Moreno Glacier including a discussion of the bursting of the ice dam in 1966 (Liss, 1970).

IV Antarctica

Relatively few geomorphological and related studies in Antarctica have concentrated on processes. The literature reveals two foci, one on soil forming processes including patterned ground development and secondly on limnological processes in the 'glacial' lakes in the dry valleys of Victoria Land. The former are exemplified by the group of papers edited by Tedrow (1966) and the latter by a study of lake waters by Yamagata (1967). Russian studies have produced papers on what is described as a new type of high latitude lake (Govorukha & Simonov, 1967; Kruchin & Simonov, 1968). The lake characteristics are warm water at the bottom and unstratified fine grained sediment deposited very slowly under a semi-permanent thick ice cover. Other studies of interest are by Souchez (1967) on the process of formation of shear moraines in South Victoria Land which are comparable with those of the Barnes Ice Cap in the field area and secondly a study by Cailleux (1968) who examined pebble roundness in McMurdo Strait as an index of glacial erosion in the periglacial environment.

V <u>Europe</u>

a) Greenland: Of the many studies in Greenland three illustrate investigation of process. Goldthwait (1960) examined processes leading to formation of the land-based ice cliff at Nunatarssuaq and to its subsequent maintenance of form. The processes of lake ice growth and decay were investigated by Swinzow (1966) on Lake Tuto which included documenting water temperatures to the lake bottom. Hooke (1970) considered process and morphology of the ice sheet near Thule.

b) Iceland: Although Iceland's proglacial environment is maritime in character some aspects of recent studies are relevant. For example Price (1970) examined multiple moraines formed at the margin of Fjallsjökull which he suggested (page 27) may be formed annually. Howarth & Price (1969) documented depth and morphology of a series of proglacial lakes southeast of Vatnajökull, some of which extend below present sea level. They also discuss a calving mechanism of interest in view of the glaciological findings of Holdsworth to be discussed later.

c) Spitsbergen: Investigations of proglacial processes by Boulton (1967, 1968) concentrated on flow tills resulting in multiple till sequences intercalated with stratified deposits but representing a single glacial phase.

d) Scandinavia: Glacier-dammed lakes ~~have~~ traditionally been topics of both intellectual and practical interest. Liestøl (1956) reviewed Norwegian examples assessing the evidence for processes leading to rapid drainage. Howarth (1968) examined a supraglacial extension of an ice dammed lake at Tunbergsdalbreen in Norway which yielded evidence for floatation of the ice barrier, which combined with the formation of a subglacial tunnel, he thought to be the processes leading to draining. Hoppe (1959) commented on various views on the formation of De Geer moraines which are marine morphological equivalents of those formed in Generator Lake. Worsley (1974) documented "annual" moraine ridges at Austre Okstindbreen which he attributed to ice push during the winter season. These Norwegian features are minor landforms up to 1 m high.

e) The Alps: No survey of ice-dammed lakes would be complete without reference to the Märjelensee dammed by the Aletsch glacier, but as Embleton & King (1975, p. 534-535) record it is now gone, but even historically it had few similarities with Generator Lake as it drained subglacially beneath heavily crevassed temperate ice.

VI <u>Rest of the World</u>

a) U.S.S.R.: As elsewhere a focal point of study is the process of catastrophic draining of glacier dammed lakes mainly in the Russian Arctic Islands (Grosval'd & Koriakin, 1962). A review by Kvasov & Krasnov (1967) on 'basic problems in the history of periglacial (proglacial?) lakes of the northwest' covers a variety of studies gathered into one volume. (It would appear that the use of "periglacial" is a reversion to the original meaning of 'adjacent to the ice' which is roughly equivalent to proglacial, a word preferred in some of the other papers in the volume). Most papers are historical in approach but the one by Aseev (1967) attempts a genetic classification of "glacial lakes of plains". The five classes are: subaerial hollows of glacial run-off, subglacial channels, trough lakes, glacial depression lakes and those of complex origin.

b) Central Asia: Hewitt (1967) studied both process and process intensity in the Karakoram looking at seasonal effects on ice front deposition.

METHODOLOGY

A primary approach is one of climatic geomorphology as it is the imprint of a low temperature condition of the ice margin which was investigated in the proglacial zone. As deglaciation proceeds it changes the morphogenetic zonation of the field area from glacial to proglacial and hence, <u>a priori</u>, the geomorphology is temperature dependent. The historical method was used to outline the chronology to set the scene for a systematic approach to process and response.

A very pertinent discussion is presented by Andrews

(1975, p. 8-11) of the desirability of using a systems approach to glacial landscapes in which he states (page 10) that "the process-response model ... should form the keystone to many studies in glacial geology and geomorphology". However, he outlines the difficulty, but often basic necessity, of working backwards from response to process as similar responses may originate from different combinations of processes. King (1966, p. 17) stressed that most geomorphological systems exhibit great complexity due to the interaction of a very great number of variables. Seeking order and meaningful linkages between these variables is one of the uses of modelling with a view to simplifying some of the complexities of the landscape.

Process-Response Models

The concept of a process-response model has been used successfully in geology, particularly by Krumbein (1963) and Whitten (1964), and M. A. Carson (1969) discussed its use in physical geography, particularly in relation to hill slope development under mass failure. Whitten (1964, p. 455) asserts that often geological projects have vague objectives (a condition probably not confined to geology) and that the utility of a model is to provide a conceptual framework. Krumbein & Sloss (1963) consider a model an aid in identification of generalizing principles and as a basis for prediction.

A process-response model, sometimes posed as a cause and effect model, is by no means a simple relationship since interaction or feedback tends to blur the distinction. This is particularly so where multiple processes are operating but it is intellectually rewarding to attempt to make distinctions and assess the dominant processes creating the responses. Responses are the easier to identify.

Each part of the landscape of the Generator Lake basin is a response to several processes which have changed through time from all subaerial to all glacial to its present, and most interesting, intermediate phase. Inevitably it is the more distinctive responses, the discrete landforms, which draw most attention. The discrete landform is also significant from the practical point of view as it implies a localized change in surface materials and/or slope. These changes, in turn, may indicate significant change in granulometric composition, slope stability and bearing strength when thawed, all of which are of potential concern for overland vehicle movement or engineering work. Some depositional landforms also have potential significance as sources of construction aggregate. Discrete landforms represent localized responses resulting from a combination of processes, time, and the form of the geometrical basement.

Therefore, the intent is to develop a process-response model based on the data derived from Generator Lake. This qualitative model is introduced in Chapter 4 following an introduction to the present physical milieu in Chapter 2 and the regional and local chronological setting in

Chapter 3.

STRATEGY

Owing to the paucity of existing data it was necessary to be selective in gathering data potentially offering useful insights into processes operating and their intensity or tempo. Obviously generally only sufficient measurements to yield order-of magnitude quantitative data were possible. In regional terms it was desirable to establish a credible link between the official meteorological data available for central Baffin Island and that from the Generator Lake basin, (details of which are set out in Chapter 2 under 'Climate'), but the essence of which is the use of the maps showing regional temperature deviations from the norm for June, July, August and September in each year from 1946. Hence in this way the summer temperature norm may be defined independently, even if imperfectly, and a medium term climatic trend established: it is getting colder.

Fluvial and lacustrine processes were obviously suitable for monitoring to examine quantitative relationships with summer meteorological data. Glaciological processes at the ice-lake interface were also potentially active enough for a short term observation program to be rewarding, and glaciological processes at the land-based margin have been examined by others (Goldthwait, 1951; Hooke, 1973a, 1973b, 1976; Hudleston, 1976).

To set these current processes in a reasonable chronological context and to enable deduction of rates for slower processes such as moraine formation it is necessary to establish a chronology for the lake basin. Two approaches were considered and tried, those of lichenometry and radiocarbon dating. Chronological separation of environments within the basin was known to be possible by lichenometry (Sim 1961; Andrews, 1963) but conversion of lichen dimensions to a reliable time scale was shown to be somewhat uncertain earlier (Barnett, 1967). Prior to 1968 no datable material suitable for radiocarbon dating had been found in the area but careful searching of deltaic sediments yielded plant detritus in eleven different localities. Discussion of the samples is deferred to Chapter 3 and details of the collection sites, sample size and treatment are set out in Appendix A.

METHODS

I Introduction

This section is divided into field and laboratory techniques and the former into innovative and conventional subsections. Fuller treatment is given to the innovative part but the conventional part is detailed enough to allow the reader to duplicate the technique.

II Field techniques

a) Innovative: These techniques are either not well established or variations of generally accepted methods. Elaboration is appropriate here and some further comment is made in context, where necessary, when presenting results.

i) Echo sounding: The use of an echo sounder to determine lake bottom morphology is well known but two variations on the conventional method were used. First, the quasi-permanent ice cover on Generator Lake necessitated sending the sound pulse through more than one metre of ice and picking up the return pulse through the same amount of ice. In a second variation, the transducers were rotated 90° and directed toward the ice cliffs and lowered slowly through the lake ice to the bottom of the lake in order to get a series of vertical profiles of the underwater cliff-face of the ice cap. This was achieved by placing the transducers in a metal cage (Figure 1-2) equipped with long cables which was then lowered through a series of holes cut in the lake ice close to the ice cliffs. The writer is indebted to Dr. J. G. Fyles, Geological Survey of Canada, for suggesting this latter technique.

The Kelvin-Hughes MS26B is a bulky ship's echo sounder normally used for navigational purposes, with a range of 1320 metres and yielding a continuous graphic record of water depth. This is achieved by two transducers, one sending an impulse and the other receiving an echo of the sound wave; the elapsed time is translated automatically into depth.

Power was supplied by two 12-volt car batteries mounted in series and connected with a rotary converter to give AC current. Batteries were frequently charged and power output

Figure 1-2: Cage mounting for transducers in rotated mode. Pulse would be directed toward the camera. (GSC Photo 168029)

Figure 1-3: Echo sounding equipment mounted on a Nansen sled. One of the transducers is visible in position to the right of the front of the sled. The other transducer was located on the opposite side of the sled. (GSC Photo 203028 B)

verified daily by counting revolutions of the stylus at slow speed against a stop watch. The instrument and power source were then mounted on a Nansen sled which was towed by motor toboggan (Figure 1-3). Otherwise, no technical adjustments were necessary. The transducers were free (rather than mounted as in a ship) and were generally placed up to 1.5 metres apart, one on each side of the sled. The sounding was then at the mid-point beneath the sled which travelled along the profile line, station spacing being measured by a steel tape.

Both transducers had to be well bonded to the ice surface, thus limiting the technique to spot readings only, rather than the ideal - a continuous depth trace along the profile lines. A bonding medium was necessary as the ice surface was rarely found to be completely smooth. Initially a solution of calcium chloride was tried, but this was found to be markedly inferior to a small amount of heavy mineral oil. Before bonding was successful, many surfaces had to be prepared by removal of some rough or candled ice, as the presence of air pockets or large concentrations of bubbles in the ice decreased the intensity or occasionally completely absorbed the return pulse.

Transmission of an impulse through ice affects the elapsed time as ice is a slightly less dense medium than water. However, the ratio of ice thickness to water depth was sufficiently small not to affect the accuracy of the measurement significantly. Table 1-1 shows good agreement

TABLE 1-1

COMPARISON OF ECHO SOUNDING AND LINE SOUNDING (METRES)

E/S Stn.#	E/S Depth	Line Sounding	Difference
6/1	16	16.19	0.19
7/-1	23.25	23.25	0.00
7/48	28.75	30.00	1.25
7/98	22.75	23.00	0.25
7/148	21.25	21.00	0.25
7/248	46.25	45.50	0.75
8/32	20*	21.56	1.56
8/47	30.25	29.67	0.58
9/1	43.75**	40.61	3.14
10/1	47.5	47.60	0.10
10/2	52	52.39	0.39
10/3	21	21.67	0.67
		$\bar{X}$:	0.76

* Poorly defined top to trace - soft sediment (?); 8 days between E/S and line sounding.

** Possible interference from proximity to ice cliff.

All echo soundings interpolated to nearest 1/4 metre. Ice platform rose with lake level thus introducing a small error depending on time elapsed between echo and line soundings.

of result from line sounding and echo sounding at selected holes.

When in the submerged mode and beamed horizontally, the transducers were theoretically free to record continuously, in practice this was unworkable because of 'noise' pick-up confusing the trace and a tendency for the transducer cage to swing. Some of this 'noise' originated with both the power and suspension cables running against the hard lake ice. Consequently, spot readings were taken in increments of two metres of depth with the transducers stationary. At first the cables were controlled on a single drum on a gas driven winch. Subsequently better control of the cables was achieved by separating them. The cables were attached to the rear of two motor toboggans, one on each side of the hole; then the transducers in their single mounting were lowered to the lake bottom and raised slowly by driving the toboggans away from the hole two metres at a time. This arrangement added stability to the cage which tended to swing when in the deeper parts of the lake and eliminated much of the 'noise' on the trace.

ii) Radiocarbon dating: Although radiocarbon dating is a well known technique, suitable materials for dating in the field area were scarce and difficult to locate. Careful examination of deltaic materials associated with specific lake phases in eleven cases revealed sufficient detrital vegetation to make dating possible. This is the first time such materials from the Generator Lake proglacial environment have been used for dating. Pleistocene proglacial materials have commonly been dated from the Great Lakes area, e.g. Dreimanis (1969), but Holocene examples are much fewer and have sometimes presented difficulties for interpretation. C. E. Carson (1968) radiocarbon dated some Alaskan lacustrine strandlines from vegetative matter but this material was in situ and glacial ice was not involved in lake level changes.

iii) Lichenometry: This dating technique developed by Beschel (1961) has not found complete acceptance thus far, with the most vigorous detractor being Jochimsen (1966). Ideally, lichenometry can be used as both a relative dating and an 'absolute' dating method. The former is more generally accepted than the latter. In this study the lichen chronology (developed first) (Barnett, 1967) is used as a basis for comparison with the radiocarbon chronology - the first time that this has been possible. The chronologies are presented in Chapter 3, although it can be noted here that the two differ by a significant amount. The lichen chronology is more likely to be in error as no independent growth rate was established for the field area. Part of the lichenometry problem lies in the technique of data collection which has no generally accepted standardized format. Reduced to its simplest elements, one has to locate and measure the diameter of the largest lichen thallus of a diagnostic species on the particular substrate to be dated. Personal experience led the writer to abandon a specifically predetermined sample area for location of the largest lichen. Lichen population density varies with age so that the adoption of a standard ten metre square (minimum area advocated by Beschel, 1961, p. 1047) leads to measurement of the largest thallus from very different available populations in different squares. Hence the more sparse the population, the larger the area it is desirable to sample, provided that the larger areas do not transgress significant micro-environmental boundaries. Of particular concern is the presence of late-lying or semi-permanent snow patches which often carry a sparse population of small lichens and, therefore, tend to yield 'young' dates. Other areas to be avoided are splash zones beside water bodies, moisture retaining hollows on individual rock faces and bird perches, all liable to promote lichen growth and give 'old' dates. The method adopted was to search intensively an unmeasured area in excess of 10 m^2. Larger areas were searched when the lichen population was low. The ten largest lichen thalli in each area were recorded and appear as Figure 12 in Barnett (1967).

iv) Pebble morphology: Techniques for examining pebble morphology are varied and usually include measuring several different parameters, but investigators have not conformed to a single standard but have developed their own. Equally, no standard method appears to have evolved for measuring degree of roundness. In the present study the parameters suggested by Cailleux were used. This technique was chosen as it was reasonably simple and was shown by King & Buckley (1968) to be satisfactory for distinguishing geomorphic environments in west central Baffin Island.

The technique requires the measurement of the a, b, and c axes of a pebble and the minimum radius of curvature in the principal plane (Blenk, 1960). This latter parameter is the most difficult to measure, but is more sophisticated than using Krumbein's reference set of ten pebble outlines for rapid visual classification (Krumbein, 1941, Plate 1). King & Buckley (1968) used a set of concentric reference arcs carefully scribed on stout card at 5 mm intervals and then positioned the rock particle or pebble on the card to get a direct, though still visual, approximation of the minimum radius of curvature. In the present study a similar, but potentially improved method, was used: the arcs were scribed on a piece of transparent 'perspex' acrylic sheet and the rock particle viewed through the sheet, thus leaving all arcs visible at all times to ensure the best fit. This system still retains a subjective component which was tested statistically to assess operator variance. This aspect of the technique was found to be unsatisfactory in this study and is discussed in Chapter 4.

b) Conventional:

i) Meteorological data: Meteorological conditions affect processes and govern their effectiveness. They were monitored in detail during July and August of the 1969 field season. The following parameters were measured continuously at base camp: air temperature, relative humidity, air pressure and total run of wind augmented by twelve-hourly observations (0000Z and 1200Z) of maximum and minimum air temperature, wind speed and direction, type and amount of precipitation, type, estimated height and amount of cloud cover. In addition, near-shore lake surface temperatures, lake level variations and temperatures of a small stream were monitored every twelve hours. Weather conditions visibly affect snow cover, lake ice cover and glacial ice when exposed, which in turn affect run-off, stream competence and load, and deltaic and lacustrine sedimentation. Meteorological conditions also affect the tempo of littoral processes since they depend upon the size of the lead around the lake ice. Whether the lake ice is in a single sheet or several pans governs its capacity to do work in a wind, although the effectiveness of the ice push is also dependent upon the slope of the shore zone, the degree of decay of the ice, and the availability of loose material in the push zone.

ii) Levelling: As no readily accessible bench marks were available all levelling was done to locally established temporary bench marks and data are presented as heights above or below each local datum. No attempt is made to relate heights to sea level datum. The level was a Zeiss Ni-2 and a Hultafors 4 m levelling rod was used. The longest traverse (22 km) was along the southern shoreline of the 76 m lake and was outlined in Barnett (1967, p. 185). It demonstrated both shoreline continuity and that it was basically undeformed within the limits of the measurements. Thus it denies the hypothesis that the shoreline may be tilted isostatically. The end points were only 22.5 cm different in height with intervening stations deviating slightly more above and below the line of apparent tilt. Determination of the level of the former water plane was of the same order of accuracy: $\pm$ 15 cm.

iii) Till fabric: The three-dimensional fabric measured was of clasts having a measured a:b axis ratio of at least 2:1 (the 'mesofabric' of Derbyshire, McGown & Radwan, 1976). Paired pits were dug close to the moraine crest and each pit consisted of 50 pebbles. One pit was from the distal and one from the proximal side of the moraine. The choice of location of the pits close to the crest was, in part, because they blow snow free and drain well, and therefore potential reorientation of the mesofabric by frost action would be minimized. The sites thus selected are therefore comparable with those of the Isortoq valley (Andrews & Smithson, 1966). The data are set out in Appendix D.

iv) Sediment sampling: Suspended sediment was collected by multiple casts using a hand-held DH-48 depth-integrating suspended sediment sampler mounted on a series of one foot rods. The container was a one pint capacity milk bottle. The samples were filtered in the field through a Munktells "OO" 18.5 cm diameter filter paper with an average ash content of 0.00012 g. The used filters were stored in clean polyethylene bags.

v) Current metering: An Ott current meter was used (Meter serial #12862) with a 100 mm diameter "V Arkansas" propeller completely submerged at intervals across the stream profile (minimum four intervals). The revolutions were counted automatically by a battery operated counter with a digital recorder. The relevant equations for 'n' are:

if $n < 4.71$ rps : $0.4064\ n + 0.131$ ft./sec.

if $n > 4.71$ rps : $0.4231\ n + 0.052$ ft./sec.

vi) Dye plumes: Safranin 'B' dye was used for tracing the path of stream flow and also for establishing flow-travel times. Concentrations used were not measured but simply based on a trial-and-error basis. A slurry of dye was mixed in a five gallon pail and timing was by wristwatch with a sweep second hand.

vii) Lake bottom sampling: Two Dietz-Lafond grab samplers (Figure 1-4) were used for sampling bottom materials through holes drilled in the lake ice at one kilometre intervals. The samplers were suspended on aircraft cable and attached by carabiners. Retrieval was by a gas driven portable winch (Hydro Products). The sampler works when a protruding foot strikes the bottom and releases a spring which closes the jaws. Slow recovery minimizes wash out and samples were scooped into polyethylene bags after initial inspection in the jaws of the sampler.

A Phleger corer (Figure 1-5) with detachable 3 x 25 lb.

Figure 1-4: Preparing to lower the Dietz-Lafond grab sampler. Note the slushy water on the ice and the power winch to provide slow but steady retrieval. (GSC Photo 167988)

Figure 1-5: Phleger gravity corer with 1 m barrel and one 25 lb. weight. Note candling lake ice and thermal decay at the hole margin. (GSC Photo 168031)

lead weights was also used for bottom sampling. It is equipped with transparent plastic core liners one metre in length and detachable cutting rims with bayonet mount. All combinations of weights and no weights were tried to collect cores from several different drill holes but the attempt was not successful. All the cutting rims available (three or four) were considerably damaged by the free falls from which it was concluded that impact must have been on bedrock with only a few centimetres of lake silts above.

viii) Lake temperatures, depth and water sampling: Lake temperature and depth was measured by a Wallace and Tiernan bathythermograph (Figure 1-6) which automatically scribed the readings on a coated glass slide. The original data is all available in Barnett, Forbes & Whytock (1970). Attached to the same cable was a Knudsen bottle with spring loaded closures and reversing thermometers both of which were triggered by sending a brass messenger (Figure 1-6) down the cable when the bottle was in position at the lake bottom.

ix) Glaciological observations: Direct observations of glaciological processes were limited but included monitoring the morphology of the ice cliffs whenever possible and noting the character and occurrence of any calving blocks. Longer term processes such as glacial movement have been and continue to be studied by the Glaciology Division of the Inland Waters Directorate of Environment Canada who kindly provided access to some of their material.

Figure 1-6: Bathythermograph (right), brass messenger (centre) and Knudsen bottle with reversing thermometers (left) in front of sampling hole in lake ice. Fraying of cable due to dragging over lake ice for many kilometres.
(GSC Photo 167997)

x) Airborne pollen sampling: A program to sample airborne pollen was basically unsuccessful. The pollen trap supplied by Dr. J. Terasmae (formerly of the Geological Survey) had been used successfully in more temperate latitudes but in central Baffin Island the collecting medium, glycerine jelly, hardened too quickly preventing adhesion of the grains. Two modern pollen samples were subsequently collected from mossy vegetation in shallow pools by Dr. G. Holdsworth, Glaciology Division, Environment Canada, at the writer's request, to enable comparison with fossil pollen data as set out in Appendix B.

III <u>Laboratory techniques</u>

All laboratory work on samples was carried out by the staff of the Geological Survey of Canada as part of routine processing to meet or exceed prevailing standards.

a) Radiocarbon samples: Samples received careful routine treatment as detailed in Appendix A. The laboratory maintains a very low background count thus generating a relatively small statistical error term which is quoted at the two standard deviation level. Most samples were corrected for C_{12}/C_{13} fractionation, the earliest corrections being run by Isotopes Inc. but a majority being run by the Geological Survey. The three samples without corrections were the result of losses within both laboratory systems (Table 3-3).

b) Palynological samples: Samples were examined routinely for their assemblage by R. J. Mott and his findings are listed in Appendix B. In addition, the writer asked for a calculation of absolute pollen abundance to get an indication of the relative richness of pollen in the lake bottom sediments. They were considered by Mott to show low abundance. Mott's interpretative comments are included in the Appendix.

c) Bryological samples: All samples are from palaeoenvironments and were collected from detrital sediments. This latter information was not provided for Dr. M. Kuc. His moss identifications are recorded as supplied but his interpretative comments required some editing to achieve linguistic plausibility but in general his sentence structures and interpretations are retained.

d) Grain size analysis: A full account of sedimentology laboratory procedures followed at the time the Generator Lake samples were processed is set out in McDonald & Kelly (1968). Adjustments to some procedures have occurred since that time. As the grain size analysis results are not used in the dissertation for rigorous analysis it is inappropriate to reproduce all the detail here but simply to outline the basic technique.

The bulk sample was air-dried, weighed and material less than 4 mm was split four ways using a riffle sampler. Moisture content was measured from a 10 g. split dried at 105°C for 10 min. A 50-55 g. split was used for grain size analysis. A dispersant, 0.5 N sodium hexametaphosphate (100 ml) was used after a distilled water wash in a 500 ml conical beaker. The mixture was then stirred for 15 sec. using a milk shake machine and then heated to boiling point. After allowing it to cool, it was stirred and agitated ultrasonically for 20 sec. The mixture was then wet-sieved through a 44 micron sieve using distilled water. The material retained on the sieve was oven dried at 105°C. and allowed to cool, and then sieved for 10 min. at $\frac{1}{2}$ phi spacing in a Ro-Tap sieve shaker. Material passing the 44 micron sieve was added to that from the wet sieving. If the sieving retained 95% or more of the sample then pipetting was omitted. Pipetting took place from a 1000 ml settling cylinder set in a constant temperature bath at 25°C. The mixture was stirred mechanically for 20 sec. and then the suspension brought to 1000 ml exactly with distilled water. After 45 sec. of shaking by hand the cylinder was replaced in the bath. Draws were taken by a 25 ml pipette at the following time intervals: 1 min. 43 sec., 3 min. 24 sec., 13 min. 33 sec., 27 min. 17 sec., 1 hr. 49 min., 7 hr. 16 min., and 25 hr. 54 min. representing one phi intervals. Each draw was dried at 105°C in a tared beaker, cooled in a dessicator and then weighed to four decimal places. Thus the results give amount of material between 2mm (-1 phi) and 4 mm (-2 phi) $\frac{1}{2}$ phi divisions for sand size material to 0.063 mm (4 phi), single phi divisions through silt and clay with the break at 0.004 mm (8 phi), and the final point at 0.002 mm (9 phi) or coarser.

Method loss is calculated (not measured) and was found to be greatest in test sampling (not the Generator Lake samples) when the median diameter was close to 44 microns.

e) Suspended sediment analysis: Ashing of the filter paper was conducted in the following manner. The filter paper and contents were placed in a tared porcelain crucible, burned over a bunsen burner, then placed in an ashing oven for 3 hr. at 550°C. The crucible was cooled to 110°C in the oven, then transferred to a dessicator and cooled to room temperature. The final weight was recorded and weight of suspended sediment calculated.

IV. Presentation of results

The dissertation is divided into five chapters, each one of which presents some new material or new interpretations or both. An attempt is made to keep peripheral and historical material pertinent but brief. Some of the peripheral voluminous data is keyed to, but separated from, the main text as Appendices. Where feasible, lengthy description is avoided by generous use of annotated photographs.

a) Measurement systems: In keeping with current trends in scientific journals and government policy the metric system is used. However, some field instruments, government data and contour maps, all circumstances beyond the control of the writer led to original data being in imperial measure. Hence, in keeping with a long standing tradition of dissertations data are included in the original system of measurement. In this way, and only in this way can users adopt a degree of rounding error in translation to metric measure suitable to their own particular investigation and at their own discretion. A number of potential pitfalls are avoided by adopting an original data standard. These include the danger of giving a false impression of accuracy by showing a '100 ft.' measurement as 30.48 m implying accuracy to one centimetre when it is more probably to one foot. This technique has even led to requiring a footnote explaining that a false impression of accuracy was introduced by metric conversion (Embleton & King, 1975, Table 19.2, p. 539). On the other hand, rounding to 30, 61 and 91 m for 100, 200 and 300 ft. contours is difficult to defend as metric map makers do not opt for those particular elevations and would in any case use constant contour intervals. In the two hypothetical cases cited there is a strong probability that the reader could deduce exactly which value was the field original, but it is contended that a field measurement of, say 71 ft. translated as 22 m would leave the reader with some uncertainty as to the original figure and a series of different users would probably deduce slightly different 'original' figures, possibly ranging from 70 to 75 ft. Such a loss may or may not be significant but with the original field figure the reader may judge accordingly. The importance of this point may be brought out if ever a controversy arises which is dependent upon some aspect of measurement which has taken on additional significance as further insights into geomorphic processes occur. In such cases going back to the original measurement could be crucial. Thus a principle of consistency has been followed - that of presenting original data.

b) Editorial Style Manual: For consistency in editorial matters the Geological Survey Guide to Authors (1975 Edition) (Blackadar, Dumych & Griffin, 1975) has been used.

c) Prior Publication: Some material submitted here has already appeared in Government publications and other scientific journals as part of the duties of the writer during data collection and preparation of this dissertation. The writer is appreciative of this use of material being permitted by the dissertation committee and the University of Western Ontario. However, to facilitate reference to this material copies of the three relevant papers are included with the dissertation as Appendix F.

CHAPTER 2

THE PHYSICAL MILIEU OF THE PROGLACIAL ENVIRONMENT

THE FIELD AREA

Baffin Island is the southernmost and largest of the Canadian Arctic Islands with an area of 476 190 km^2. It extends from the sub-Arctic shores of Hudson Strait close to 62°N. to Lancaster Sound, entrance to the Northwest Passage at 74°N. The island also extends from almost 62°W. at Cape Dyer on the Arctic Circle to 90°W. on the western side of the Brodeur Peninsula on the Gulf of Boothia (Figure 1-1). The eastern rim of the island is mountainous facing Baffin Bay and is cut by deep narrow fiords whose heads reach back to the interior upland. Many of the coastal mountains, 915 m-2440 m high, carry glacial ice or semi-permanent snow fields. The Barnes Ice Cap (Figure 2-1) is topographically and glaciologically distinct from these as it lies westward of the mountains and lies across the consequent stream valleys of a dissected upland which slopes gently toward Foxe Basin in the west. The ice cap was considered to be a Wisconsin remnant by Goldthwait (1951b) and Hooke (1976) has concluded from oxygen isotope ratios that Pleistocene ice is still present at the base. The ice has shown a gradual net decline throughout the past 4500 years with which this study is concerned.

Bostock's map (1970) of the Physiographic Regions

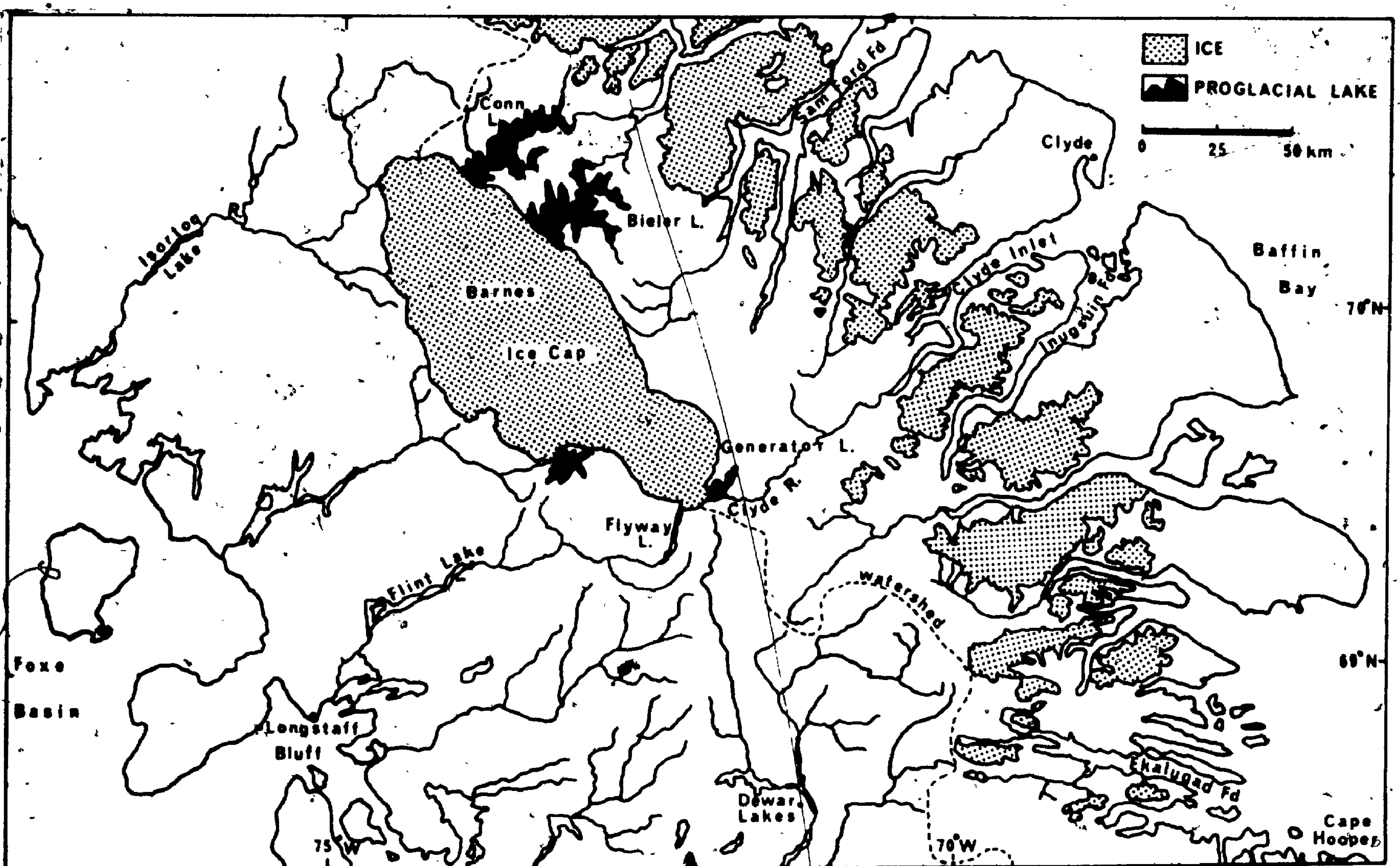


Figure 2-1: Map of central Baffin Island showing the location of the Barnes Ice Cap, the proglacial lakes, the regional watershed, east coast fiords and the existing mountain glaciers and snowfields.

of Canada shows the field area in the Baffin Upland division of the Davis Region of the Canadian Shield. Bird (1967, p. 61) states that "the backbone of Baffin Island is a complex upland with a subdued upper surface". This description is of Bird's "Baffin Surface" which he identifies in its less dissected parts at an elevation of 600-700 m (2000-2300 ft.) with a relief of about 60 m (200 ft.). The Generator Lake basin is included in the area so designated, but has somewhat greater relief 275 m (830 ft.) and generally lies below the 2000 ft. contour with the highest land summit lying between 2650 ft. and 2675 ft. That portion of the Barnes Ice Cap from which snowmelt drains into the basin, however, rises to approximately 2900 ft.

The Generator Lake basin at the southeastern margin of the Barnes Ice Cap lies in the upper reaches of a consequent valley and covers an area of 540 km^2 of which 130 km^2 are covered by part of the Barnes Ice Cap. Generator Lake itself has an area of 40 km^2 and is up to 60 m deep. An earlier and deeper phase of the lake covered 120 km^2 of the basin. The basin is elongated in a northeast to southwest direction reflecting the shape of the preglacial consequent valley and extends from 69°48'N. and from 71°25'W. to 72°06'W. (Figure 2-2). The watershed between Foxe Basin drainage and Baffin Bay forms the western boundary of the Generator Lake basin.

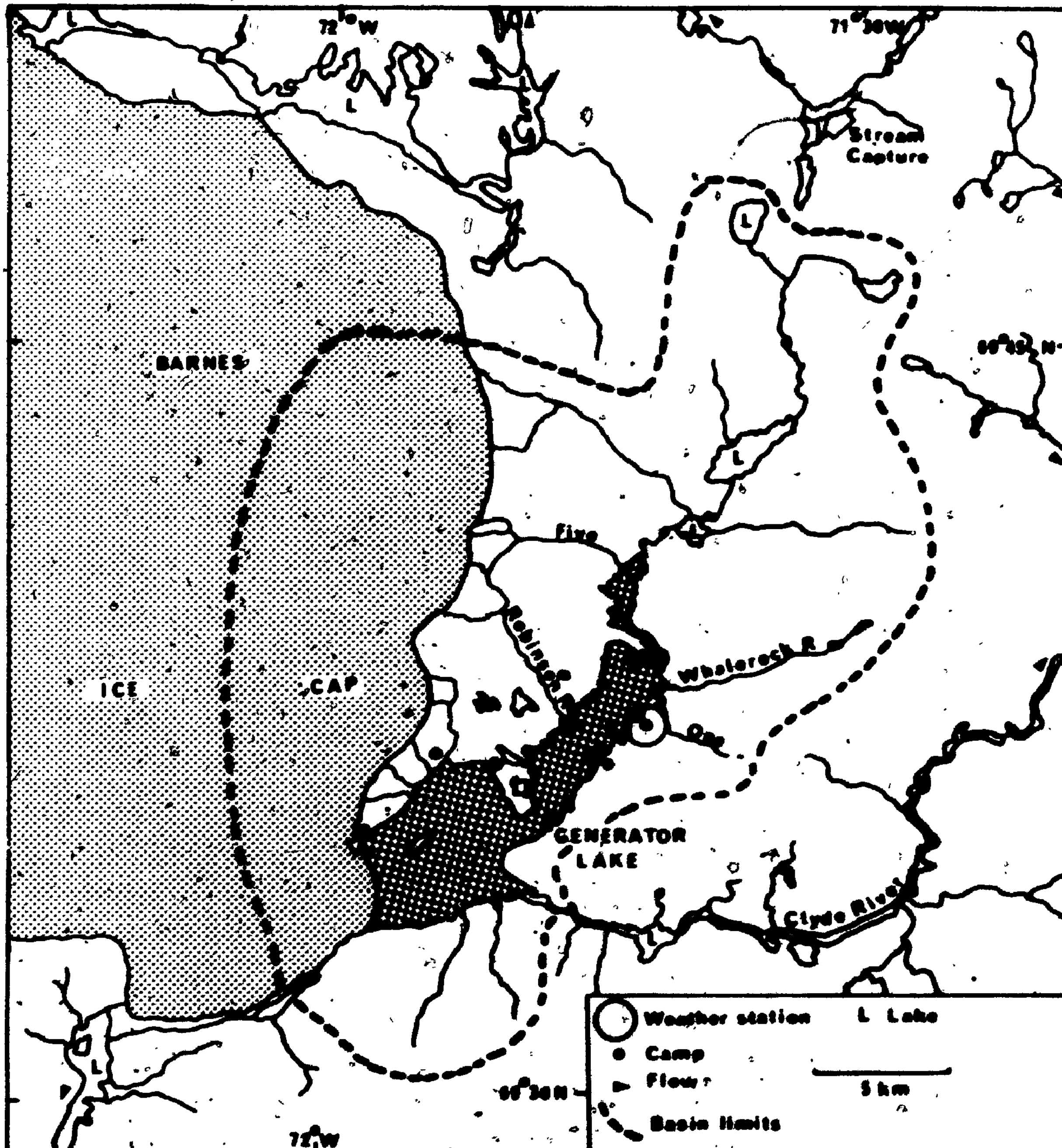


Figure 2-2: Map of Generator Lake basin showing principal streams, weather station and camp locations.

SCIENTIFIC HISTORICAL PERSPECTIVE

Somewhat paradoxically Baffin Island was discovered early by Martin Frobisher in 1576, was revisited by William Baffin early in the seventeenth century, but remained little known and inadequately mapped until after the Second World War. Of course, after the voyages of Frobisher and Baffin many other vessels sighted Baffin Island, and landings were made by whaling crews or by those searching for the Northwest Passage. Furthermore, several early exploratory-scientific expeditions visited Baffin Island, including those of Franz Boaz to Cumberland Sound and the west shore of Davis Strait in 1883-4 (1884), and R. Bell for the Geological Survey of Canada to southern Baffin Island in 1897 (1898). Knud Rasmussen's Fifth Thule expedition visited northern Baffin Island in 1921-4 and J. M. Wordie's northeast Baffin Island expedition worked there in 1934. All involved feats of endurance as part of the price or reward (depending upon one's viewpoint) of the undertaking. Post-war investigations concentrated more exclusively on scientific endeavours usually with some form of air support, though rigorous field conditions were by no means eliminated.

The discovery, on aerial photographs, in 1948 of Prince Charles and Air Force Islands in Foxe Basin just off the west coast of Baffin Island concluded the purely exploratory phase (Fraser, 1953). The 1949 voyage of the 'Nauja' marked the transitional stage in scientific exploration. It was primarily scientific (geological) but with exploratory overtones (Burns, 1952). Nevertheless, Blackadar (1970, p. 6) describing conditions before Operation Admiralty, a geological reconnaissance carried out by the Geological Survey of Canada in northwestern Baffin Island in 1963, indicated that the island "....one of the largest in the world, remained geologically unknown".

Settlements on the island are few, with the earliest based on a native fur trapping economy managed by the Hudson Bay Company at places such as Arctic Bay, Pond Inlet, Clyde, Broughton Island and Pangnirtung (Figure 1-1). The post-war recognition of the military strategic significance of the whole of the Arctic brought more and bigger settlements to Baffin Island, notably the Distant Early Warning Line (DEW Line) sites and Frobisher Bay which grew from its establishment in 1942. Concomitant with this development was improved access to many parts of the island as a result of construction of landing strips with ancillary weather stations and the installation of sophisticated communications systems. With these developments came an increased government presence augmenting the original 'multipurpose Mounties' of the R.C.M.P. Almost all settlements are now less than six hours flying time from Montreal. There are as many as eight jet flights a week to Frobisher Bay, and at least weekly service beyond, which is a vast change from the once-a-year sea-lift.

Scientifically there has been a rapid increase in knowledge since the first trimetrogon air photograph coverage was flown in 1948. These photographs, a very considerable scientific data source, were later augmented by full vertical coverage at a scale of 1:60 000 and recently by localized special photography of areas of particular interest. The Rand series of physiographic reports of Baffin Island indicated the change of pace and scale of scientific examination made possible by air photograph interpretation (Rand, 1963). The island is among the few areas of the world where many of the original scientific investigators are not only still alive but are still working.

The Generator Lake basin was essentially unknown until the 1950 Arctic Institute of North America expedition. The only known previous sighting by a scientific group was by two members of the 1934 J. M. Wordie expedition to Baffin Island (Wordie, 1935, p. 331). P. D. Baird and M. Ritchie climbed Pioneer Peak and saw the then unnamed Barnes Ice Cap. They returned in 1950 with Baird leading the expedition (Baird, 1950). Only part of this group's effort was devoted to the Generator Lake area, and much of this part concerned glaciological and related meteorological observations (Ward, 1952; Orvig, 1951). These were augmented by Goldthwait's brief but detailed study of moraine formation at the ice cap margin (Goldthwait, 1951a), and by a short lichenological investigation by Hale (*in* Baird, 1950).

In 1950 only inadequate and often inaccurate maps were available for Baffin Island and the air photograph coverage was trimetrogon only which added to the considerable logistical problems of scientific investigation in a remote environment. The degree of sophistication achieved particularly by Orvig, Ward and Goldthwait was remarkable.

In 1961 the Geographical Branch of the Canadian Department of Mines and Technical Surveys (now Energy, Mines and Resources) began an intensive investigation of the physical geography of north-central Baffin Island. The study area, centred on the Barnes Ice Cap, was considered critical in the examination of recent glacial history, since Goldthwait (1951b) had identified the Barnes Ice Cap as a vestigial remains of the Wisconsin Ice Sheet.

From these studies maps of glacial features were produced (Ives & Andrews, 1963; Sim, 1964), as well as several important papers: on lichenometry as a means of dating glacial landforms (Andrews & Webber, 1964); on evidence of recent extensive thin ice cover over much of northern Baffin Island (Ives, 1962); on glaciometeorological data from the Barnes Ice Cap crest to examine the mass balance (Sagar, 1966) and on till fabric analyses of cross-valley moraines to examine process of formation (Andrews & Smithson, 1966).

This intensive physical geography program contrasted with the extensive systematic geological inventory of the island carried out by the Geological Survey of Canada during Operation Admiralty in 1963 (northwest Baffin), Operation Bylot in 1968 (north-central Baffin) and Operation Penny in 1970 (south-central Baffin). The former two had Quaternary research components.

The writer joined the Geographical Branch Baffin Island Program in 1963 as a graduate assistant and later became a field officer; research for the present study began with a preliminary reconnaissance in August 1965 at Generator Lake and continued during part of the 1966 summer. In 1967 departmental reorganization led to transfer of the writer to the Quaternary Research and Geomorphology Division (subsequently renamed Terrain Sciences Division) of the Geological Survey of Canada. Research was resumed in 1968 and 1969 and became increasingly oriented toward production of a dissertation. As such a goal was not specifically defined in the beginning, some results are already published (Barnett, 1967; Barnett, Forbes & Whytock, 1970 and Barnett & Holdsworth, 1974; all included as Appendix F) and the emphasis has changed through time.

Much additional research on Baffin Island has been conducted by personnel from the Institute of Arctic and Alpine Research (INSTAAR) since the fieldwork reported here but none has taken place in the Generator Lake basin although several regional syntheses have included the field area, e.g. Dyke (1974).

GEOLOGY AND GEOMORPHIC SETTING

I Bedrock Geology

Baffin Island forms the northeastern rim of the Canadian Shield which is an immense structural block of resistant Precambrian rock. Minor outcrops such as the Precambrian sedimentary Iron Formation occur in scattered pockets along the long axis of Baffin Island (Jackson, 1969). Some more extensive Paleozoic rocks outcrop in depressions in the shield such as Foxe Basin in which limestones predominate (Trettin, 1969).

As the Generator Lake basin is underlain solely by the Precambrian basement it represents a constant in the model. It should be noted that the southernmost watershed is formed by a narrow ridge of Iron Formation first recorded by Kranck (1953), and confirmed by Jackson (1969, p. 172 & 175).

The bedrock basement is characterized by migmatites, gneisses and granitoid rocks (Jackson, 1969, p. 173) the history of which is not of concern here. The landscape created on such a base is one of open rounded terrain, often rubble covered, becoming felsenmeer in places, with bedrock rarely far from the surface. Aspects of interest are that coarse-grained crystalline rocks primarily yield coarse fraction particles on weathering. Locally jointing or fracturing is strong sub-parallel with the ground surface and spalling was observed (Figure 2-3) in association with the main abandomed shoreline of Generator Lake.

Figure 2-3: Outcrop at the 76 m shoreline showing fracture patterns which yielded boulders of similar thickness in the ice pushed ridges.
(GSC Photo 168111)

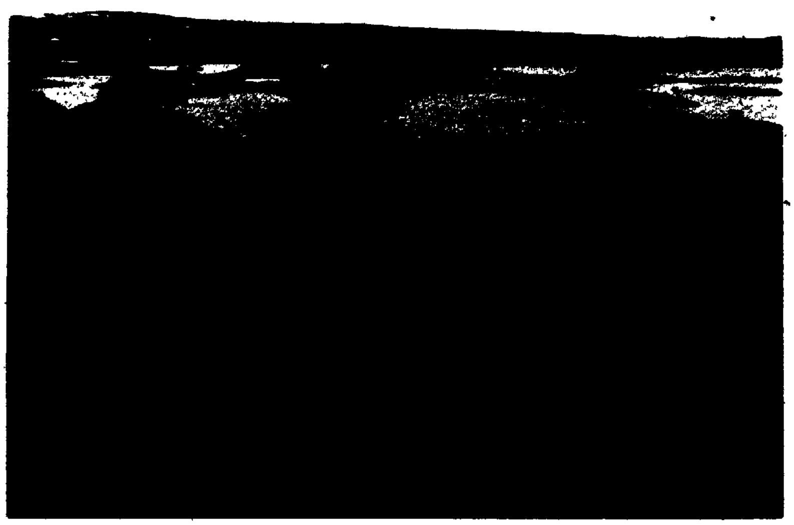


Figure 2-4: General view of exposed sublacustrine moraines showing arcuate shape, asymmetry of profile and spacing. The ice margin is to the right.
(GSC Photo 203028)

Boulders forming ice pushed ridges (see Chapter 4) exhibit a characteristic thickness reflecting the jointing pattern.

II Surficial Geology

Very coarse bouldery ground moraine mantles much of the exposed part of the lake basin but bedrock is usually very close to the surface. Indicators of ice flow are present but not abundant showing flow just north of east. Locally, water action has concentrated sand size material into deltas associated with former and present phases of Generator Lake. Occasional beach deposits, ice pushed ridges and a glacio-lacustrine terrace are associated with the highest shoreline. Multiple moraines are conspicuous features of the lake basin area now drained (Goldthwait, 1951b; Andrews, 1963). These landforms (Figure 2-4) and other less conspicuous features were described earlier elsewhere (Barnett, 1967). A comprehensive treatment of the moraines, using the term sublacustrine moraines, was published by Barnett & Holdsworth (1974) and is now covered in Chapter 4.

As the eastern margin of the Barnes Ice Cap retreated more of the Generator Lake basin was exposed. The lake tended to become both larger and deeper as retreat was down the regional slope. In addition the area of exposed land surface within the basin became larger. Variations on this trend are dealt with in detail in context.

The present glacier margin is a combination of ramps and cliffs at the lake-ice interface (Figure 2-2 and Figure 2-5) and in 1969 was going through a phase of increased cliffing due to accelerating retreat at a zone of strong transverse crevasses which have remained quasistationary since the earliest record on the 1948 air photographs. In contrast, the ice-land boundary in the basin is a tapering edge which is difficult to identify precisely. This is because of the concentration of debris in the shear zone, making a clear distinction between fresh ground moraine and heavily debris-laden ice difficult. It is in these most recently deglacierized areas that the greatest amount of fine particles occur which subsequently find their way into streams and later onto the lake bottom itself. The ice cap margin is a major source of material for fluvial transport. As glacial ice melt maintains the flow of streams on the north side of the lake basin after the main snowmelt, greater competence is available to transport the additional material than is available from streams on the south side of the lake basin. The result is asymmetry of both supply and of deposition. This is very strikingly shown by the size and location of raised deltas.

III <u>Topography and Drainage</u>

The land surface within the lake basin is characterized by low bedrock domes mantled with very coarse bouldery till (ground moraine). They typically show relief between 75 m and 150 m. Although the dome summits are quasi-accordant in elevation, this small drainage basin (540 km^2, including

Figure 2-5: Aerial oblique photograph of ice-cored surge moraine (foreground), ice cliff and ice ramps. (GSC Photo 168064)

the ice) does not fit typically into Bird's macro-scale topographic description of the Baffin Surface as having relief of about 60 m (200 ft.) (Bird, 1967).

Second or third order streams which drain the interdome areas are typically small, ungraded and often braided with occasional shallow lake expansions along their courses. Drainage density of the basin is extremely low, averaging only 0.951 km/km^2 based on the drainage lines shown on the 1:50 000 topographic map, including those shown on the Barnes Ice Cap and adopted curvilinear routes through lakes. This value is significantly less than the 1.5 km/km^2 considered by Scheidegger (1970, p. 24) to be among the lowest values to be found. The present condition is indicative of a drainage net which is not fully integrated and in fact some areas of the map lack drainage lines but carry shallow lakes without significant connecting channels. Details of run-off characteristics and sediment load of five integrated streams are presented in Chapter 4.

The streams drain into Generator Lake, the dominant hydrological feature of the basin, and out as the headwaters of the Clyde River which drains eastward to Clyde Inlet and Baffin Bay (Figure 2-1). Formerly preglacial drainage was to the southwest into the area now occupied by Foxe Basin. This former route was substantiated by gravimetric profiles obtained through the ice cap by Littlewood in 1950 (1952, p. 123). Further support is offered by the present sublacustrine topography which shows a slightly meandering gradually deepening lake as the ice cliffs are approached (Barnett, Forbes & Whytock, 1970, Figure 4). At the head of the preglacial valley the stream was beheaded by stream capture (Figure 2-2) by a south bank tributary draining northeastward into the Sam Ford system. This stream is exploiting a zone of apparent structural weakness running northwest-southeast for approximately 50 km. Jackson (pers. comm., 1971) found no evidence of faulting along this line. A reconstruction of Arctic drainage development by Bird (1967, Chapter VII) is compatible with the above interpretation, but his Figure 23 (p. 82) suggests the further possibility that the preglacial stream may have been at a much earlier time a subsequent tributary to a southeast draining consequent stream through Foxe Basin.

The ice dam retaining Generator Lake and the whole crestline of the Barnes Ice Cap constitute part of the major drainage divide between the Arctic Basin and Hudson Bay (Rivers, Lakes & Ice - National Atlas of Canada Special Interim Series, 1968). Disappearance of the ice cap would result in the drainage divide being relocated much closer to the western flank of the coastal mountains.

IV Geometry

Geometry is defined by the Oxford dictionary as the science of properties and relations of magnitudes in space.

Geometry has affected both process and response within the lake basin. Bunge (1962) made a strong plea for greater attention to geometry in geographical research. Haggett

(1965, p. 14) in assessing geography places geomorphology in overlapping sets including the geometrical sciences and both Bunge and Haggett conclude that insufficient emphasis has resulted despite the prominent role played by geometry in the classical Greek concept of geography.

If one considers the Generator Lake basin as an irregularly shaped container with one moveable side (the ice cap) then the geometry of the container and the position of the moveable side control the volume of lake water. The geometry is also a prime determinant of the position and basal morphology of lacustrine and sub-lacustrine depositional landforms. Lake fetch and the form of the heat reservoir beneath the semi-permanent lake ice cover are also influenced by the basic geometry.

The gross geometry of the basin is the result of past orogenic-lithologic-climatic events. The sum of bedrock geological events is considered as a geometrical constant and the climatic events are variables. Basic dimensional data of the drainage basin are shown in Table 2-1.

PRESENT CLIMATE

I General

Generator Lake lies in the upper reaches of a consequent valley. Taking the geometry of the valley as given then the dominant variable in the geomorphic system is climate, for the geometric variable of the ice dam is controlled either directly or indirectly by climate (considering the glacial

TABLE 2-1

THE GEOMETRY OF GENERATOR LAKE BASIN

	Total Basin	Former Lake (76 m)	Present Lake
Area	540 km^2	120 km^2	42 km^2
Length	31.5 km	26.5 km	13.2 km
Depth (max.)	-	115 m	60.5 m
Width (max.)	32 km	8.63 km	4.57 km
Elevation (max.)	951 m (ice) 678 m (land)	477 m	402 m
Relief (max.)	609 m ice-lake bottom 336 m land-lake bottom 276 m land-lake surface		
Volume	-		

Drainage Density: 0.951 km/km^2 total basin (Barnes Ice Cap + deglaciated basin)

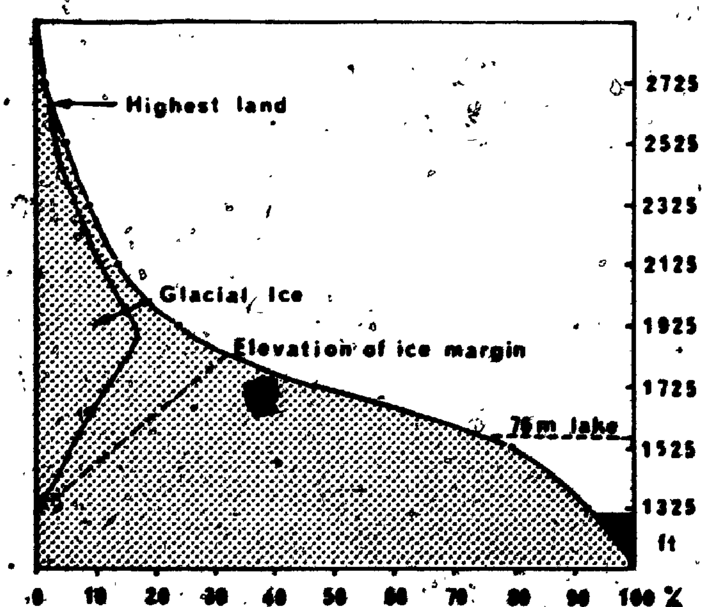
201 km in 130 km^2 1.55 km/km^2 (ice only)

312.5 km in 410 km^2 0.762 km/km^2 (land only)

Area Ratio:	
ice to land + lake	1:3.15
lake to land	1:8.76
lake to total basin	1:12.87
ice to total basin	1:4.15

Hypsometric Data:

Basin area	
above 2725 ft.	1.10%
2725 -2525	3.31%
2525 -2325	3.98%
2325 -2125	4.75%
2125 -1925	9.89%
1925 -1725	22.60%
1725 -1525	33.23%
1525 -1325	13.16%
below 1325	7.95%
Total	99.97%



surge (Løken, 1969; Holdsworth, 1973b) as being an indirect climatic response).

The time span under consideration is somewhat less than 5000 years though little direct information on climate is available for this time span. Certain inferences may be drawn about the severity of the climate relative to that prevailing today but even data on the climate of the present proglacial environment are negligible. Evidence from the fields of palynology, botany and glaciology yield climatic inferences of very limited range and debatable accuracy.

II Existing Records

North-central Baffin Island has a sparse recording network of meteorological stations and even these have comparatively short records. Clyde on the east coast has the longest record (September 1947 - present). Lonstaff Bluff (west coast), Dewar Lakes (central) and Cape Hooper (east coast) (Figure 2-1), all three maintained on contract for the Atmospheric Environment Service, Environment Canada, only offer records from the late fifties. Of the four stations, Dewar Lakes is the closest to the Generator Lake basin in distance, altitude and topographic setting, but Løken & Sagar (1968, p. 288, Table III) have shown that winter precipitation amounts are greater by three or four times at the ice cap. Precipitation variability for July and August 1969 was also great but with opposing trends: 5.4 times as much rain fell at Dewar Lakes as at Generator

Lake in July, but almost six times as much rain fell at Generator Lake as at Dewar Lakes in August. It has long been recognized that the Barnes Ice Cap creates its own local climate (Sagar, 1966; Løken & Andrews, 1966; Løken & Sagar, 1968), but lack of more than a few months weather records from the ice cap (Sagar, 1966) has led to comparison of these partial records with available records from the closest weather stations to see if parallel patterns emerge. If so, then it would be reasonable to extrapolate from the longer records to the ice cap. Such an undertaking is speculative at best for the Clyde and Cape Hooper stations are east of the coastal mountains, which are often more than 1000 m high. Such stations are often subject to local summer fogs blowing off the ice-choked coastal waters of Baffin Bay. Longstaff Bluff on Foxe Basin often suffers from coastal fog as late ice is also characteristic. Hence, regionally prevalent weather patterns may be masked by local conditions at any or all of these stations, and the east coast stations may be not infrequently in a weather system confined to Baffin Bay.

A further unknown factor is that of operator error or omission at the contract stations. Personal discussion with some observers indicate that because maintenance of meteorological recording equipment was not their prime responsibility, on occasions, when conflicting demands were made on the observer's time, the weather instruments were given second priority. As a specific example, the build-up of hoar frost on the cups of the anemometer at Dewar Lakes has occasionally led to low readings in strong winds.

III The Regional Climate

Despite paucity of long term accurate data, the broad climatic pattern can be sketched with confidence. In terms of the Köppen system the Generator Lake basin is ET (tundra) and the adjacent ice cap is EF (perpetual frost), in both cases the 'E' indicating that the warmest monthly mean temperature is less than 50°F or 10°C. Annual precipitation is estimated to be about 12 inches (30.5 cm) in water equivalent (Atlas of Canada, 1957, Plate 25), including perhaps 60"-70" of snow, often concentrated during the winter-summer and summer-winter transition periods.

Although as Stamp (1966) concludes, 'desert' is a difficult term to define accurately, many definitions would accommodate the area in question, particularly in qualitative terms such as "inhospitable" and "an area of very sparse and stunted vegetation". However, it would be totally misleading to use the word desert as meaning 'arid' for although the total annual precipitation is low, ground moisture is abundant to excessive during the brief, but vigorous growing season. There are two basic reasons, one is the concentration of the total run-off into perhaps 8-10 weeks and secondly, the presence of continuous permafrost (Brown, 1967) which prevents infiltration below the shallow active zone (<1 metre). In areas where the surficial cover is less than one metre thick, run-off is correspondingly greater.

A characteristic of deserts is their extreme variability of precipitation. Table 2-2 shows such characteristics for both Dewar Lakes and Longstaff Bluff (1958-1969) for the months May to September inclusive. The greatest variation was between August 1958 and August 1964 at Dewar Lakes, the latter being >568 times the amount of the former. Comparable figures for Longstaff Bluff are >362 times (July 1963-July 1969). Even based on the five month total at Dewar Lakes 3.86 times as much precipitation fell in 1964 as in 1965 and 4.59 times as much in 1967 as in 1966 at Longstaff Bluff. This variability might be expected to have considerable influence on the intensity of fluvial processes, but when considered carefully, this may not be so. Firstly, the geomorphic effect of sporadic precipitation in true summer tends to be masked or overshadowed by the main snow melt run-off and secondly, much of the July-August precipitation (the main "rain" months) occurs on single days but rarely as heavy rain. Hence, moderate amounts at low or even moderate intensity can be accommodated by the drainage system without dramatic effect. This is particularly so when the snowmelt period is brief and vigorous as this tends to "flush out" much of the loose material in the drainage system. Fluvial activity in a drainage basin containing glacial ice often is greater on a warm, sunny day than following measurable rainfall.

TABLE 2-2

PRECIPITATION

Year	Dewar Lakes May	Jun	Jul	Aug	Sept	rain	pptn	Longstaff Bluff May	Jun	Jul	Aug	Sept	rain	pptn
1958	NR	.30	.55	TR	.53		1.38	NR	.05	1.68	.13	.84		2.70
		.20	.15		.20	.55			.05	.48	.10	.53	1.16	
1959	2.11	.66	2.08	.58	.70		6.13	.23	.06	.48	1.14	.85		2.76
	.80	.17	.73	.42	.70	2.82		.10	.05	.17	.57	.28	1.17	
1960	.75	.51	.59	.19	1.98		4.02	.35	.15	.57	.21	1.53		2.81
	.35	.20	.39	.17	.44	1.55		.30	.08	.17	.09	.39	1.03	
1961	.91	.13	.26	3.35	1.18		5.83	.81	.72	.01	1.98	1.54		5.00
	.40	.10	.26	.80	.34	1.90		.30	.55	.01	1.05	.44	2.35	
1962	.73	.50	1.28	1.17	1.62		5.30	.75	.05	2.55	2.84	2.26		8.45
	.33	.40	.81	.27	.41	2.22		.20	.05	1.40	.78	.99	3.42	
1963	.87	.27	1.11	1.37	1.28		4.90	.43	.27	3.62	.84	.96		6.12
	.31	.15	.60	.53	.24	1.83		.14	.10	1.89	.37	.30	2.80	
1964	.32	.98	3.33	5.68	2.29		12.60	.60	1.13	1.98	.50	1.03		5.24
	.10	.22	1.27	1.71	.83	4.13		.18	.37	1.19	.18	.35	2.27	
1965	.15	.19	.53	.93	1.46		3.25	.77	.56	2.75	1.69	1.44		7.21
	.11	.13	.43	.60	.35	1.62		.27	.49	1.69	.39	.30	3.14	
1966	.15	.46	1.68	.87	.76		3.92	.01	.84	.57	.43	.79		2.01
	.15	.14	1.43	.40	.40	2.52		.01	.49	.57	.14	.40	1.61	
1967	.85	1.56	2.07	3.74	.99		9.21	.60	1.19	2.55	2.55	2.34		9.23
	.26	.34	.72	1.24	.41	2.97		.12	.33	1.10	.65	.38	2.58	
1968	.17	1.39	1.37	1.25	2.80		6.98	.77	.34	.77	1.23	1.38		4.49
	.12	.54	.30	.40	.86	2.22		.21	.16	.11	.37	.65	1.50	
1969	2.32	.14	.54	.09	1.00		4.09	1.08	.03	TR	.03	1.13		2.27
	.50	.09	.46	.09	.40	1.54		.30	.02	TR	.03	.30	.65	
Total	9.33	7.09	15.39	19.22	16.59		66.24	6.40	5.39	17.53	13.57	16.09		58.35
$\bar{X}$	.848	.591	1.283	1.602	1.383	$\bar{X}$	6.022	$\bar{X}$ .582	.449	1.461	1.131	1.341	$\bar{X}$	4.863

NR NO RECORD TR TRACE ALL DATA WATER EQUIVALENT (IN.)

IV Local Climate

No year round records are available for the Generator Lake basin and summer data are sporadic. A complete record was obtained for the period from July 5 to the end of August 1969. This included continuous thermograph, hygrograph and barograph records plus precipitation, total run of wind, cloud cover, maximum and minimum temperatures on a 12 hourly basis at the station shown on Figure 2-2. Lake water level and lake water temperature close to shore were also recorded. Subjectively the two months were atypically good with only 0.1" rain in July and 0.595" in August (0.44" fell on August 29) and the mean temperature for July was 44.4°F and for August 45.5°F. Support for this subjective assessment is apparent when examining the sequence of maps showing deviations from average temperatures in the Monthly Record of the Department of Transport (now Atmospheric Environment Service). The maps are national in scope and therefore variable in accuracy due to the variations in density of stations and length of records at each. Only the months of June, July, August and September were examined for central Baffin Island for 1946-1973 inclusive and interpolation to the nearest 1°F was adopted. The results are presented in Table 2-3. Both July and August 1969 have positive anomalies of 3° and 6°F respectively, the latter being the second highest of the decade. However, the 1964-1973 decade was the coolest one since 1946 for this area. Although limitations of the data must be

TABLE 2-3

MEAN MONTHLY TEMPERATURE ANOMALIES (°F) - CENTRAL BAFFIN ISLAND

	1946	1947	1948	1949	1950	1951	1952
June	-2	1	2	4	2	2	3
July	-2	-1	5	1	1	N	1
Aug.	-2	-1	2	1	N	-2	-1
Sept.	-1	-1	2	N	N	2	-1
Net	-7	-2	11	6	3	2	2

	1953	1954	1955	1956	1957	1958	1959
June	1	-1	2	-1	-1	1	-1
July	-1	1	2	-1	3	2	N
Aug.	2	5	1	1	-1	-2	N
Sept.	-1	1	1	-1	4	1	N
Net	1	6	6	-2	5	2	-1

	1960	1961	1962	1963	1964	1965	1966
June	2	1	-1	-2	-1	-1	-4
July	2	4	3	-2	-2	-1	-1
Aug.	2	-3	1	N	N	-4	7
Sept.	2	-3	N	-3	2	N	2
Net	8	-1	3	-7	-1	-6	4

	1967	1968	1969	1970	1971	1972	1973
June	-4	-4	-2	-4	N	-8	N
July	-1	-6	3	-6	N	-6	-2
Aug.	-2	-1	6	-2	-2	-4	2
Sept.	-4	2	N	2	N	-3	3
Net	-11	-9	7	-10	-2	-21	3

N Normal

25 year Record Net: 9°F 0.36°F/yr

28 year Record Net: -11 -0.39°F/yr

Decade Values

	Net Sum	Mean
1946-55	28°F	2.8°F/yr
1951-60	29°F	2.9°F/yr
1956-65	Nil	Normal
1961-70	-31°F	-3.1°F/yr
1964-73	-46°F	-4.6°F/yr

stressed it is clear that a cooling trend has set in, at least locally.

The mean wind speed in July was 11.28 km/hr. and in August increased to 14.69 km/hr. Air frost occurred on July 8 and did not reoccur until August 18, a frost-free period of 36 days. Mean cloudiness for July was 0.608 based on 53 observations and this somewhat remarkably decreased to 0.585 in August (62 observations). There was no recorded overcast between August 8 and 21 inclusive. The data are summarized in Table 2-4. Non-systematic observations of daily maximum and minimum temperatures were made on the north side of the lake closer to the ice cap from July 9 to August 23, 1970 by Dr. G. Holdsworth of Environment Canada. The mean daily values for July and August were 41.0°F and 37.8°F, both less than 1969 values and showing the more expectable decline from July to August. No air frost occurred in the first 21 days of record.

VEGETATION

I General

The following remarks on present vegetation are based primarily on casual observations and without the benefit of formal botanical training. However, some consideration is pertinent as the detailed chronology of the lake basin is dependent upon detrital vegetation which yielded radiocarbon dates. Samples of the fossil plant material, particularly the mosses, were kindly examined by Dr. Marian Kuc

TABLE 2-4

METEOROLOGICAL DATA SUMMARY, GENERATOR LAKE 1969

	July (date)	August (date)
Temperature, mean	44.4°F (6.9°C)	45.5°F (7.5°C)
Temperature, max.	67.8°F (19.9°C) (28)	64.9°F (18.3°C) (3)
Temperature, min.	28.8°F (-1.8°C) (7)	29.7°F (-1.3°C) (23)
Precipitation	0.1 in. (2.54 mm)	0.595 in. (15.1 mm)
Maximum fall	0.04 in. (1.02 mm) rain (17)	0.46 in. (11.7 mm) rain (29)
Wind run:		
12 hr mean	131.33 km	141.35 km
12 hr max.	320.3 km (10)	340.8 km (16)
12 hr min.	12.1 km (7)	38.0 km (5)
Prevailing wind	NE (19 % of observations, 2 per day)	NE (24 % of observations, 2 per day)

formerly of the Geological Survey of Canada and his bryological reports with interpretive comments are included in Appendix C.

The Atlas of Canada (1957, Plate 38) shows the vegetation of the lake basin to be primarily "stony-sedge-moss-lichen tundra" apart from the area of vegetation-free ice. Obviously the boundary between the two lies within the basin but is not coincident with the land-ice boundary because of the time lag in colonization of plant life in newly exposed areas. Furthermore, not all plants are equally hardy in establishing themselves quickly, hence the sedge-moss-lichen association is not optimally developed close to the ice. There are some Arctic willow (<u>Salix arctica</u>) in the basin but they are few, small and have a mat form flush with the ground surface. Lichens are frequently the first plant life to colonize a freshly deglaciated area, usually on a bare, fairly smooth rock surface. The progression then appears to be from lichens to mosses, to sedges and then willow.

II <u>Lichens</u>

The greatest significance of ~~these lowly plants~~, particularly <u>Rhizocarpon geographicum</u>, is their potential for establishing a chronology to be discussed in the next chapter. However, not all types are good chronological indicators. Species identified but not used, included <u>Umbilicaria</u> and <u>Buellia</u>.

III Palynology

Initial attempts to collect modern airborne pollen were by leaving a series of small glass slides coated with glycerine open to the air at 65 cm above ground. The purpose was for comparison with any fossil pollen found in either the lake bottom sediment or in association with detrital vegetation from deltaic sands. Unfortunately, the collecting medium used, glycerine jelly, proved unsatisfactory due to rapid hardening in the very cool environment. The writer is indebted to Dr. G. Holdsworth of Environment Canada for subsequently collecting two pond vegetation samples which have yielded modern pollen and an ice cap sample of concentrated fines which has also yielded exotic pollen. Palynological reports appear as Appendix B.

SUMMARY

The foregoing outlines the location and essential nature of the environment examined and emphasizes the relative paucity of quantitative data on even such routine parameters as mean annual air temperature, or annual precipitation. The next step is to consider the chronology of the environment and compare the methods for determining it.

CHAPTER 3

CHRONOLOGY OF THE PROGLACIAL ENVIRONMENT

REGIONAL CHRONOLOGY: Cockburn to Present

I General

For a specific process-response model to be meaningful it is necessary to place it in context. This must be not only in terms of time but in terms of the regional environment beyond the boundaries of the research area. In this way it will be made clear that ice-dammed lakes have been common features of the Baffin Island landscape for at least the last 8000 years and that they have changed frequently in location and size as the ice margin fluctuated. A summary of regional events is given in Table 3-1 and localities on Figure 3-1. Figure 3-2 shows former lake locations.

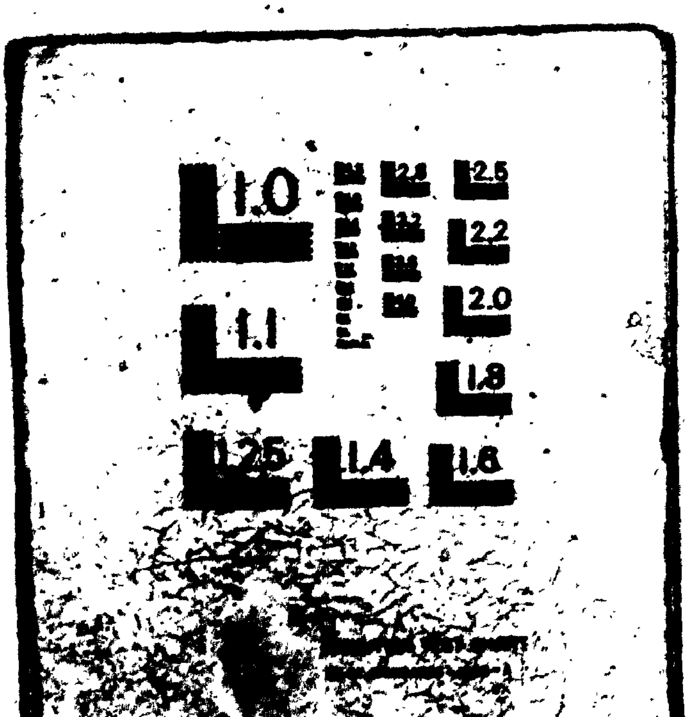
Until recently, present day ice masses were usually classified as being formed of either 'polar' or cold ice and temperate or alpine ice (Sharp, 1960, p. 25). The recent interest in surging glaciers has led to the recognition of a sub-polar category (Schytt, 1974, p. 305) which recognizes a polar ice margin but a temperate core area where the ice thickness is sufficient to reach pressure melting point. The Barnes Ice Cap falls into this intermediate category (Holdsworth, pers. comm., 1976) and Boulton (1975) concluded that this type is a good

TABLE 3-1

REGIONAL CHRONOLOGY - BAFFIN ISLAND AND ADJACENT AREAS

Name	Time Range (B.P.)	Author	Primary Basis	Ice Characteristics
Cockburn Phase	9000-8000	Falconer et al. (1965)	C14 (Shells)	Moraine building
	8400-8000	Andrews & Falconer (1969)	C14 (Shells)	Moraine building
Greenland Ice core	8300	Dansgaard et al. (1969)	O18 (Ice)	Minor cooler interval
Ekalugad Moraine	8160-7480	Andrews et al. (1970)	C14 (Shells)	Moraine building
Tyrrell Sea	8000-7000	Lee (1968)	C14 (Shells)	Rapid retreat
Deglaciation Fiord Heads (Inugsuin, Clyde)	8000	Barnett; Løken (1965)	C14 (Shells)	End of moraine building
Penny Ice Cap	7800	Dyke (1974)	C14 (Shells)	Separation from Northern Baffin Ice Cap
Keewatin Ice Divide	7000	Lee (1959)	C14 (Shells)	Disintegration
Isortoq Phase	7000	King & Buckley (1967)	C14 (Shells)	Moraine building
	6700	Andrews (1966)	C14 (Shells)	Moraine building
Central Labrador-Ungava	6600-6400	Prest (1969)	C14 (Peat)	Disintegration
Sam Ford Fiord	6200-6100	Smith (1966)	C14 (Shells)	Moraine building
Ottawa Islands Strandline	5300-4700	Andrews & Falconer (1969)	C14 (Shells)	Stable/Readvance
Flint Phase	5000-4700	Andrews (1970)	C14 (indirect)	Readvance
Beta Strandline	3700	Andrews (1970)	C14 (indirect)	Stillstand
[illegible]nk Phase	1700	Andrews & Webber (1964)	Lichenometry	Advance
Barnes "surge"	700	Løken & Andrews (1966)	Lichenometry	Advance
	50	Holdsworth (1973)	Glaciology	
Lewis Phase	250	Andrews & Webber (1964)	Lichenometry	Advance

2



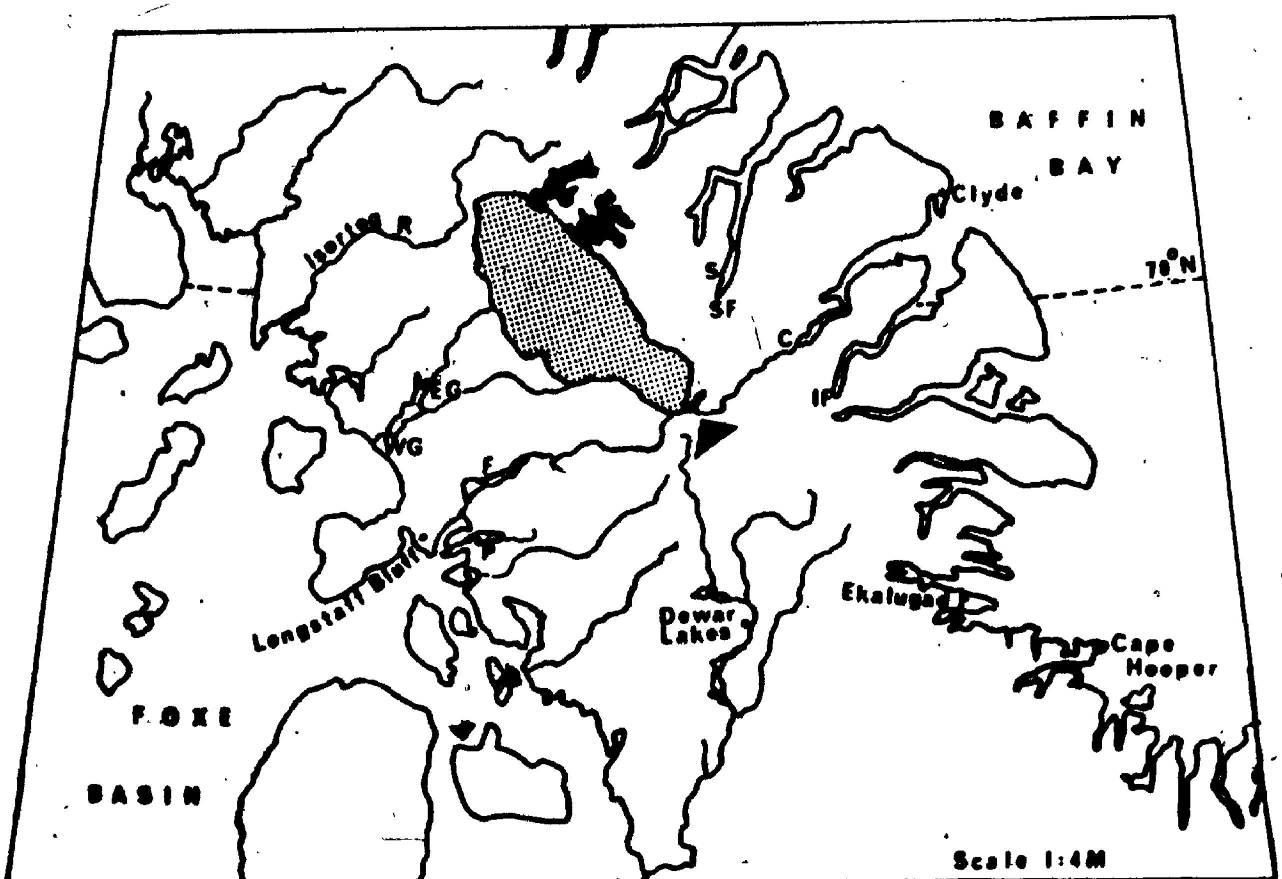


Figure 3-1: Localities of significance to the regional chronology. Localities identified by letters are explained in Table 3-1.

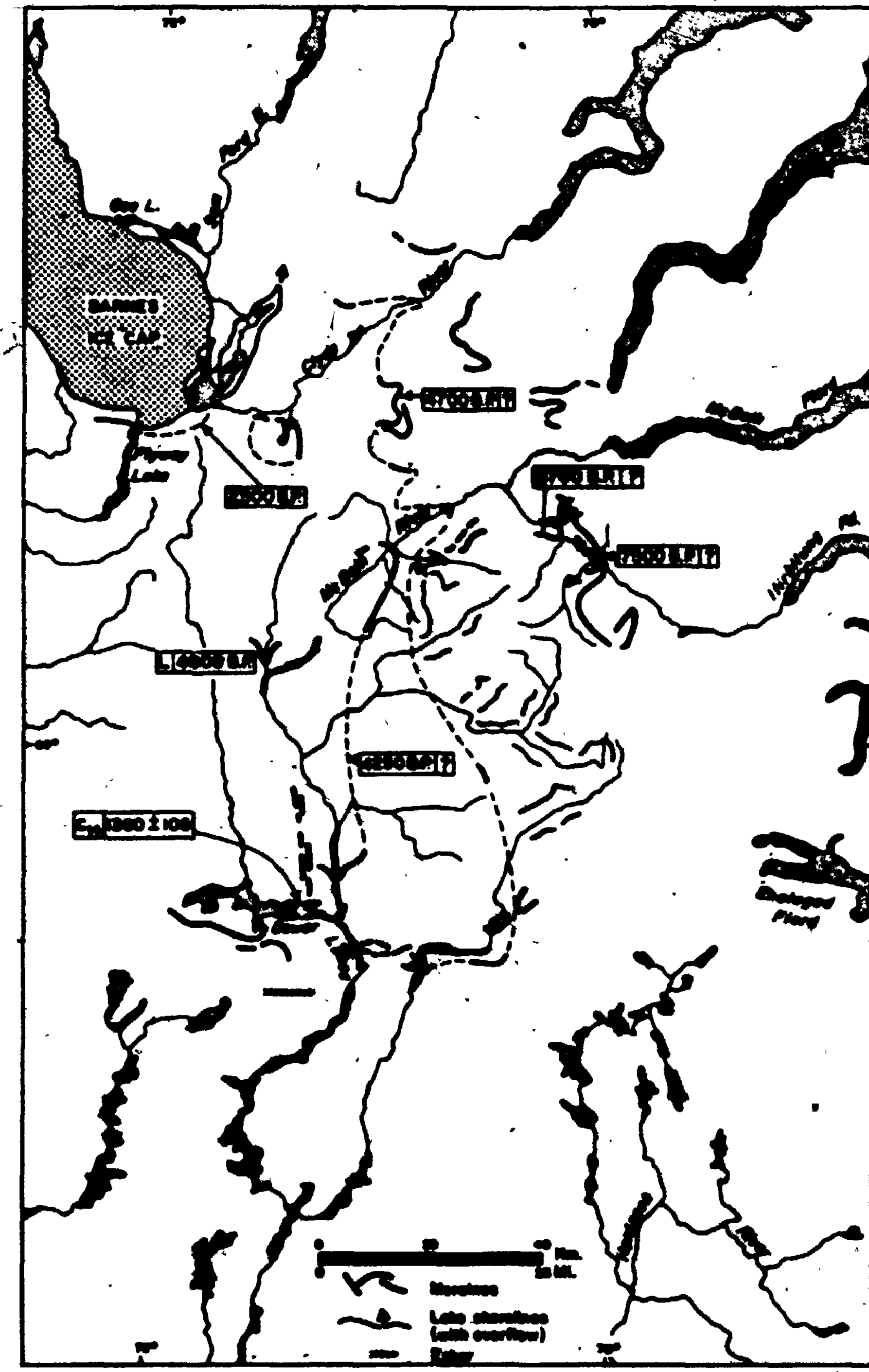


Figure 3-2: Map showing post-Cockburn moraine segments southeast of Generator Lake. Correlations are tentative and uncertainties numerous. C_{14} date is on plant detritus and does not date the moraine segment. The 'L' on the 4000 B.P. date indicates it is based on lichenometry. Note the lake shorelines adjacent to the 4700 B.P. (?) boundary.

analogue model for Quaternary ice sheets. With basal melting in the interior, he theorized that the resulting water discharge on refreezing close to the cold margin could maintain the marginal ice at or near melting point from the latent heat of freezing (Boulton, 1975, p. 39). The Laurentide Ice Sheet had a latitudinal extent of some 34° at its maximum. Therefore, there must have been a thermal continuum in the ice between the latitudinal extremes as significant spatial and temporal differences in solar radiation intensity must have led to differences in thermal conditions and thickness of the ice and hence to the ease or difficulty of meltwater penetration. One early recognition of this condition is the 'Cochrane surge' shown on Prest's map (1969) of ice marginal positions. Therefore, it is hypothesized that some proglacial processes formerly operating in "southern" ice-dammed lakes such as Lake Algonquin in central North America must have differed from those operating in present "northern" lakes such as Generator, Conn or Bieler Lakes. Lake Ojibway-Barlow would then have experienced processes transitional between the two latitudinal extremes. Careful examination of differences of the landforms (responses) in these different proglacial environments could give significant leads in evaluating both geomorphic processes and former ice conditions.

It is not the intention to directly examine process or response in lake basins other than Generator Lake but part of the utility of a process-response model lies in its applicability to other proglacial lake environments.

II The Cockburn phase

The starting point for the chronology is the sub-continental ice mass termed the Cockburn Phase of the Laurentide Ice Sheet originally proposed by Falconer, Andrews & Ives (1965) and initially tentatively fitted into the 9000 B.P. to 8000 B.P. time envelope. Andrews & Falconer (1969, p. 1272) now relate the phase to the 8400 B.P. to 8000 B.P. time period. It is increasingly clear that this phase marked a 'late-Wisconsin' maximum (Andrews & Ives, 1972; Dyke, 1974; Andrews et al., 1975) and is related to 8500 B.P. to 8000 B.P. An ice mass of rather similar dimensions was depicted by Farrand (1964) when considering what he called the deglacial hemicycle.

This time boundary of the ice mass is based on distinct morphostratigraphic evidence of multiple moraines, particularly along the north and east margins, marking the final phase of the single sub-continental ice mass immediately preceding dramatic and rapid events in the deglaciation of Canada. These were the splitting of the ice sheet into three ice masses over Keewatin, Labrador-Ungava and Foxe Basin-Baffin Island (Prest, 1969). The Foxe Basin-Baffin Island ice was the most northerly and the most persistent of the three and the present Barnes Ice Cap is thought to be a remnant of the Wisconsin Ice (Goldthwait, 1951b).

The 'type area' for the Cockburn Phase is Cockburn Land

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alled the deglacial hemicycle.
the ice mass is based on distinct
ce of multiple moraines, parti-
d east margins, marking the final
ntinental ice mass immediately
id events in the deglaciation of
itting of the ice sheet into

in northeast Baffin Island about 80 km from Generator Lake. Therefore if we accept the Barnes Ice Cap as a 'Wisconsin' relic, then landforms created in the last 8000 to 8500 years are concentrated in a 65-80 km wide zone between the fiord heads of the east coast of Baffin Island and the eastern margin of the Barnes Ice Cap which includes the Generator Lake basin.

During the last 8000 years there has been a marked difference in process intensity, and perhaps process type, between the rapidly retreating southwestern margin and the slow retreat on the northeastern margin. Retreat of approximately 3000 km from Manitoba in the southwest contrasts with only 80 km or less from such fiord heads as Royal Society Fiord in the northeast. This contrast must have led to differences in 1) volumes and intensities of meltwater flow (tempo contrast), 2) stability of ice margins, 3) duration and size change of proglacial lakes, 4) depositional characteristics of constructional landforms. These contrasts are intended to put the model area in context but not to imply simplicity of process or uniformity of rate in any one sector.

III Uncertainties and limitations of existing dates

Four sources of uncertainty are often associated with radiocarbon chronology. They are briefly reviewed at this point because, a) the Generator Lake chronology is based on this technique, b) the accuracy of the chronology is important for evaluating rates at which processes operate,

c) it is necessary to demonstrate the difficulty of dating even major events closely. Sources of uncertainty are:

i) the time lag between the event to be dated and the colonization of, or introduction to the environment of, the datable material, including the assumption that the one dates the other. Godwin (1969) has stressed the importance of avoiding misassociation of material and events particularly in relation to plant materials but it is equally applicable to other materials.

ii) Some uncertainty about the viability of ages derived from shell dates has been introduced by the recent work of Mangerud & Gullikson (1975) and Mangerud (1972). These workers found modern shells (collected early this century and in museum collections since) could give apparent ages of up to 750 B.P. first with variations showing up along the Norwegian coast and then between Ellesmere Island and northwestern Europe. Additionally the recent work of Karrow on the effects of secondary carbonates on apparent ages of shells must be borne in mind (Dreimanis, pers. comm., 1977). Apart from acknowledging these additional uncertainties with regard to shell dates little can be done at this time but accept the general chronology (Table 3-1) which appears to fit together reasonably coherently despite much of it being dependent upon shell dates.

iii) The relationship between radiocarbon years and calendar years. This latter variable of non-parallel time scales has been the subject of much painstaking research and is being solved gradually. Tables are now available for conversion from one scale to the other via bristlecone pine calibrations, e.g. Ferguson (1970).

iv) A tendency to use the central value of a date in textual material even when it has a large dating laboratory error term (perhaps related to editorial neatness and/or the desire for precision). A modification of this is used by Andrews (1970) where he uses both ca. as a prefix or $\pm$ as a suffix without indicating a range. All laboratories do not use the same criteria in reporting the error term, some using one and some two standard deviations. Only two laboratories' data are involved here and Isotopes Inc. use one standard deviation and the Geological Survey two; hence, the error terms with GSC dates are 95% confidence limits as opposed to 68% confidence limits for Isotopes Inc. dates. Comparability is achieved by doubling the Isotopes error term or halving the GSC one.

Although not related directly to Baffin Island it is instructive to consider an example of chronological uncertainty of even a major event such as the formation of the Tyrrell Sea which occupied the Hudson Bay area and which reached "its maximum extent about 8000 years ago" (Lee, 1968, p. 517). This was not possible until ice was removed from Hudson Strait allowing marine waters into Hudson Bay which implies that deglaciation from the Cockburn Phase to the Tyrrell Sea maximum was achieved in 400 years or less. However, Lee's text was written several years before publication, during which time many new dates were obtained and on page 514 he was less specific suggesting the Tyrrell Sea "reached its maximum extent 7000 or 8000 years ago".

Even more speculative chronology based on the long climatic record established for Greenland is interesting; it is based on oxygen isotope-derived temperature fluctuations from a very long ice core (Dansgaard, Johnsen, Møller & Langway, 1969, Figure 4, p. 379). For the time envelope 8000 B.P. to 9000 B.P. two warming trends are separated by a cooler interval at about 8300 B.P.

Andrews (1969) reviewed the laboratory error term of a radiocarbon date used for establishing the deglaciation chronology of Ekalugad Fiord in eastern Baffin Island and argued on geomorphological grounds for a more precise date than the central value but within the two standard deviation limit. Even where this is possible other variables remain as possible sources of error. When a major event such as the disintegration of the Cockburn Phase ice occurs very quickly, then a two or three hundred years error (whether real or statistical) could either miss dating the event or relate to the beginning or end of it. Therefore process and response on a much smaller scale should be judged within such a time framework, where precision is still limited to a few hundreds of years.

IV <u>Post-Cockburn phases</u>

As Foxe Basin is separated from Hudson Bay by only a few islands of which Southampton is the largest, it is not surprising that relatively rapid deglacierization also occurred in this basin, despite the higher latitude. Dated molluscs associated with a moraine system (the Isortoq Phase) quasi-parallel with the western shore of Baffin Island indicate land-based ice at about 6700 years B.P. (Andrews, 1966, 1968; King & Buckley, 1967). Parts of this moraine were earlier identified by Sim (1964) during terrain analysis primarily by air photo interpretation. King & Buckley date the phase at "about 7000 years" (pages 22 and 29) as the oldest date 6725±250 B.P. (I-406) was from shells collected by Sim (1964). Preference by King & Buckley for the older date allows time for colonization by the molluscs. However, the detail of the chronology is still based on this single date which has a large error term (1 S.D.) and, therefore, should be treated with due caution, particularly as Paleozoic carbonates are common in Foxe Basin.

The later chronology on the western side of Baffin Island was developed in three essentially complementary studies. The first by Andrews (1965, 1968) examined the chronological sequence of landforms in the Isortoq valley right up to the Lewis Glacier snout on the northwest margin of the Barnes Ice Cap. The second study by King & Buckley (1967) examined the area immediately south of the Isortoq valley but the work did not extend to the present ice margin. Finally, the Flint River landform sequence was examined by Andrews from the coastal Isortoq Phase equivalent up to the present ice margin (Løken & Andrews, 1966; Andrews, 1970). Eight distinct phases were identified from 6700± B.P. to 3700± B.P. (Andrews, 1966, Figure 11, p. 184). These were based on the morphology of strandlines, including ice-contact deltas, six radiocarbon dates and a predicted uplift curve, the latter based, at least in part, on the dates. Of the events outlined he believed the most important to be the Isortoq Phase (6700± B.P.) a readvance or stillstand (Andrews, 1966, 1970), the Flint Phase, a readvance (5000 B.P. to 4700 B.P.) and a well marked strandline at 3700 B.P.

Younger phases indicated by prominent moraines were dated by lichenometry (Løken & Andrews, 1966). They were the King Phase 1700 B.P. (an advance), a further advance at 700 B.P. and another at 250 B.P. (the Lewis Phase). The King and Lewis Phases are named for moraines northwest of the ice cap and discussed in Andrews & Webber (1964).

This degree of detail is necessary as responses on the west side of the Barnes Ice Cap might reasonably be expected to have equivalents on the east side except that any such evidence should be more restricted geographically. Evidence for chronology east of the ice cap is less complete although preliminary maps were prepared by the writer for presentation to the Geomorphology of the Eastern Canadian Arctic Symposium at the 1967 annual meeting of the Canadian Association of Geographers. They form the basis of Figure 3-2. Deglaciation chronology of several fiord heads is well documented: Sam Ford Fiord (Smith, 1966, p. 78, 6240±140 B.P. I-1556), Clyde Inlet (Barnett, <u>in</u> Andrews & Drapier, 1967, 7940±130 B.P. I-1932), Inugsuin Fiord (Løken, 1965; Barnett, <u>in</u> Andrews & Drapier, 1967, 7970±340 B.P. I-1673) Ekalugad Fiord and adjacent fiords (Church, 1972, p. 21, 6100 B.P.; Andrews, 1969, 5900±130 B.P.; Andrews, Buckley & England, 1970, 8160 B.P. to 7480 B.P.).

An estimate of a time envelope for undated moraines east of the ice cap (Figure 3-2) can be made by interpolation between dated events, and the assumption of a mean value for net retreat of the ice margin. Moraines indicate pauses in retreat or even readvance of the ice margin so that such interpolation is far from satisfactory by itself. However, a finite series of geomorphologically significant events were identified and by correlation with similar events elsewhere on the island or adjacent areas, a tentative chronology has been developed. This method has the distinct disadvantage of creating false correlations when, or if, poorly defined events are interpolated into an inadequately controlled chronological framework.

Net retreat of the ice margin from 8000 B.P. to the present was eight metres per year from close to the head of Clyde Inlet and almost seven metres per year from a similar position in Sam Ford Fiord (Table 3-2). Slightly over nine metres retreat per year is indicated by dividing the distance between the head of Inugsuin Fiord and the present ice margin by 8000 but this is less easily comparable to the

TABLE 3-2

APPARENT NET RETREAT RATES OF THE ICE MARGIN WEST AND EAST OF THE BARNES ICE CAP

West of Barnes Ice Cap

Localities (Fig. 3-1)	Linear Distance (km)	Dates (BP)	Rate (m/yr)
(P) Piling Lake - Barnes	115	6700-Present	17
(WG) W.Gillian Lake - Barnes	82	6700-Present	12
(EG) E.Gillian Lake - Barnes	42	4700-Present	9
(WG) W.Gillian Lake - E.Gillian	40	6700-4700	20
(F) Flint Lake - Barnes.	50	5000-Present	10

East of Barnes Ice Cap

Localities (Fig. 3-1)	Linear Distance (km)	Dates (BP)	Rate (m/yr)
(C) Clyde Inlet Head - Barnes	65	8000-Present	8
(S) Sam Ford Fiord - Barnes	60	8000-Present	7
(IF) Inugsuin Fiord Head - Barnes	75	8000-Present	9
(S) Sam Ford Fiord - Head of fiord	17.5	8000-6270	10*
(SF) Head of Sam Ford Fiord - Barnes	40	6270-Present	5
Generator L. initiation - ice cliffs	30	4000-Present	8
Generator L. initiation - draining from 76 m	24	4000-2600	17*
Generator L. present level ice cliffs	6	1300-Present	5
Generator L. adjustment phase	0.75*	2600-1300	0.6

Notes: Dates are rounded to nearest century of central value of C14 date.

Rates are rounded to nearest metre except where less than one.

Generally faster net retreat is apparent west of the Barnes Ice Cap than to the east.

* Ice margin subject to calving into water bodies.

above as a topographic barrier exists between the two localities hence ice did not flow directly from one to the other. In contrast, net retreat rates from the west side of the ice cap (Table 3-2) are almost double those to the east. It is highly unlikely that this unequal retreat is attributable to significantly different temperatures or radiation but more likely to an asymmetrical distribution of snow either as precipitation or due to subsequent drifting. This process is currently active and Løken & Sagar (1968, p. 282) suggest that northeastward migration of the crest of the ice cap is still taking place. This interpretation of migration is complicated by the identification of a surge which has changed the location of the crest line (Holdsworth, 1973b).

CHRONOLOGY OF GENERATOR LAKE BASIN

I . Sequence of events

Deglaciation of the Generator Lake basin is divided into three phases.

a) Phase I - the 76 m lake: Phase I began as the ice margin retreated westward across the main Baffin watershed. In the absence of the Barnes Ice Cap this would be the principal watershed separating east flowing streams to Baffin Bay and west flowing streams to Foxe Basin. As the ice margin retreated down the regional slope a lake was formed in the head of the valley emerging from beneath the ice. This meltwater lake which was narrow and shallow, filled to the level of the lowest col in the watershed and outflow was by a channel at the head of the valley into the Sam Ford Fiord drainage system (Figure 3-2). As deglaciation proceeded the lake became larger and deeper reflecting the broadening and deepening to the west of the preglacial valley. The overall sequence is shown schematically on Figure 3-3 (Maps 1-4).

A secondary overflow channel also flowing into the Sam Ford drainage and shown at 's' on Figure 3-3 (Map 2) may have been in operation for a very brief period. The contour map shows the channel to be marginally below the main overflow level. Systematic measurable shoreline deformation was discounted by a levelling program which showed no significant tilt over 22 km of levelling (Barnett, 1967, p. 185). However, a minor deltaic accumulation on the lake end of the channel attests to the water flow coming into the lake rather than leaving it by this route. This inflow indicates that the ice margin lay closer to the channel than at present.

The main overflow channel persisted until the lake had grown to an area of 120 km^2 resulting from a net retreat of the ice margin of 26 km, at which time a lower ice marginal route southeastward into the Clyde River became operational and marked the beginning of Phase II. A number of geomorphologically significant events took place during the 26 km of retreat. Glacial, proglacial and non-glacial drainage entered the lake which maintained its level as a response

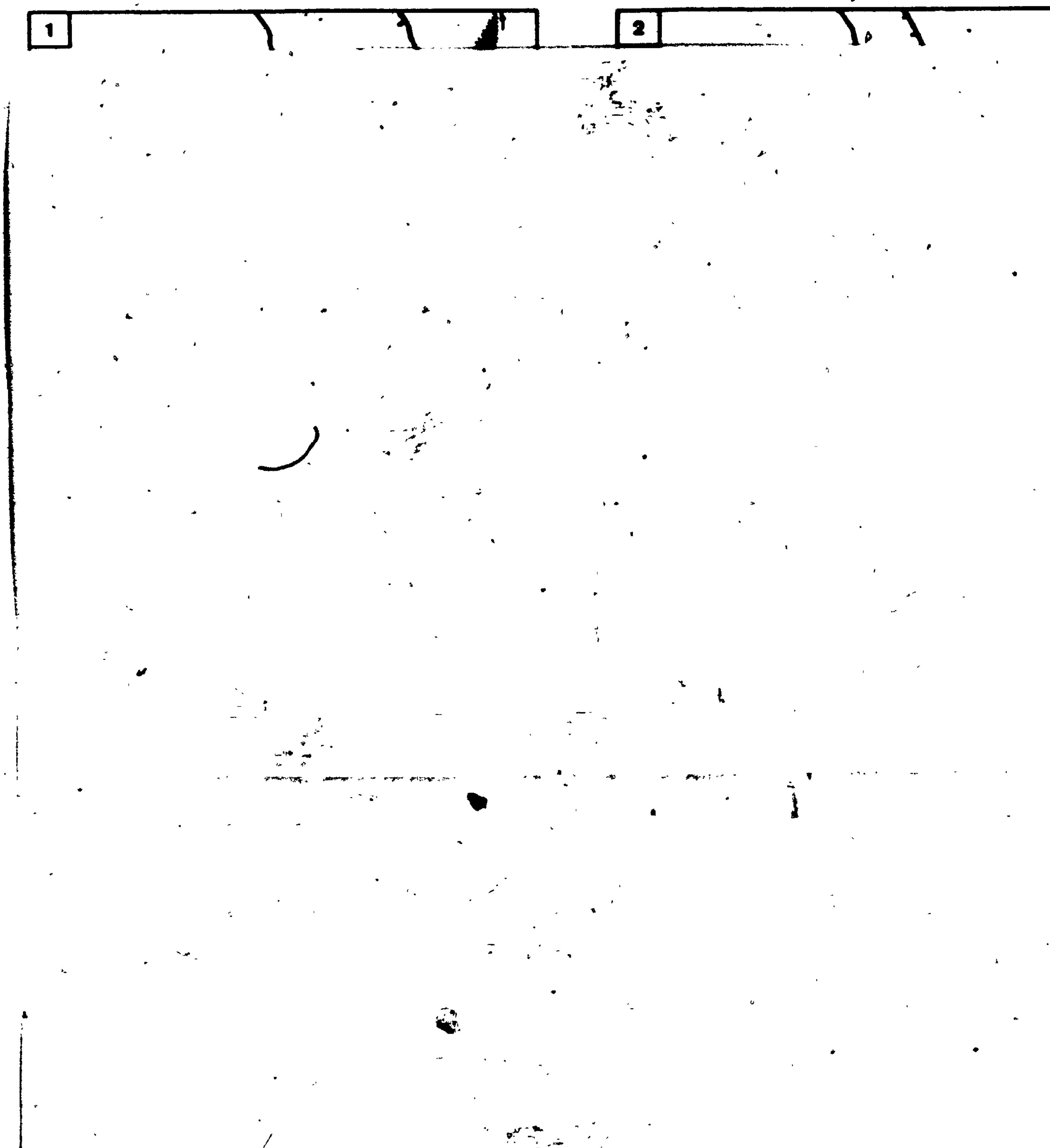
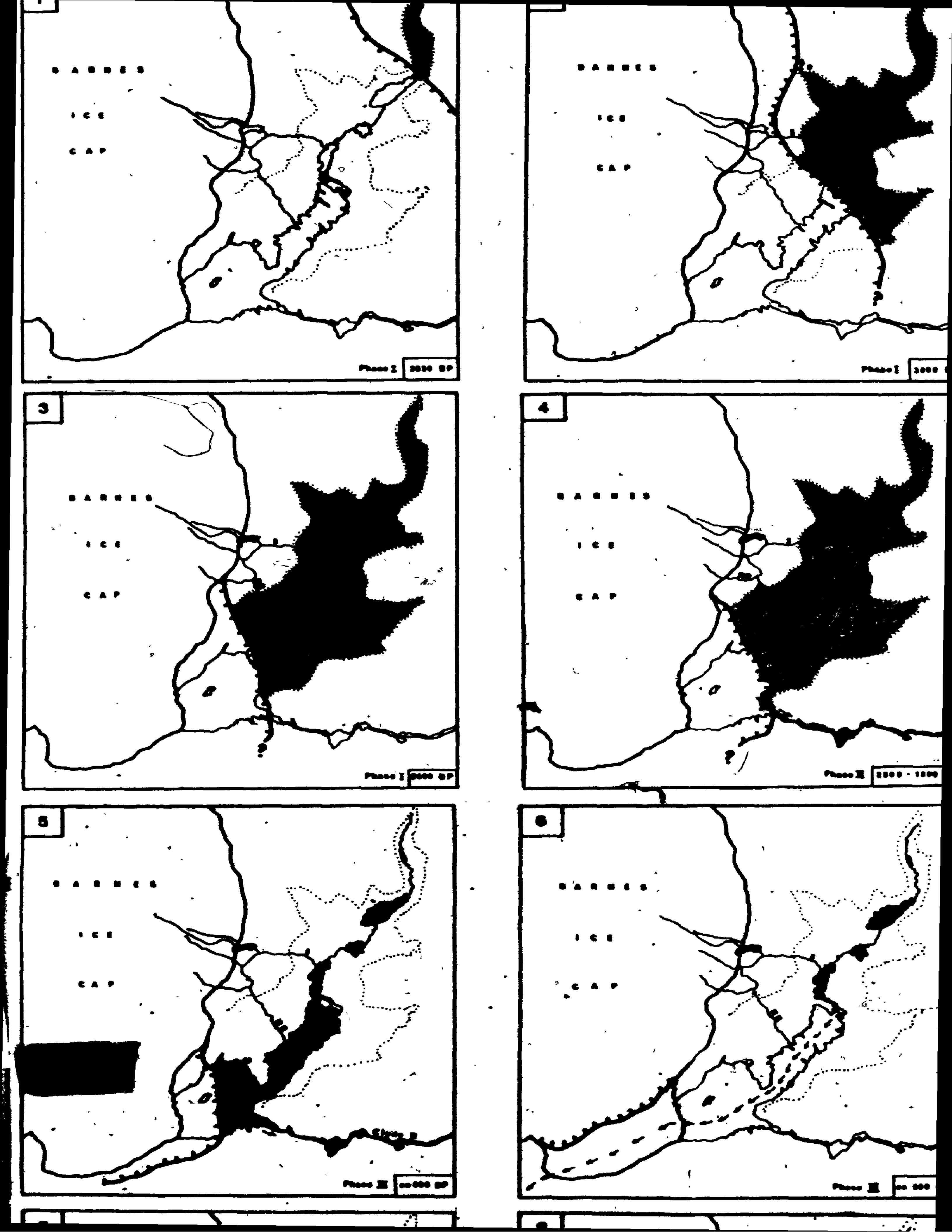
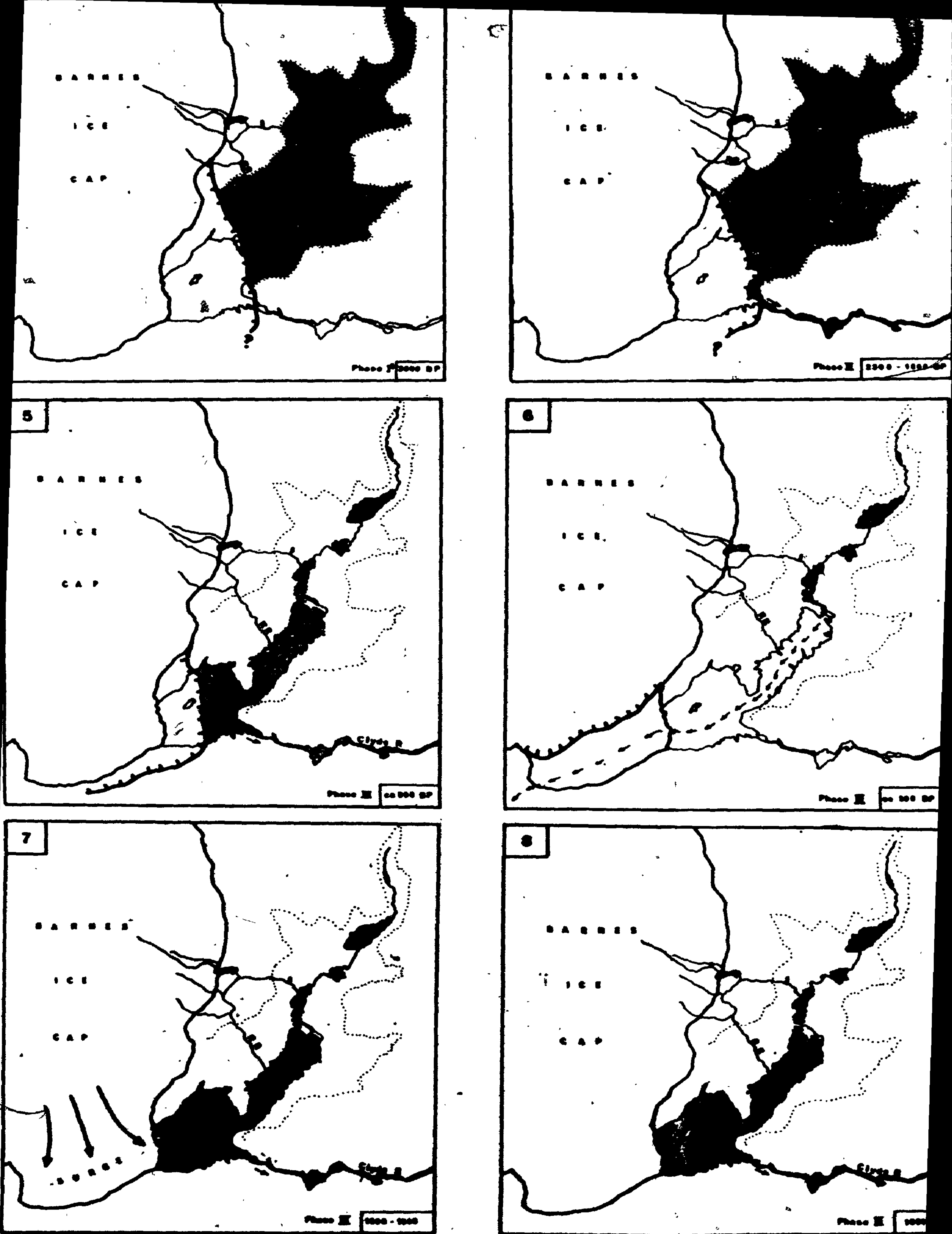


Figure 3-3: Eight maps showing three phases in the evolution of the ice margin and the resulting lake bodies at specific times identified in the chronology.



3
4
5
6
BARNES ICE CAP
Phase I
Phase II
Phase III


BARNES
ICE
CAP
Phase
BARNES
ICE
CAP
Phase
5
BARNES
ICE
CAP
Clyde R
Phase
6
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ICE
CAP
Phase
7
BARNES
ICE
CAP
Clyde R
Phase
8
BARNES
ICE
CAP
Clyde R
Phase

to basin geometry and the level of the overflow channel. Deltaic deposition occurred where streams entered the lake. The stream marginal to the ice carried a larger load than those further from the ice as it had a large available load, derived from fresh moraine, and a generally greater volume derived from the melting ice cap. Consequently, deltas on the north side of the lake were larger than those on the south side and a chronological sequence may be derived from those deltas (Figure 3-4) formed in response to successive ice marginal drainage routes. Several shallow ice marginal drainage channels (Figure 4-2) indicate detail of the phases of retreat.

b) Phase II - the adjustment phase 76 m to present level: Following the breakthrough of lake drainage to the lower Clyde River route (Figure 3-3, Map 4) a series of dramatic events occurred culminating in the establishment of the lake at a level about 76 m lower than the previous one. This adjustment phase (Barnett & Holdsworth, 1974, p. 390) (Figure 3-5) included at least four intermediate levels of lake stabilization of sufficient duration to allow formation of distinct deltas (Figure 3-4) principally by the ice marginal stream. These adjustments resulted from the initial drainage into the Clyde River leaving an unstable ice cliff as the lake level lowered rapidly. Stabilization of the cliff was achieved by an ice advance (to lower the slope of the ice) which curtailed, but did not cut off, drainage into the Clyde River. As greater

Figure 3-4: Flight of deltas along the 'Robinson' River including the present one visible below water level.

(GSC Photo 202522 Q)

Figure 3-6: Large Rhizocarpon geographicum thallus with uniform outline located on steep surface (no water ponding or bird perch).

(GSC Photo 202522 R)

stability of the cliff was achieved, down-cutting by the ice marginal stream again took place, leading to lowering of lake level and further instability of the ice cliff. Adjustment of the ice cliff profile by ice advance again took place and lake level stabilized again without completely cutting off drainage into the Clyde River.

This sequence repeated with decreasing magnitude of fluctuations until lake level stabilized at 76$\pm$1 m below the former level. Five distinct deltas (Figure 3-4) are present including the 76 m one, along the 'Robinson' River on the north side of the lake.

The supporting evidence for these events apart from the deltas themselves is a) a series of partially overridden sublacustrine moraine segments in the zone just east of the Clyde outlet which indicate localized readvance of the ice, b) a lake bottom profile (Barnett, Forbes & Whytock, 1970, Figure 2, Profile 7) (Figure 4-2) which shows a distinct general rise of the lake bottom in this zone just east of the narrows - it could be an accumulation of till or a bedrock rise, or a combination of the two, c) bedrock washed clean of debris at the intermediate levels on the down-stream side of the Clyde outlet, and a little further downstream an accumulation of outwash material well above the present river level.

c) Phase III - the present lake including the drained episode: Following the lowest lake level stabilization the ice margin retreated westward pausing briefly at a point some two kilometres west of the Clyde outlet to form a subdued lateral moraine (Figure 4-2) and then retreat continued until the lake drained via the preglacial route to Foxe Basin. The evidence for this route is the topographic continuity of the valley under the ice demonstrated by sounding (Littlewood, 1952). Some time following this establishment of uninterrupted drainage, part of the southern margin of the Barnes Ice Cap surged (Løken, 1969; Holdsworth, 1973b) and blocked the valley again leading to the re-establishment of Generator lake and rising lake level until the reactivation of the Clyde River outlet. The eastward limit of the surge is marked by a major moraine against which the ice front stood at the time of the 1948 photography. How much earlier than 1948 this surge occurred is uncertain but Holdsworth (1973a) believes that it was this century and the writer would surmise it was during the second quarter of it, as the 1948 ice margin (air photographs) is virtually contiguous with the surge moraine indicating negligible net retreat since the surge occurred.

Since 1948 the ice margin has retreated rapidly but irregularly. In part this was due to the intersection of the cliff line with a heavily crevassed zone (where retreat was most rapid) contrasting with the ramp zone where forward movement of the ice was greatest (where retreat was less rapid). Westward of the surge moraine a small number of sublacustrine moraines (Figure 4-6) occur which are less than 5 m high as determined by depth sounding. By 1969 the lake was slightly more than 60 m deep at its deepest part just south of the mid-point along the cliff line. Net retreat of the margin is continuing.

A tentative chronology based on lichenometry (Barnett, 1967) must be re-evaluated in light of radiocarbon dates (Barnett & Holdsworth, 1974), the details of which follow. The relative chronology based on lichenometry remains the same (even though the draining and surge were not recognized at that time) but the 'absolute' chronology is now thought to be considerably longer.

The next step is to document the timing of these phases as closely as possible.

II Chronology of the phases

a) Radiocarbon samples: Eleven radiocarbon dates relating to Phases I and II were determined by the Geological Survey of Canada Radiocarbon Laboratory for small organic samples collected from deltaic deposits within the Generator Lake basin. Sample locations are on Figure 4-2. These dates require careful evaluation. All dates younger than 5000 B.P. are now automatically corrected for $C^{13}C^{12}$ fractionation, a procedure which, thus far, yields corrections most commonly less than 100 years which suggests that the three dates not corrected (Table 3-3) may well be within 100 years of the corrected value. A review of other modern aspects of the technique is given in Olsson (1968). Table 3-3 summarizes the dates, corrections and other data.

Prior to assessing the dates in detail the circumstances

TABLE 3-3

RADIOCARBON DATA

GSC sample no.	^{14}C Date B.P. (yr)	$^{13}C/^{12}C$ Corrected date B.P. (yr)	Range: 95% Confidence limits (yr)	Lake level above present lake level (m)
GSC 1087	3650 ± 140	3690 ± 140	3830–3550	76
GSC 1244	3690 ± 250	3730 ± 250	3980–3480	76
GSC 1621	2240 ± 390	N/A	2630–1850	76
GSC 1276	3080 ± 170	3090 ± 170	3260–2920	76
GSC 1168	4600 ± 290	N/A	4890–4310	76
GSC 1304	2480 ± 150	2520 ± 150	2670–2370	76
GSC 1315	2600 ± 150	2620 ± 150	2770–2470	55
GSC 1622	2180 ± 240	N/A	2420–1940	23
GSC 1325	1480 ± 160	1530 ± 160	1690–1370	23
GSC 1177	1560 ± 140	1660 ± 140	1800–1520	12
GSC 1239	1240 ± 210	1270 ± 210	1480–1060	3

NOTES: 1. All samples were small fragments of plant detritus, dominantly mosses.
2. Corrections for GSC 1168, 1621, and 1622 unavailable due to laboratory problems.

of their occurrence and collection will be reviewed. All dates within the basin were based on detrital vegetation deposited in a deltaic environment currently above lake level. The lake basin is very sparsely vegetated with scattered lichens, mosses and a few vascular plants. Organic matter in the sandy foreset beds of deltas is, therefore, not abundant and in some cases only a few grams were located and in others none at all. All samples were carefully collected from sites free from surface vegetation and associated roots and apparently free from slumping. The organic material was stored in polythene bags or bottles both before and after air drying in the field. Seven of the dates are from a flight of deltas (Figure 3-4) along the 'Robinson' River which is only seven kilometres long from the ice margin to the lake. Only one sample, GSC 1037, was located from the southeast side of the lake.

b) Dates relating to Phase I: Initiation of this lake phase is interpolated to date at about 4500 B.P. to 4000 B.P. This estimate is not based on closely bracketing dates, but between radiocarbon dates from the heads of Sam Ford Fiord, Clyde Inlet and Inugsuin Fiord (Barnett, in Andrews & Drapier, 1967) and those in the lake basin. A date close to 4000 B.P. is preferred for initiation but 4500 B.P. cannot be ruled out. Four different deltas on the 76±1 m shoreline have yielded six datable samples ranging from 3690 B.P. in the northeast to 2520 B.P. on the 'Robinson' River. Sample GSC 1276 gave a date of 3090±170 B.P. and is from a delta no longer associated with the ice marginal stream which created it. A markedly underfit stream uses the same channel but it is little more than a snowmelt gully. This circumstance is of considerable importance for interpretation as the divorce of the processes from the response - the stream and lake waters from the delta - indicates that the organic material was deposited before any of the deltas to the west were formed, unless one invokes an ice marginal fluctuation for which no geomorphic evidence has been recognized. The flight of five deltas immediately west of the one in question were created by the same stream after its course changed as the ice margin withdrew. This, therefore, establishes a benchmark date. Of the three dates available northeast of GSC 1276 (3090±170 B.P.) two are older and of the seven to the southwest six are younger (Table 3-3). Such a trend is compatible with the qualitative evolution of the lake and its landforms outlined in Barnett (1967). The two dates which are not obviously compatible with this sequence are GSC 1168 (4600± 290 B.P.) and GSC 1621 (2240±390 B.P.) each of which differs substantially from one other date from a slightly different location in the respective deltas. No human or laboratory errors are suspected for either sample but both were minute (4.5 g. and 1.2 g. respectively), an undesirable characteristic for accurate dating.

Contamination of the samples with material of a different age is a possibility. Betula pollen has been identified by my colleague, R.J. Mott (GSC), from both deltaic and lake bottom samples. No exotic macrofossils have been identified despite a diligent search and no living specimens of <u>Betula</u> have been found in the area. The nearest known occurrence of fossil <u>Betula</u> pollen is in the Isortoq plant-bearing beds 200 km to the northwest (Terasmae, Webber & Andrews, 1966). The concentration of pollen grains, though small (4200 grains per ml sediment), exceed expected airborne values. It is stressed that despite this uncertainty, one of the two dates is older and the other younger than the 'preferred' chronology, suggesting that if contamination was responsible then it was not a simple matter.

The two other dates northeast of the 3090±170 B.P. site are 3690±140 B.P. (GSC 1087) and 3730±250 B.P. (GSC 1244) which are statistically indistinguishable; the site of GSC 1037 is older geomorphically and the 95% confidence limits (Table 3-3) allow this date to be between 350 and 70 years older than GSC 1244. The site of GSC 1244 is then between 1060 and 220 years older than the site of GSC 1276 (3090±170 B.P.) also using the outer confidence limits.

The remaining accepted date relating to the 76 m level shoreline is GSC 1304 (2520±150 B.P.) which is between 890 and 250 years younger than the benchmark date of 3090±170 B.P.

Within these statistical limits it is possible to argue on geomorphological grounds for further refinement of the probable duration of each portion of Phase I of the evolution within the chronological framework.

c) Dates relating to Phase II: Phase II is a critical phase in the evolution of Generator Lake; any data relating to this phase are particularly valuable aids in unravelling a complex history. The phase requires that a major lake be lowered by 76±1 m by means of an ice marginal route but with four distinct pauses in the process allowing four distinct intermediate-level deltas to form. The four levels are distinguishable by their radiocarbon dates. This chronological separation and the morphological distinction of the deltas aided as constraints in the evolution of the glaciological explanatory diagram prepared by Holdsworth (Barnett & Holdsworth, 1974, p. 392) and which is reproduced here as Figure 3-5.

Five radiocarbon dates were run on organic material collected from the four delta levels below the 76 m one. A declining trend is immediately apparent from 2620±150 B.P. (GSC 1315) at 55 m to 1270±210 B.P. (GSC 1239) at 3 m above the present lake level, with the other three dates falling between the two and relating to levels at 23 m and 12 m.

The 23 m delta yielded two samples dated at 2180±240 B.P. (GSC 1622) and 1530±160 B.P. (GSC 1325) suggesting persistence of the delta for between 650 and 250 years using maximum and minimum values from the laboratory confidence limits of both dates. The organic materials were found at opposite ends of the delta with the upstream end yielding the older material and the downstream end the

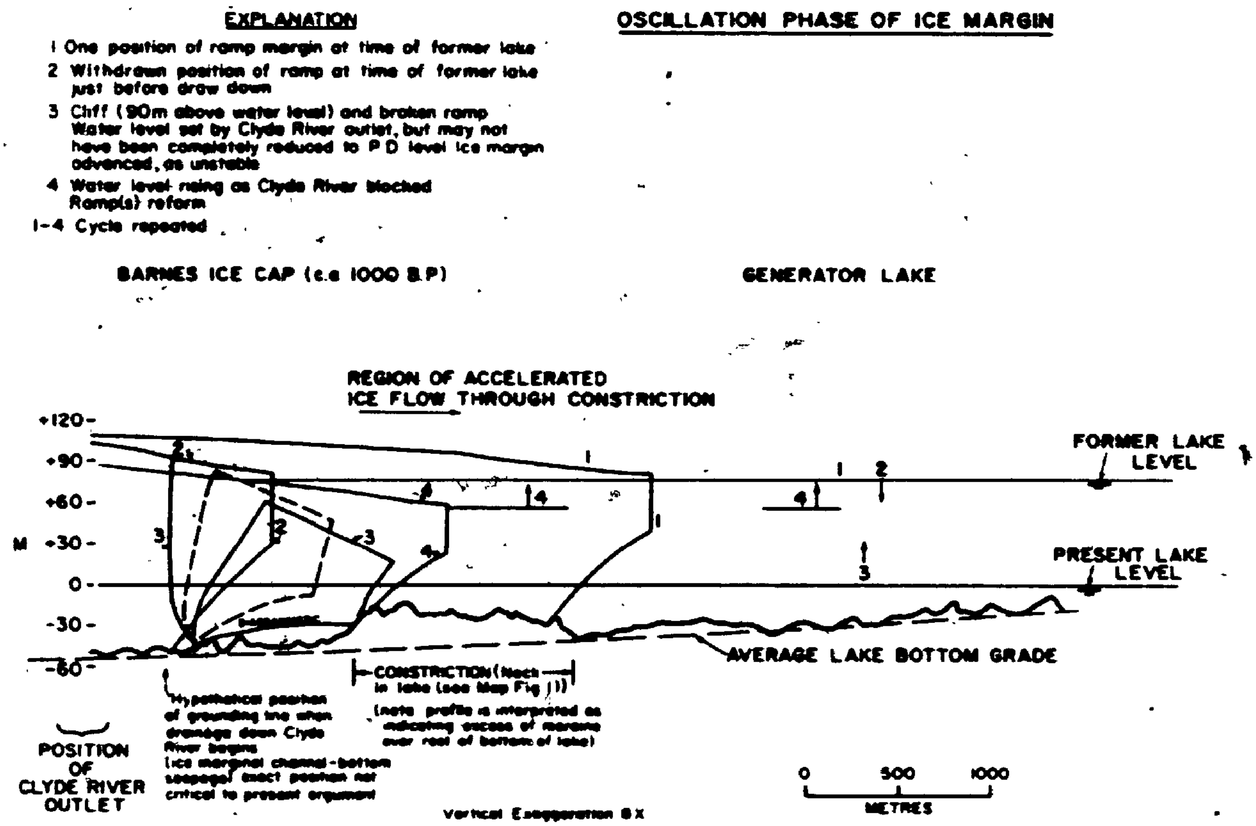

Figure 3-5: Diagram illustrating Phase II of the lake history. Ice cap profile is a reconstruction based on present profiles.

younger. The date of 1660$\pm$140 B.P. (GSC 1177) from the 12 m delta overlaps the younger date from the 23 m delta, but even in this case the outer confidence limits allow a statistical chronological separation to be made to reflect the morphological distinction.

As outlined in Figure 3-5 glaciological considerations of relative ease of fluvial erosion of the ice suggest a rapid drop in lake level is expectable when overflow drainage became ice marginal (initiating the Clyde River outlet) but that any rapid drop was immediately followed by an accelerated forward movement of the ice to stabilize the ice cliff, exposed by the sudden drop in lake level, by lowering the ice surface gradient. This physical response would be a direct result of the geometry of the unsupported ice margin. If this rapid drop did take place then the relative time spent at the lower levels was considerably shorter than that at each of the delta levels. Thus lake level probably followed a cyclic pattern of rapid drop and build-up to a stable level somewhat lower than the preceding one. The sequence of drops in net level of 21 m, 22 m, 13 m and 9 m suggest such a trend. In each case the Clyde outlet had to be partially sealed off by advancing ice.

During these glacial fluctuations the partially ice-marginal 'Robinson' River creating the delta levels was in continuous summer operation, a circumstance which has significance when assessing process rates in the next chapter.

The volume of sand deposited in the delta 12 m above lake level is larger than for any other except that at 76 m. Assuming similar stream discharge and similar available load throughout Phase II then the longest stability occurred at the 12 m level. The duration at each level during Phase II appears to have been of the order of two centuries.

d) Dates relating to Phase III: No radiocarbon dates relate directly to this phase. The GSC 1239 date (1270±210 B.P.) places an upper limit on the phase which may have been initiated as late as 1060 B.P. Between that date and the 1948 aerial photography no definite chronological control is available. Since 1948 considerable detail is available concerning changes of the ice margin through time. A period of rapid net retreat is indicated with substantial local variation of rate across the width of the ice cliffs.

Within this time frame of 1270±210 B.P. radiocarbon years at least three morphologically distinct events occurred. An unimposing moraine, both above and below the lake surface some three kilometres east of the ice cliffs (Barnett, Forbes & Whytock, 1970, Figure 4) (Figure 4-2) offers morphological evidence for a pause or minor re-advance of unknown duration in the retreat of the ice margin during Phase III. After the formation of this moraine the ice margin probably retreated sufficiently to drain Generator Lake completely (some indirect evidence from lake bottom sedimentation is presented in Chapter 4), thus re-establishing fluvial drainage to Foxe Basin. Later a surge of the south dome of the Barnes Ice Cap advanced across the valley and Generator Lake re-established itself to the level of the Clyde River outlet currently in use. Holdsworth (1973b) presented the glaciological case for the surge, initially postulated by Løken (1969). Holdsworth concluded that the surge occurred in the early twentieth century.

Radiocarbon years and calendar years (introduced in Phase III) may differ by up to some hundreds of years (Suess, 1970) but no attempt is made at reconciliation. The possible divergence of the two scales does not, in principle, invalidate interpolation between such dates but simply adds a small additional uncertainty to the derived values.

RADIOCARBON CHRONOLOGY AND LICHENOMETRY

I Basic considerations in lichenometry

As the radiocarbon chronology presented in Barnett & Holdsworth (1974) superceded the provisional lichenometry presented in Barnett (1967) without discussion of the implications for lichenometry, it is thought necessary to do so now. The relative chronology based on lichen diameters has not changed but the "absolute" chronology has lengthened considerably.

The earliest attempt at fitting a chronological framework to the geomorphology of the Generator Lake basin used lichenometry, a technique about which there are mixed opinions. There are at least three viewpoints ranging from rejection of the fundamental assumptions on which it is based (e.g. Jochimsen, 1966) through acceptance of the technique for relative dating of lichen covered surfaces (essentially the position adopted earlier, Barnett, 1967) through to using additional data to create a lichen chronology with an acceptability approaching that of radiocarbon dating (e.g. Beschel, 1961; Andrews & Webber, 1964). The variation in chronologies shown in Table 3-5 indicate the wide discrepancies between dates for a single event depending upon the basis used for establishing growth rates, and the older the event, the less reliable the date (Figure 3-7).

The Generator Lake basin data offer the first opportunity for evaluating lichenometry based on Rhizocarpon geographicum (Figure 3-6, Table 3-4) by means of a set of closely related radiocarbon dates. As the dates are from within one small area, the climatic complications are minimized, a factor which must enter into any evaluation of lichenometry. Miller & Andrews (1972) came close to this opportunity but the radiocarbon dates in that instance were much older and from a larger area with greater topographic diversity than at Generator Lake. Basic assumptions of lichenometry are: 1) the largest lichen thallus is the oldest, 2) there is a close-to-linear relationship between thallus diameter and time throughout the major portion of its life cycle after an initially more rapid growth rate, 3) lichens colonize lichen-free freshly deglaciated surfaces

TABLE 3-4

RHIZOCARPON GEOGRAPHICUM GROWTH RATES AND CHRONOLOGY: A COMPARISON

Growth Rate (mm/yr)	Basis	Lichen Diameter (mm)	Apparent Age of Surface	Lake Condition
A. Chronology derived from growth rates				
0.097	Sin (Andrews & Webber 1964, p. 101)	73	1250±150 B.P. 1280 (my calc.)	drained
0.075	Barnett (1967)	120	1470 B.P.	at 76 m
0.075	Barnett (1967)	84	1120 B.P.	at 76 m
0.053	Lower limit Løken & Andrews (1966)	110	2075 B.P.	at 76 m
0.097	Upper limit	110	1130 B.P	at 76 m
0.053	Lower limit Løken & Andrews (1966)	84	1580 B.P.	at 76 m
0.097	Upper limit	84	870 B.P.	at 76 m
0.075	Barnett (1967)	71	950 B.P.	drained
0.053	Lower limit Løken & Andrews (1966)	71	1340 B.P.	drained
0.097	Upper limit	71	730 B.P.	drained
B. Growth rates derived from radiocarbon dates				
0.027	Radiocarbon dating (GSC 1304)	71	2600 B.P.	draining from 76 m
0.028	Radiocarbon dating (GSC 1315)	71	2500 B.P.	draining from 76 m
0.035	Radiocarbon dating (GSC 1276)	110	3100 B.P.	at 76 m
0.024	Radiocarbon dating (GSC 1304)	62	2600 B.P.	draining from 76 m
0.032	Radiocarbon dating (GSC 1304)	84	2600 B.P.	inception of draining
0.033	Radiocarbon dating (Müller & Andrews, Fig. 4, 1972, p. 1136)	80	2400 B.P.*	Archeological site Cape Hooper-E.Coast

Notes: All calculated dates based on lichenometry are rounded to nearest decade.
All but first and last lichen measurements made by writer.
Range of growth rate based on C14 0.024 - 0.035 - 0.011 mm/yr mean 0.030
Range of growth rate based on A.&W. 0.053 - 0.097 - 0.044 mm/yr mean 0.075
Location of lichen stations (except last one) shown on Figure 4-2 and on Figure 1 (Barnett, 1967).
* (GaK 1978 2400±90 : 1 s.d.)

Conclusion: Radiocarbon derived growth rates are much lower and more uniform than those previously derived and, therefore, lichenometric chronologies underestimated the age of surfaces (see Table 3-5).

TABLE 3-5

PUBLISHED LICHEN CHRONOLOGIES COMPARED USING RADIOCARBON DERIVED GROWTH RATES

		Published	Revised
Lewis Glacier (N.W. Barnes Ice Cap) (Løken & Andrews 1966, p. 359)			0.030 mm/yr
18 mm	Isophyse	240±60	600±150
30 mm	Isophyse	400±90	1000±225
70 mm	Isophyse	940±220	2330±550
130 mm	Isophyse	1740±400	4330±1000
Southwest of Barnes Ice Cap Andrews & King Data (Løken & Andrews 1966, Fig. 6, p. 354-5)			
20 mm	Advance	250±	625±
30 mm	Advance ('.. an important phase')	400±90	1000±225
39 mm	Retreat (Stillstand?)	520±	1300±
53 mm	Advance	700±150	1750±375
Bruce Mountains, E. Baffin Island (Harrison 1964, p. 68)			
110 mm	Minimum age	2100±	3660±

Notes: Lichen diameters for Rhizocarpon geographicum s.l.
Error terms as published; conversion increased proportionately.
All dates are B.P.

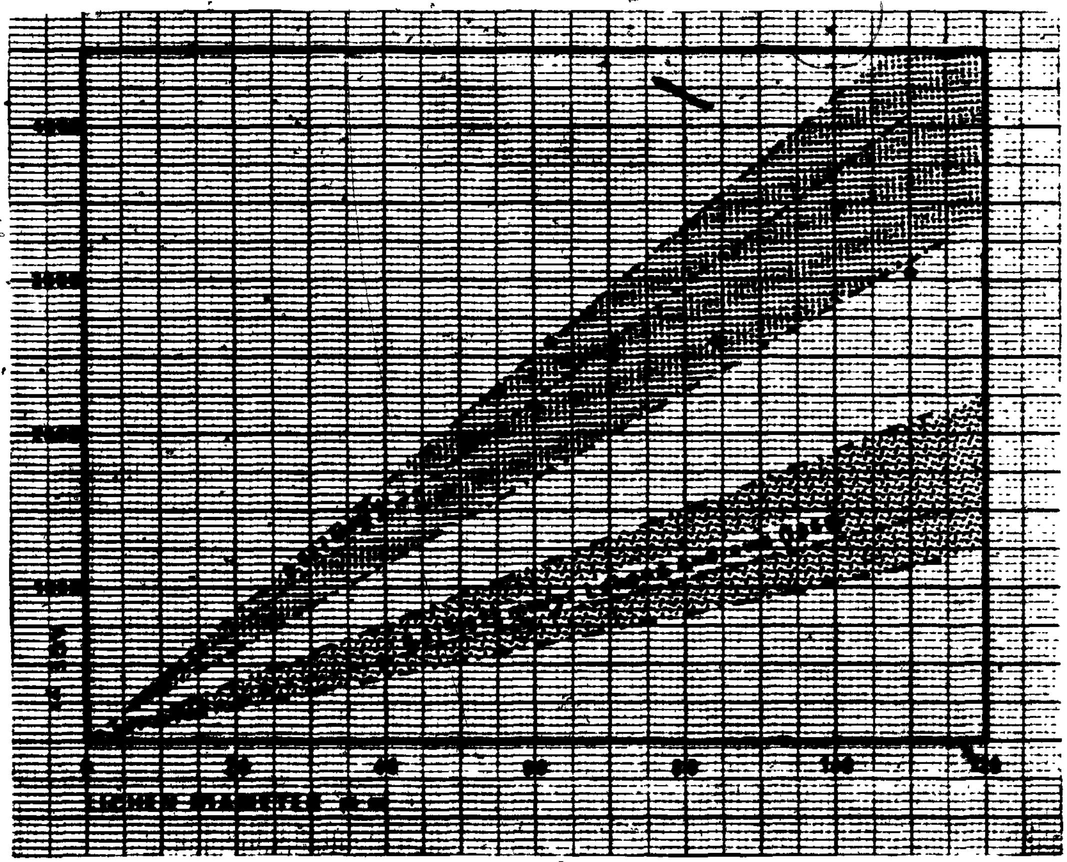

Figure 3-7: Graph showing radiocarbon derived growth rate for Rhizocarpon geographicum (0.027 mm/yr) and comparative growth rates published by Løken & Andrews (1966) for the southern margin of the Barnes Ice Cap. Points plotted are those shown on Table 3-4.

quickly, perhaps within a decade, 4) sampling will locate the largest thallus, 5) unduly favourable growing sites such as bird perches will be recognized and avoided, 6) only one species or sub-species is measured, thus assuming accurate identification.

In all cases the lichens must be related to some known chronology to derive a rate or rates of growth. In the case of remote environments this is generally difficult as habitation and related artifacts (cf. the gravestones used by Beschel in the Alps) are rare and radiocarbon dated materials are often over 1000 years old, whereas lichenometry probably has its optimal chronological value in the order of hundreds of years.

From the diversity of evidence available for Rhizocarpon geographicum sensu lato it seems probable that it is the best species for dating as it is distinctive in appearance, uniform in shape, has a long life span, and a wide geographic distribution. Equally, it seems highly improbable that its normal growth pattern is such that a simple linear relationship holds between thallus diameter and time from year one to perhaps several thousand years. Hence the question changes to: is there evidence that the relationship is sufficiently close to linear to be usable and if so, is there a possibility, as Miller & Andrews (1972, p. 1133) speculate, that a single universal growth rate prevails?

These large questions cannot be answered from data within the Generator Lake basin but some insights are available. It seems that the technique of lichenometry is valid as it has been used in several areas with at least moderate success although usually with some reservations. That a universal linear growth rate occurs seems highly improbable due to the disparity of environments of occurrence, ranging from alpine to arctic with very different insolation and thermal regimes. Ease of nutrient availability must be also a variable of some significance. The number of differing growth rates already proposed for Rhizocarpon on Baffin Island (Figure 3-7) would appear to detract from the idea of uniformity (Andrews & Webber, 1964; Løken & Andrews, 1966; Miller & Andrews, 1972; Harrison, 1964).

II Generator Lake lichen data

Twelve lichen stations were established in six pairs. In each case one was above and one below the 76±1 m shoreline. Data gathered in 1961 by Sim showed that a visible difference in lichen diameter occurred above and below the shoreline. This was readily confirmed with the differences ranging from 55 mm on the southside to 10 mm on the northside of the lake basin (Barnett, 1967, Figure 1).

Taking the largest thallus measured at each site and using the radiocarbon chronology outlined in Barnett & Holdsworth (1974) a range of apparent growth rates are produced from 0.018 mm/yr to 0.042 mm/yr with a mean of 0.027 mm/yr (Figure 3-7) (12 values). Only two values exceed 0.029 mm/yr and only one is below 0.020 mm/yr. These deviations from the mean, particularly the highest values (0.042 mm/yr and 0.040 mm/yr) reaffirm a concern expressed in an earlier discussion (Barnett, 1967, p. 182-185). Although the technique depends fundamentally upon the largest diameter thallus, in the case of Generator Lake, the largest thalli at three sites, despite being good to excellent specimens, were considered of doubtful viability for dating. This doubt related to the considerable gap between the diameter of the largest and second largest thalli measured at each of the three sites. Translated into years using the mean growth rate of 0.027 mm/yr the difference between the age of the oldest (largest) thalli and the second oldest amounts to 700, 850 and 1100 years respectively. These values should not be confused with error terms (as such they would be equivalent to those associated with radiocarbon dates) but they represent chronological discontinuities of considerable magnitude, when, or if, lichenometry is viewed uncritically. Such a discontinuity occurs at three of the six paired lichen stations, but not at the other three where close grouping of the ten largest thalli occurs. Interpretation, therefore, becomes a dilemma of whether to accept or reject the largest thalli.

Of the three large, possibly anomalous, thalli diameters, one, the easternmost at station IV, yields a radiocarbon-derived growth rate of 0.029 mm/yr which is close to the mean of 0.027 mm/yr. This isolates the other two large thalli even more, with apparent growth rates of 0.040 mm/yr and 0.042 mm/yr.

Excluding the two largest derived growth rates the mean becomes 0.025 mm/yr (10 stations) which is remarkably close to the long term value of 0.03 mm/yr proposed by Miller & Andrews (1972). Therefore, it seems reasonable to conclude that a growth rate of the order of 0.02 mm/yr to 0.03 mm/yr may be applicable for the lake basin but it is premature to go beyond that conclusion for the lichenometric value of Rhizocarpon geographicum sensu lato.

III Implications of the data

Of more importance to the question of process and response is the supportive nature of the evidence. The thalli diameters support the general evolution of the lake basin as originally outlined in Barnett (1967). Secondly, they support the chronological frame work (Barnett & Holdsworth, 1974) as the data indicate a reasonably uniform growth rate is derived from the radiocarbon chronology. This mutual support is important as in sum the evidence is greater than taken individually.

SUMMARY

The regional chronology was presented in sufficient detail to demonstrate the evidence for a 4500 year time span to be covered in the relatively small area on the eastside of the Barnes Ice Cap (contrasting with the much more rapid deglaciation of the westside). This introduction, therefore, set the outer limit on time involved and more

particularly the tempo of events liable to occur within the lake basin. A more sedate pace is proposed than that to the west of the ice cap.

The presentation of the internal chronology of the lake basin followed two lines - the former, radiocarbon chronology, the latter lichenometry and the examination of their relationship and is summarized as follows: Greater reliance is placed on the radiocarbon dating than lichenometry but despite imperfections of both sets of data a reasonable chronological framework is proposed within which the process-response model must be viewed next.

The 76 m lake developed between 4000 and 2600 B.P. as a gradual net retreat of the ice margin took place. The adjustment phase of lake level lowering took place between 2600 and 1270 B.P. or slightly younger. The lake drained completely at approximately 500 B.P. and became re-established between A.D. 1925-1945 following a surge of the Barnes Ice Cap.

CHAPTER 4

PROCESS AND RESPONSE: TOWARDS A QUANTITATIVE MODEL

Introduction

This chapter has two primary purposes, firstly to introduce and explain the structure of the qualitative process-response model and secondly to present the quantitative results of the field and laboratory investigations of selected sub-systems of the proglacial system. In attempting to present the data in sub-systems it becomes apparent just how interdependent some of the processes really are, as each environment - glacial, sub-aerial and lacustrine, interfaces with each other and each interface has been, and still is, subject to change in location through time. Consequently no single order of treatment of environments can be developed without leaving an apparent discontinuity. An attribute of the qualitative model is its incorporation of the sub-systems into one unified diagram (Figure 4-1).

THE QUALITATIVE MODEL: Its structural components:

I General format and conceptual constraints

The geomorphic unity of the drainage basin is symbolized by the circular diagram encompassing the whole process-response system. The outer circle symbolizes the watershed of the drainage basin and is divided into three segments of equal size which symbolize the three components of the

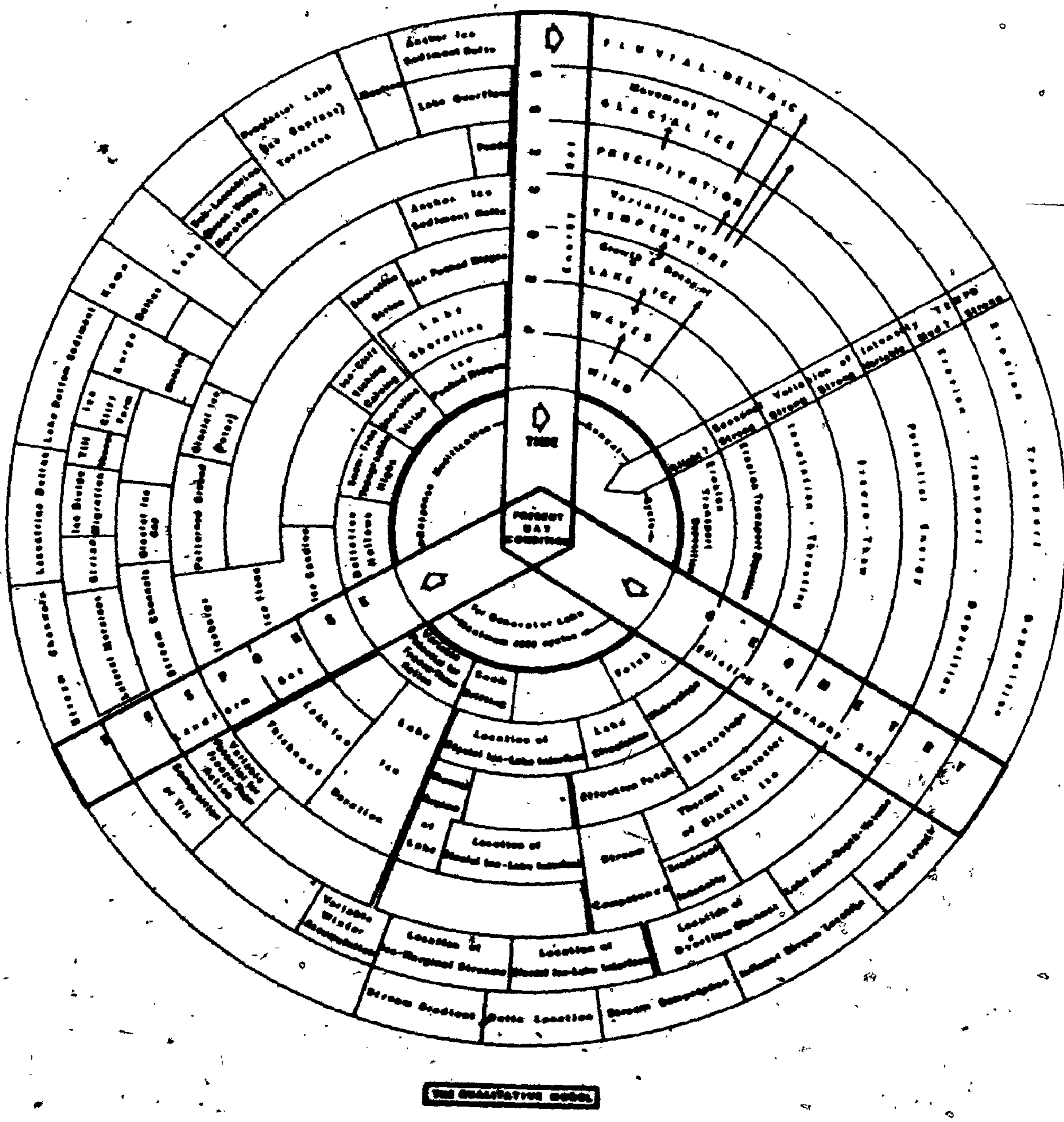

Figure 4-1: Diagrammatic representation of the Qualitative Model.

system. Beginning with exogenic processes (the first component), they then collectively interact with the basin geometry (the second component) thus initially creating a landscape (the third component) which is then subjected to the processes again which modify the surface further by erosion or deposition.

Radial linkages are indicated not only in the process sector but also in the geometry and response sectors as no process acts in isolation. For example, wind not only has direct effect through deflation and deposition but has considerable influence on the growth and decay of lake ice and on wave generation. Therefore the creation of ice pushed ridges is at least dependent upon wind, temperature, waves and the decay of lake ice. These in turn must act upon a moderate slope which has the necessary loose construction materials available for generation of the response. Similar circumstances with a lesser availability of loose materials could result in shoreline striae. Thus the essential aspect of the threefold division is the connecting link in the logic between process and response because a given process or combination of processes of given tempo do not lead to specifically predictable landforms without regard for the morphology and condition of the surface acted upon. This fundamental aspect of morphology appears to have gained little explicit attention in the literature and yet is essential in analysis of paleoenvironments. For example, interpretation of the disharmony of fluvially cut channels either crossing watersheds or running obliquely across valley sides, or marine or lacustrine raised beaches exhibiting tilt or flexure, require assessment of the morphology of paleosurface acted upon.

Endogenic processes are not considered in this model. The energy set (processes) is qualified by reference to seasonal intensity (tempo).

The geometrical role of the broad structural facets of the pre-existing landscape is one which appears to have been under-represented in previous process-response studies. It is a prime determinant of the locus of erosion and deposition as well as of the basal morphology of the depositional landforms. In the short term it can be considered reasonably static despite being subject to morphological modification by the processes operating. At this stage the model is such that any drainage basin would fit such a scheme but additional constraints are outlined next.

Of the processes listed, the presence of glacial ice distinguishes the basin from the majority of others. This distinction is further qualified by the ice occurring in a polar environment. Other constraints built into the model are the recognition of the growth and decay of lake ice as a significant process and the tempo or marked seasonal variation of intensity of most of the processes. Tempo is not often considered in geomorphology but has received thoughtful reflection by Gage (1970) and Wolman & Miller (1960). In the absence of other evidence it can only be assumed that processes were operating at a reasonably uniform rate or intensity (basically summer season to summer season) between dated events. The potential pitfalls of such an assumption were illustrated by Gage (1970, Figure 1). Fluctuations about such a mean rate are to be expected. These latter two constraints are by no means exclusive to the proglacial environment but in this case probably exhibit characteristics close to the extreme.

II <u>Detailed format</u>

a) Process: The Energy Set. Seven process types were identified: fluvial-deltaic, glacial, precipitation, thermal (= periglacial), thermal-lacustrine (= littoral), lacustrine and aeolian. The most convenient way of examining these process types is to combine them in groups where multiple interactions have lead to significant responses. Five groups of processes are considered: hydrological (fluvial, general lacustrine and precipitation); shoreline (fluvial-deltaic, thermal, aeolian and glacial); sublacustrine (lacustrine, thermal, glacial); glacial and sub-aerial (thermal, precipitation). Of the five the glacial and sub-aerial groups will receive minimal attention as in the former case other scientists have tackled the topics and the latter case the responses are minor influences on the landscape.

b) Geometry: Existing Topography Set. This component is fundamental. The location of the glacial ice margin across the valley and the fact that ice movement is up the regional slope is the very basis for the existence of the proglacial lake basin. The size, shape and depth of the lake at each of its evolutionary phases was dependent upon the existing topography and the position of the ice margin. This interaction of topography, ice and lake was in turn responsible for the locus of many of the distinctive landforms and their basal morphology.

c) Responses: Landform Set The landforms identified in the response sector are those specifically identified in the Generator Lake basin. The response sector also constitutes an inventory of potential landforms in any polar proglacial lake basin. Given this latter assumption two questions are then pertinent: are any landforms peculiar to this type of lake basin? Secondly, which landforms are known from other environments, perhaps as minor features, but are unusually well developed in this particular morphogenetic environment? Answers to these two questions will yield information on whether distinctive landforms are associated with a polar proglacial lake. The first answer is that multiple sublacustrine moraines are postulated to be peculiar in process, although not in form, to this type of environment and the second answer is that ice-pushed ridges are well developed but not unique. Anchor ice sediment rafts are also interesting ephemeral features which may be known from other environments but are not extensively reported, perhaps because of their ephemeral nature.

Therefore, few if any landforms or responses are considered unique to the type of environment particularly as multiple subaqueous moraine ridges are known from other types of environment (Hoppe, 1957; Mawdsley, 1936). The distinction of the environment is, therefore, in the degree of development of unusual forms and in the association of these well developed unusual forms.

d) Elements: The smallest boxes representing the landforms are arranged in the 'Response' component of the diagram. Their location on the diagram is determined by the primary processes operating so that where more than one primary process is identified the element will have a greater radial extent. Elements on the same process arc of the diagram represent different responses depending upon the landscape geometry on which the process operates. As the model is qualitative the relative size of each element is not significant, nor is there particular significance to neighbouring elements. Additionally some elements are left blank - this indicates that other elements could be added both for a general model or also in the case of Generator Lake. For example, fluvial landforms such as meanders are not included as they are non-distinctive in this morphogenetic context.

The elements in the 'Geometry' component identify conditions and locations. For example, stream length, gradient and competence are dependent upon the shape of the surface over which the water flows but the competence is also dependent upon precipitation and temperature. Additionally, competence is also dependent upon tempo, as heavy rain or temperatures rising rapidly above $0^{o}C$ increase competence much more quickly than gentle rain or slowly rising temperatures.

Similarly, the composition of till (without regard to its morphology) is a function of the tempo of movement of glacial ice and the nature (gradient, roughness and composition) of the surface crossed.

III Time

Time may be incorporated in the model diagram by considering annual revolutions of a pointer attached to the centre of the circle so that the greater the number of revolutions of the pointer, the more well defined the responses are likely to be. Essentially the most significant parts of three annual revolutions were observed in an attempt to examine the last 4500. This concept of time is infinitely flexible, as, if one considers very rapid processes, then the revolution of the pointer could be envisaged as daily, hourly, each minute or even each second. In this way, each morphological change has altered the geometry of the surface on which the process is acting, thus a flexible qualitative feedback loop is introduced.

Although the quantitative aspects of time are largely dependent upon the accuracy with which this set was outlined in Chapter 3, the distinctiveness of this part of the model is the considerable range of time available within the small field area. Thereby it offers the opportunity to identify benchmark dates, bracketing distinctive geomorphic events, which enable process rates to be calculated for different phases of the evolution of the basin. This in turn offers the opportunity for comparing these zones with an average rate for the whole basin. In this way not only are the process rates assessed but the chronology further tested for consistency. Any apparently anomalous rates or dates must then be re-evaluated accordingly.

IV Materials

The additional variable, materials, is considered as a dependent variable. It will be recalled from Chapter 2 that the whole basin is underlain by granite-gneiss veneered by a discontinuous coarse till. Glacial ice and lake water are materials given special attention. It is the independent variables affecting the distribution of the materials which are the focus of the dissertation.

V The quantitative goal

The ultimate goal is to identify individual quantitative relationships between processes and responses thus allowing prediction of geomorphic change through time on the basis of formulae. This goal is not completely achievable. Practical constraints include the slowness of some geological processes, limited field time and funds. Some relationships may not even be possible to identify with present field techniques. However, wherever possible, progress has been made toward the quantitative goal. The results will indicate both the degree of success and the remaining great distance to go.

THE QUANTITATIVE MODEL: Sub-systems:

I Responses

General: The map of Generator Lake basin (Figure 4-2, in pocket) is based on the 1:50 000 topographic map (NTS 27C/12) which outlines the surface geometry with 25 ft. contour intervals (the adjacent map of the Barnes Ice Cap has a contour interval of 50 ft.). The sublacustrine contours (10 metre interval) represent the writer's interpretation of echo sounding and line sounding data outlined in Barnett, Forbes & Whytock (1970). The spatial relationships of the responses discussed are displayed and sizes of the larger features shown to scale. At 1:50 000 the smaller landforms are exaggerated of necessity, but several are shown as photographs, in diagrams, or as cross-sections to follow. Although glacial processes are one sub-system where a limited contribution of new data is made it is logical to begin with it as it is the fundamental one affecting many of the others.

II Glacial processes

a) Ice margin above Generator Lake:

i) General: As stated on several occasions earlier the presence of the Barnes Ice Cap is the raison d'être of the lake and as such some consideration of the processes operating at the ice-lake interface is appropriate. This section will deal with the ice and water only, with deferment of consideration of sublacustrine moraines until the following section. Considered here are the following processes: cliff retreat including undercutting, ramp formation and decay, crevassing, moulin formation and melt-water flow, striae and glacial surging. Unless otherwise identified the data were gathered by the writer or at his request by his assistants.

ii) Cliff retreat: The 1948 air photographs show a reasonably smooth gently arcuate convex margin for the Barnes Ice Cap at Generator Lake with 17.5 m high cliffs measured photogrammetrically. By 1950 the south end of the ice margin was cliffed more vigorously as shown on ground photographs taken by the Arctic Institute expedition (reproduced in Barnett, 1967, as Figure 4). One other ground photograph (Rand Corp. 1963, Figure 37) bears a caption '100 foot ice cliffs' but it is thought to be either a visual over-estimate or to include estimation of the sublacustrine portion of the ice. Between June 30, 1960 and July 19, 1961 the north section of ice cliffs experienced a net retreat of approximately 90 m as shown on successive air photographs, but net retreat at the south end was much more modest in amount. By 1965 the highest point above lake level on the ice cliffs was measured at 17 m and this point occurred on the north-central section of the ice margin.

Retreat of the ice cliffs is continuing and several cliffing events were noted, particularly one on July 4, 1968 (Figure 4-3) when a substantial volume of the north-central section of cliff suddenly collapsed and sank into the lake (initially) in an essentially vertical path as opposed to toppling forward into the lake ice. This indicated a very substantial undercutting of the cliff below lake level and by the continuous process of thermal erosion estimated by Holdsworth to be 3-4 m per year (Barnett & Holdsworth, 1974, p. 401). Such a process is demonstrable as during the decline in lake level in August a smooth undercut became visible and is shown in Figure 4-4.

Cliffing of the ice margin is essentially a shallow water response. Two zones occur, one at the north end and one at the south end of the ice margin where water depths are generally less than 30 m (Barnett, Forbes & Whytock, 1970). The processes of cliff decay other than by ablation are by thermal erosion, mechanical breakdown (particularly in crevassed zones) and for a short period at least, by wave action with the water often armed with ice floes and bergy bits. For the period since 1948 the tempo of the retreat of the ice margin by cliffing has been strongly influenced by a crevasse zone which has remained stationary relative to bedrock. Therefore, the zone must be a response to the geometry of the subglacial surface. As the cliff margin had by 1969 intersected the crevasse zone (Figure 2-5), cliff retreat was speeded up by the occurrence of ice already broken up into blocks, or at least subject to tensional stresses as the ice flowed forward over the

Figure 4-3: Collapse of cliffed ice margin. Dark section indicates length of collapse July 4, 1968. Height above lake 15-16 m. Note incision of supraglacial stream.

Figure 4-4: Undercut sloping surface of ramped portion of the ice margin, August, 1969. Note the ice is generally debris free. Height above lake 2-3 m.

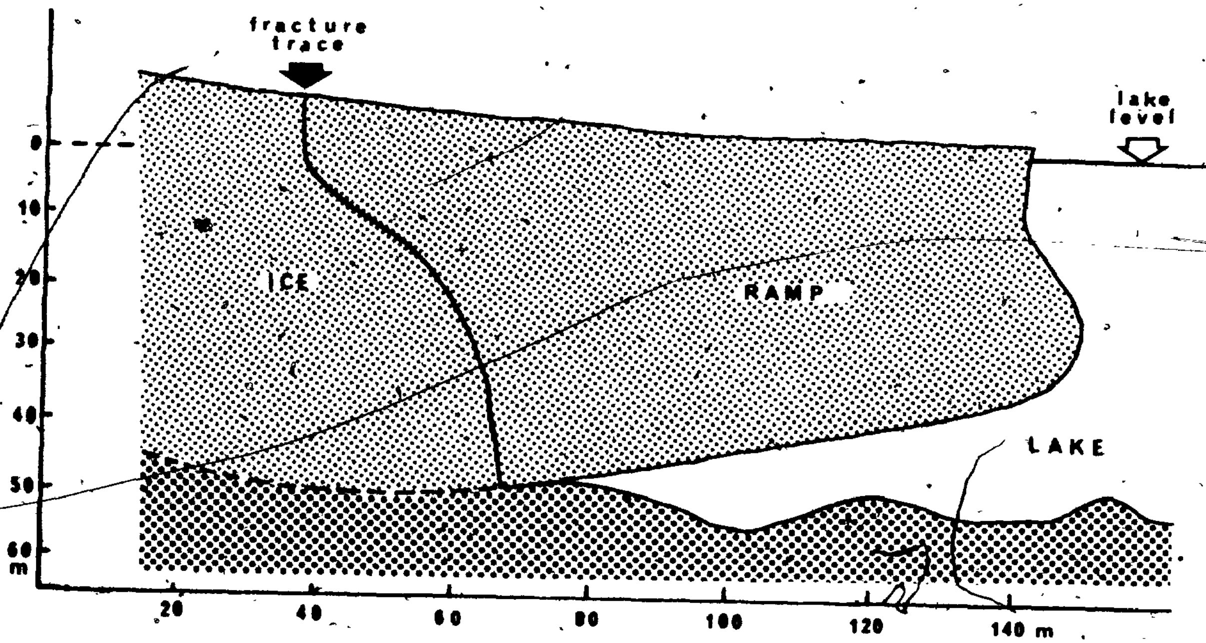

Figure 4-5: Schematic cross-section of ramped ice margin showing fracture trace. (After Barnett & Holdsworth, 1974).

subglacial irregularity.

iii) Ice ramp formation and decay: The second, and in many ways, more significant form of the ice margin is the ramp, a lobate form which tapers in cross-section towards the lake and is not in contact with the lake bottom at its leading edge (Figure 4-5). A ramp forms in water over 30-40 m deep and is subjected to buoyant stress which increases particularly as, and when, lake level rises rapidly. When discussing sublacustrine moraines next it will be noted that the space beneath the ramp is the site of the development of the moraines. The size and length of a moraine is then a function of the persistence and stability of the ramp.

The morphology of the ramp form was established by Holdsworth (Barnett & Holdsworth, 1974, p. 399) who witnessed the calving of one with a volume estimated at $0.4\text{-}0.5 \times 10^6 m^3$. From field measurement and observation of a backward tilt toward the ice cap on fracture, he was able to model the shape of the fracture plane graphically and thereby demonstrated, not only the presence of a substantial undercut of the ramp, but also that it had an underwater projection or ram, not only on the calved portion of the ramp but also on the fresh face initiating the next ramp (Barnett & Holdsworth, 1974, Figure 18, p. 400). The presence of the ram is taken to explain the diffuse echo sounder returns that had puzzled both the writer and Holdsworth when independently each had tried echo sounding very close to the ice margin.

Where well developed, the ramp form has a very low 'freeboard' at the leading edge (Figure 4-4) and one was observed partially submerged on August 9, 1968 and again on August 14, 1969. This geometry minimizes the effects of some of the following destructive processes. Ablation is reduced as a lesser area is exposed and mechanical breakdown is also greatly reduced with the absence of an unsupported wall of ice. Any wave action tends to dissipate energy as swash onto the ramp rather than concentrate its energy against the low cliff face. In addition, the rate of forward movement of the ice is greater in the ramp form due to the diminished friction relative to that prevailing at the cliffed sections. Holdsworth measured ice flow rates of 2.76 cm/day at the cliff section and as high as 7.57 cm/day on the 'TI' ramp (Barnett & Holdsworth, 1974, Figure 18, p. 400). In this way a ramp tends to develop and persist as a relatively stable form given adequate water depth.

As a ramp establishes itself as a stable form which tends to spread out into the lake, it tends to lengthen. The longer the taper the more susceptible to fracture from the buoyant forces acting each spring as the lake level rises. As already noted the particularly rapid rise in lake level, up to and including July 17, 1970, exerted sufficient stress to fracture the 'TI' ramp. Thus the process of a gradually lengthening ramp establishing itself and thinning over a period of several years, followed by a rapid rise in water level one spring leading to fracture, enables the creation of a highly plausible environment for the development of sublacustrine moraines and a cyclic mechanism allowing for multiple moraine development.

iv) Variations and predictions: The recent developments at Generator Lake differ slightly from the idealized development outlined above. These differences relate to the glacial surge, (the temporal eccentricity of which is unknown but is thought to be somewhat unusual), and to the crevasse zone. The responses to these conditions are that several contiguous small ramps rather than one big one has been the norm since 1960 and the proximity of the crevasse zone to the ice margin suggests greater instability than normal as the retreating margin passes the cause of the crevassing.

Predicted developments according to the model when the ice margin is west of the crevasse zone, and assuming a continuing negative mass balance for the Barnes Ice Cap, are as follows: the ice margin will have a low cliffed margin at the south end somewhat higher cliffs at the north end and a single spreading ramp slightly south of the mid point, reflecting the sublacustrine geometry. Beneath the ramp a sublacustrine moraine will form which will be somewhat longer and larger than the small ones (3 m) detected just in front of the 1969 ice margin (Barnett & Holdsworth, 1974, Figure 18, Profile # 4 inset, p. 400) (Figure 4-6).

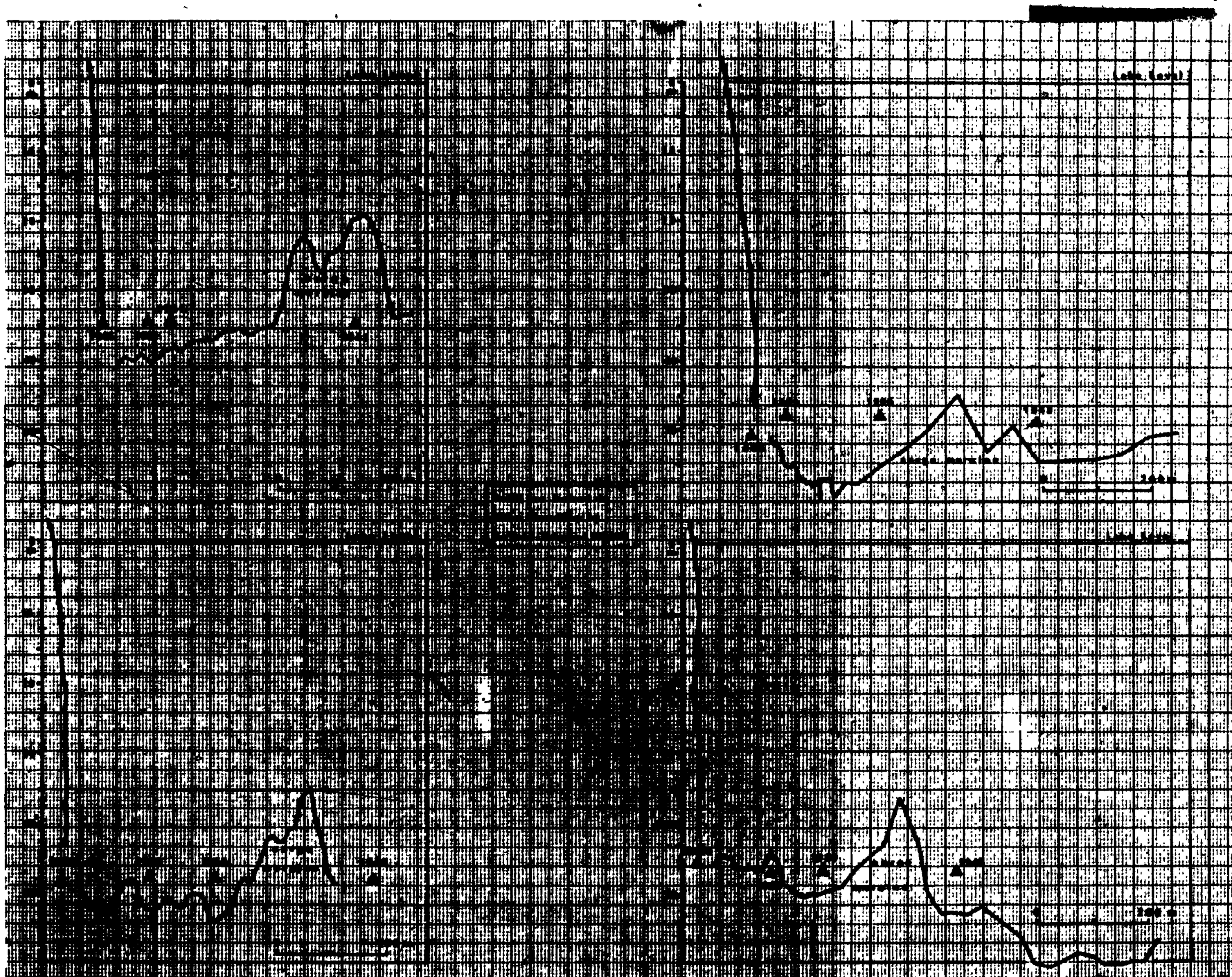

Figure 4- 6: Four profiles showing the lake bottom across the location of the surge moraine and the sublacustrine moraines between it and the ice. Triangles indicate the approximate location of the ice margin at the lake surface at the year indicated.

v) Crevassing, moulins and melt water flow: The crevasse zone, which is an uncommon feature of the Barnes Ice Cap, has allowed the development of three active moulins. Each was inspected and exhibited a near vertical drop with a corkscrew shape for an estimated 5-8 m minimum. With a view to gaining some information about the route of meltwater flow entering the ice mass, the upper layers of which theoretically would decrease in temperature from the surface downward, some Safranin dye was added to each of the three moulin streams in turn. In each case the time of entry was noted and the intent was to establish where the water would reappear and how long the journey would take, in an attempt to get some indication whether the water penetrated to the bottom of the ice or remained en-glacial. If the former condition could be established then it would be demonstrated that at that time the ice was not frozen to the bottom at that locality.

The procedure adopted after inspecting the moulins, was for an assistant to inject a dye solution at a given signal from the helicopter which immediately flew over the open water to spot any stain as well as its time of appearance and characteristics. No stain was observed and so a stronger solution was injected into the second moulin. Again, no stain was observed and so a considerably stronger dye solution was added to the third moulin. Still no stain was observed and the helicopter landed. A second assistant in an inflatable boat was patrolling the ice-free section of the lake below the cliffs and he observed a dye plume 17 minutes after injection of the first plume. A flight immediately after this revealed all three dye plumes at the lake surface, but all three were offset to the south. The alignment of the offsets was in each case obviously the result of the influence of crevasses open at the surface and shown in Figure 4-7. In each case the upwelling dye plume was localized against the ice margin. The plume water was slightly less dense at 0°C (measured at the moulin edge) than the lake water at 1°-1.5°C causing it to rise to the surface. According to the model and the observations, the water must have dropped vertically for between 5-8 m, then migrated southward and downward confined along a narrow crevasse, and close to lake level, but beyond that point involves speculation. The lack of dispersion of the dye suggests that the water was confined all the way and, therefore, did not penetrate to the lake beneath the ramp. If it did it probably would have spread while rising up beneath the gentle underslope of the tapered ramp.

Although definite conclusions cannot be reached about the detail of processes operating, some insights can be gleaned. The thermal regime of the ice, at least in the crevassed zone, permits penetration of meltwater to at least an en-glacial level. The dye plumes were so concentrated as to suggest that the meltwaters only achieved en-glacial levels along crevasses rather than subglacial penetration, although the latter cannot be ruled out.

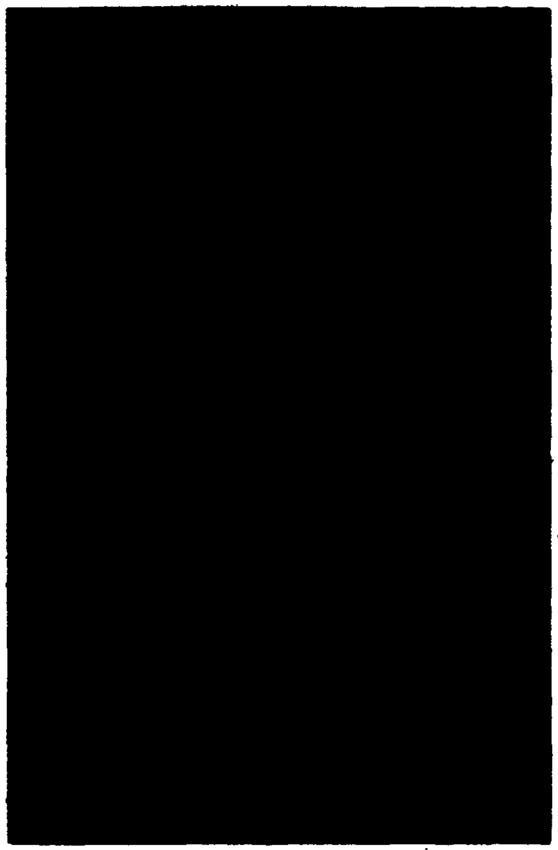

Figure 4-7: Vertical photograph of dye route and alignment of crevasses. Upwelling of dye was localized.
(GSC Photo 202522-S)

Figure 4-8: Shoreline striae on the west facing shore of Generator Lake.
(GSC Photo 168032)

By visual inspection only, the water entering the moulins was clear of sediment and none was observed in the moulin walls. Thus the process of kame construction, if it is active at the present time, would have to obtain the materials from the last few metres of flow within the basal debris zone of the ice.

vi) Striae: The occurrence of striae presents a problem of interpretation of process. An ice mass which is frozen to its bed cannot create striae. Striae can, however, be created by a lacustrine process as thrusting lake ice armed with pebbles or cobbles can produce convincing striae. This latter process can be concluded to have been operating where striae are oriented radially about the shore as noted by McLellan (1971) and Adams & Matthewson (1976) for Gillies Lake on the Bruce Peninsula in southern Ontario. Shoreline striae have been observed around Generator Lake (Figure 4-8) and are attributed to lacustrine processes. However, other striae, particularly in the inter-lacustrine-moraine swales have been noted and they are much more probably glacial in origin. As such they may have been formed at the ice margin where the thermal input of the lake has raised the glacial temperature sufficiently to allow basal sliding; or, they may represent an earlier process operating away from the frozen margin where and when the ice thickness was sufficient to reach the pressure melting point and allow basal sliding.

vii) Glacial surging: The hypothesis that the south dome of the Barnes Ice Cap had surged was first proposed by Løken (1969) and was based primarily on the morphology of the ice surface which indicated a considerable, and probably rapid, transfer of ice from the upper levels to the lower ones in a zone of appropriate shape. The toe of that surge constitutes the present ice dam. Holdsworth (1973b) confirmed the surge hypothesis by advancing more detailed glaciological evidence. The earlier morphological evidence alone could have been explained simply as a morphological adjustment of the ice surface to a deep south trending valley tributary to the valley in which Generator Lake now lies.

Evidence for the surge collected by the writer was limited to demonstrating the presence of a persistent 15 m high moraine across the floor of Generator Lake at or about the alignment of the 1948 ice margin (Barnett, Forbes & Whytock, 1970; Barnett & Holdsworth, 1974, Figure 18, p. 400; Figure 4-6). This moraine rises above lake level both north and south of the ice cliffs as an ice-cored moraine, an obviously different landform from the sublacustrine moraines.

viii) Summary: In summary, the glacial processes operating are essential elements to the understanding of the features in the rest of the lake basin. Without an appreciation of the surge the reconciliation of chronology and apparent process rates for the lake basin was difficult to establish. Without the demonstration of the tapered ramp morphology the process and environment of deposition of sublacustrine moraines was problematical. Without appreciation of the cold nature of the ice mass the glaciofluvial and thermal lacustrine responses would be difficult to explain. Many of the effects of the cold glacial processes are not immediately discernable, or more specifically, not immediately attributable to them. It is only after careful evaluation that the linkages, including the vital one of geometry, are established. However, once established it is equally clear that the polar ice environment does give rise to a distinctive assemblage of landforms. None individually can be considered unique but collectively they do represent a distinctive array of landforms with the subtle stamp of cold ice inherent in their development.

III Responses in the sublacustrine environment

Three responses will be evaluated in this section: sublacustrine moraines, lake bottom sedimentation and anchor ice sedimentation.

a) Sublacustrine moraines:

i) General: The multiple linear closely-spaced, asymmetric boulder covered moraines (Figure 4-9) were formed beneath the former lake (Barnett, 1967, p. 178) which was 76±1 m above the present lake surface. The moraines are the most striking landforms in the lake basin and are considered distinctive, if not unique, to a polar proglacial lacustrine environment. The moraines form at the

Figure 4-9: Sublacustrine moraine showing bouldery surface and slightly arcuate form. Proximal slope is to the right. (GSC Photo 161626)

lake-ice interface with optimal development beneath a tapered ice ramp (Barnett & Holdsworth, 1974, Section 7, p. 404).

The origin, chronology and morphology of these landforms was discussed by Barnett & Holdsworth (1974) (included as Appendix F) but here relevant material will be presented in terms of the model.

Approximately 140 moraines occur across the lake basin within an axial basin length of 26 km, and 132 $\pm$ 15 occur within the easternmost 19.6 km (Table 4-1). The lack of precision in establishing the number of moraines relates to uncertainty of matching discontinuous ridges where they are very small at the northeast end of the lake basin.

ii) Morphology: The maximum dimensions measured for a single moraine were 6 km long, 35 m high with maximum distal slope of 36° and maximum proximal slope of 27° but all these measurements do not apply to the same profile. The maximum moraine dimensions (till volumes) occur in the central part of the lake basin where the deepest water occurred, and decline to nothing at or immediately below the 76 m lake shoreline.

A total of 107 hand-levelled profiles show that the mean distal slope angle exceeds the proximal by about 7° (Figure 4-10). The mean cross-section areas are based on the dominant (longest) measured slope angle of a particular moraine, where one or more breaks of slope was recognized. In only 13 cases were the proximal slope angles in slight

TABLE 4-1

CHRONOLOGY, MORAINE FORMATION AND RETREAT RATES

Lake phase	Time Interval (yr)	Number of moraines	Axial retreat distance (km)	Retreat rate (myr^{-1})	Moraines per 100 yr	Average number of years per single moraine	Remarks
	4500[1]–3850 B.P.	—	7.2	10.7	—	—	Moraines nonexistent to very small
	4000[1]			42.4			
	3830–3480	75±10	9.2	26.3	21±3	5±0.6	small to medium
	3480–3090	17±2	2.6	6.7	4±1	23±2	medium to very large
	3090–2520	40±2	7.8	13.7	7±1	14±1	medium to large
I	3090–2820	24±2	3.7	13.7	9±1	11±1	medium
	2820[2]–2520	16±1	4.1	13.7	5±1	19±1	medium to very large
	3830–2520	132±15	19.6	15.0	10±1	10±1	full range
II	2520–1270 B.P. (5 phase adjustment to 3 m level; may have lasted until 1060)						deposition confined to area of lake constriction (Fig. 11)
III	1060[3](B.P.) 1948 (A.D.)	3(?)	6.0	5.7	—	—	nonexistent to very small
	1948–1971(A.D.)	3±2	0.4	17.4	13±8	8±3	very small

[1]Initiation assumed; not definitely dated but 4000 B.P. is considered more reasonable.
[2]Interpolated date.
[3]Mixed chronology. Phase includes lake draining and subsequent surge.

	SECTION		
	WEST	CENTRAL	EAST
No. of cross sections	33	34	40
Moraine Ht (m) max	35	23	20
min	4	1	1
mean	13	8	9
Slope (°) max prox	25	27	27
min prox	15	4	2
max dist	34	35	36
min dist	12.5	6	4
mean diff (dist−prox)	9	5	7
Range prox	26	23	25
Range dist	21.5	29	32
No. cases where prox slope > distal	2	7	4

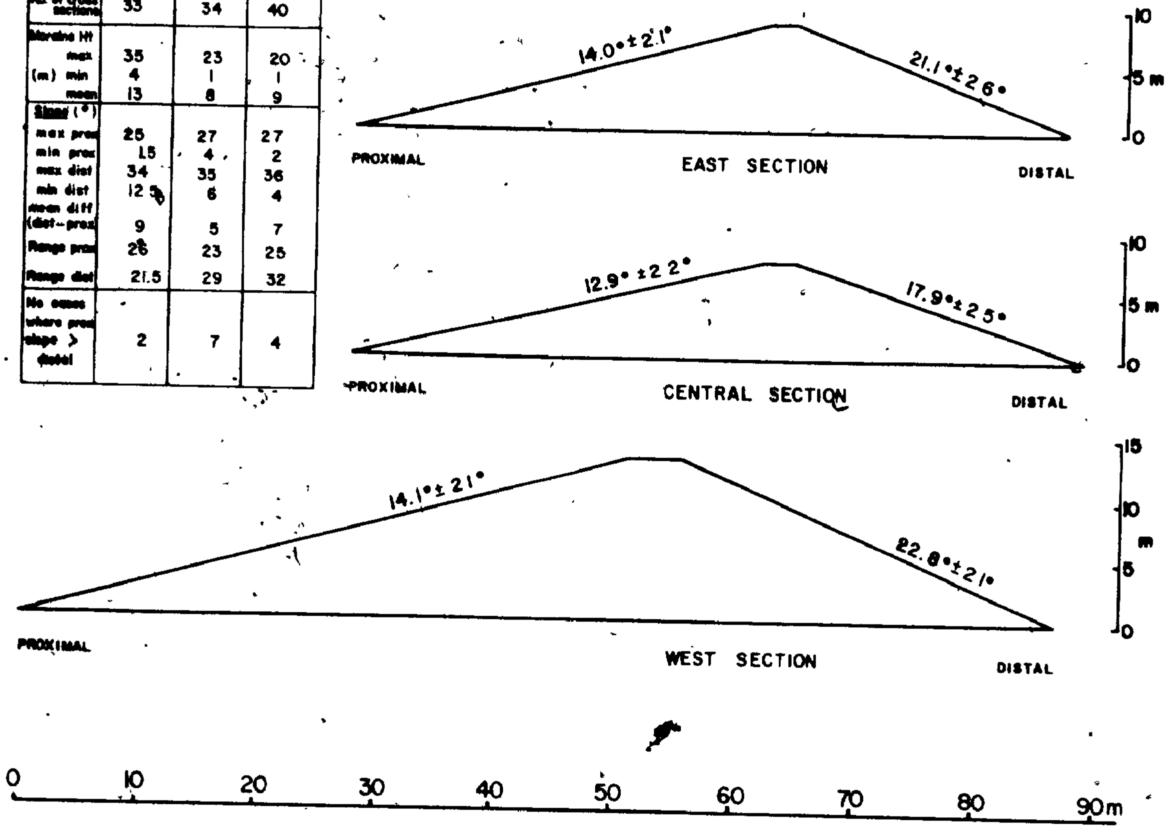

Figure 4-10: Representative profiles of the east, central and west sections derived as composites from the 107 profiles measured across selected moraines.

excess of the distal. This characteristic asymmetry must reflect primary processes as it persists both on moraines currently subject to sub-aerial and lacustrine processes. Morphological modification by ice push is limited to occasional ice pushed ridges up to two metres high which are characterized by boulders showing lichen-free faces of lighter tone than the generally darker and more stable slopes above the shore zone.

iii) Spatial arrangement and timing: The spatial arrangement of the moraines is noteworthy as they are linear to slightly sinuous and quasi-parallel but only rarely become contiguous and even then for only very short stretches. This spacing is remarkable when considering that close to 140 moraines occur in a distance of less than 20 km. A marked trend towards larger moraines in deeper water is apparent. These relationships suggest a process which is ponderous and possibly rhythmic but not one repeatedly subject to rapid minor fluctuations of the ice margin. In other environments an apparent rhythm has led to speculation on moraine formation being an annual event. Such timing was considered by Andrews & Smithson (1966, Figure 7, p. 288) as a possibility, but was not proven for the apparently similar cross-valley moraines of the Isortoq valley. Others have considered such a timing for multiple moraine ridges, for example, Hoppe (1959) for the 'De Geer' moraines in Sweden, Price (1970) for the Icelandic proglacial environment and Worsley (1974) for the Norwegian proglacial environment but in this latter case the moraines were only one metre or less in height and only a few metres long.

The chronological evidence presented from Generator Lake is unequivocal that the sublacustrine moraines are not annual features but require between 5 and 23 years for a single ridge to be formed (Barnett & Holdsworth, 1974, Table 2, p. 390) (Table 4-1). A mean value of 10± 1 years was derived for the 132±15 moraines in the 76 m lake basin.

iv) Retreat rates of the ice margin: The average net retreat rate of the ice margin for the present lake basin west of Clyde River is about 6 m/yr. This recent trend is associated with the intersection of about one half of the calving ice front with the heavily crevassed ice zone visible on the 1948 air photographs and which has ~~remained~~ fixed relative to bedrock. The 6 m/yr apparent retreat rate is lower than the complete basin average of 8 m/yr but when allowance is made for the lake draining, followed by the surge, it follows that actual retreat rates must have been much higher than 6 m/yr.

The mean retreat rate for the 76 m lake level basin was 19 m/yr based on a distance of 26.8 km along the axis of the basin and lake inception at 4000 B.P. and initial marginal drainage down the Clyde River at 2600 B.P. The mean would be only 13.4 m/yr if, though less probable, inception was as early as 4500 B.P. and drainage as late as 2500 B.P. However, a declining rate of retreat with increasing lake area (76 m lake) is discernable from the field data (using the 4000 B.P. inception) and this relationship was substantiated theoretically by Holdsworth who concluded that increased stability of the ice front was expectable with increased lake area (Barnett & Holdsworth, 1974, Section 6) due to a reduced bending moment exerted on the ice ramp. The reduced bending moment would be simply a reflection of a decreasing rise in lake level for approximately the same annual influx of meltwater into an increasing lake area. Holdsworth calculated the idealized relationship as $Z_W = \frac{70}{\bar{A}}$ where Z_W = maximum rise in lake level and $\bar{A}$ = area of lake. The measured rise of ca. 1.6 m in both 1968 and 1969 (this study) plot close to the idealized curve (Barnett & Holdsworth, 1974, Figure 13, p. 396). Therefore, other things remaining equal, a greater retreat rate for the present lake basin is expectable than for the 76 m one. This cannot be conclusively proven as the post-1948 rates are complicated by the effects of the crevasse zone. However, about 10 km of retreat took place (leading to lake drainage), a period of fluvial drainage to Foxe Basin (of unknown duration but up to several hundred years) a surge of the ice to reblock the drainage, refilling of the lake to the Clyde River outlet (more than 60 m maximum depth) and the re-initiation of retreat beginning approximately mid-twentieth century.

This theoretically based calculation for a rapid rate of retreat >20 m/yr would then explain the general absence of well developed moraines beneath the present lake, which was a previously puzzling problem. This was particularly so as the water depths and lake width were generally only slightly less than those prevailing in the 76 m lake. Therefore, simply a change in the tempo of a process, the rate of calving was sufficient to change the response from well developed moraine ridges to indefinite occasional rises on the lake bottom barely traceable by echo sounder.

v) Till volume and time: As the chronology section offered the opportunity to calculate net annual retreat of the ice margin in the 76 m lake basin and as the spatial arrangement analysis indicated a fairly steady or ponderous process of moraine formation not obviously subject to minor annual fluctuations, then examination of till volumes and time should yield valuable support or refutation for the chronology and process of moraine formation.

Till volumes at three sections 500 m wide (shown on Figure 4-2) were examined in detail by measuring multiple profiles. Each section was selected to include a least one moraine ridge with good continuity of form and well developed morphology. Section E was the longest, section C shortest, and section W intermediate in length but well developed. From these volumes a till deposition rate is derived for each section (Table 4-2). Approximations used in this calculation are: 1. the calculated average retreat rates apply over the 500 m wide strip, 2. mean cross-sectional areas at 200 m intervals (minimum 33) adequately represent till volume, 3. a planar base between end points

TABLE 4-2

TILL RELEASE AT THREE 500 m WIDE SECTIONS ACROSS THE FORMER LAKE

	Volume (m^3/m width)	No. Of Profiles Measured	Retreat Rate (m/yr)	Time (yr) For 500 m Retreat	Till Release (m^3/m width/yr)	Equivalent Uniform Depth Till (m)
West section	930.5	33	13.7	36.5	25.5	1.861
Central section	310.1	35	13.7	36.5	8.5	0.620
East section	514.4	42	6.7	74.6	6.9	1.029

Note: Location of sections shown in Figure 4-2 of each profile, 4. no significant amount of till occurs between individual moraine ridges. Bedrock was frequently observed between the moraines generally substantiating this point. 5. Sublacustrine profiles in the middle of the sections were approximated using the long profile of the lake basin, cross profiles and some lake soundings. The values derived for till volumes are, therefore, minimum estimates with no allowance made for sub-aerial erosion or inter-moraine till.

The volume values per metre width, when coupled with the time for 500 m of retreat, offer till release rates per metre width per year which range from 25.5 (W) to 6.9 (E) $m^3/m/yr$. Each value when divided by the retreat rate gives a figure for an equivalent uniform depth of till over the basin width for a 500 m section. In each case this value was less than 2 m and in the case of the central section only 62 cm.

Holdsworth measured the concentration of debris in basal ice from three samples taken from the north side of the present ice cliffs. The sample sites were above lake level from a zone of ice folded by a surge. He concluded that an average value for basal debris was $8 \pm 2\%$ in the basal 8 m only, (Barnett & Holdsworth, 1974, p. 401). Using that value range from present day conditions it is instructive to work back from debris concentration through the ice supply rate to give a sufficient volume of till to create the moraine volumes measured. This calculation is shown graphically in Figure 4-11. The first conclusion is that all three measured till volumes lie in an acceptable range of forward ice movement with the west section being very close to present day values, near 20 m/yr and the east and central sections suggesting much slower rates of ice movement generally less than 5 m/yr.

As the rate of retreat influences other calculations and is based on an assessment of the error terms of radiocarbon dates it is desirable to examine the results in more detail.

Holdsworth calculated a till release rate based on glaciological considerations (Barnett & Holdsworth, 1974, Section 5.7, p. 402) which he expressed as

8 m thickness basal ice	x	17 ± 3 m/yr basal flow rate	x	$8 \pm 2\%$ per m width debris content	$= 11 \pm 5$ $m^3/m/yr$

Therefore, using this glaciologically derived value and using its maximum value of 16 $m^3/m/yr$ in place of the largest measured value of 25.5 $m^3/m/yr$ derived from the 500 m West section, a retreat of 500 m would take 58.1 yr which yields an annual net retreat rate of 8.6 m/yr (instead of 13.7 m/yr). Using this new value between the ice margin locations represented by dates GSC 1304 and GSC 1276 (7.8 km) it would require 900 years at 8.6 m/yr assuming the same ice flow rate of 16-20 m/yr. This time span lies on the 95% confidence limits (laboratory statistics) of the two dates 3260-2370 = 890 years. Therefore, agreement on average till release rates calculated

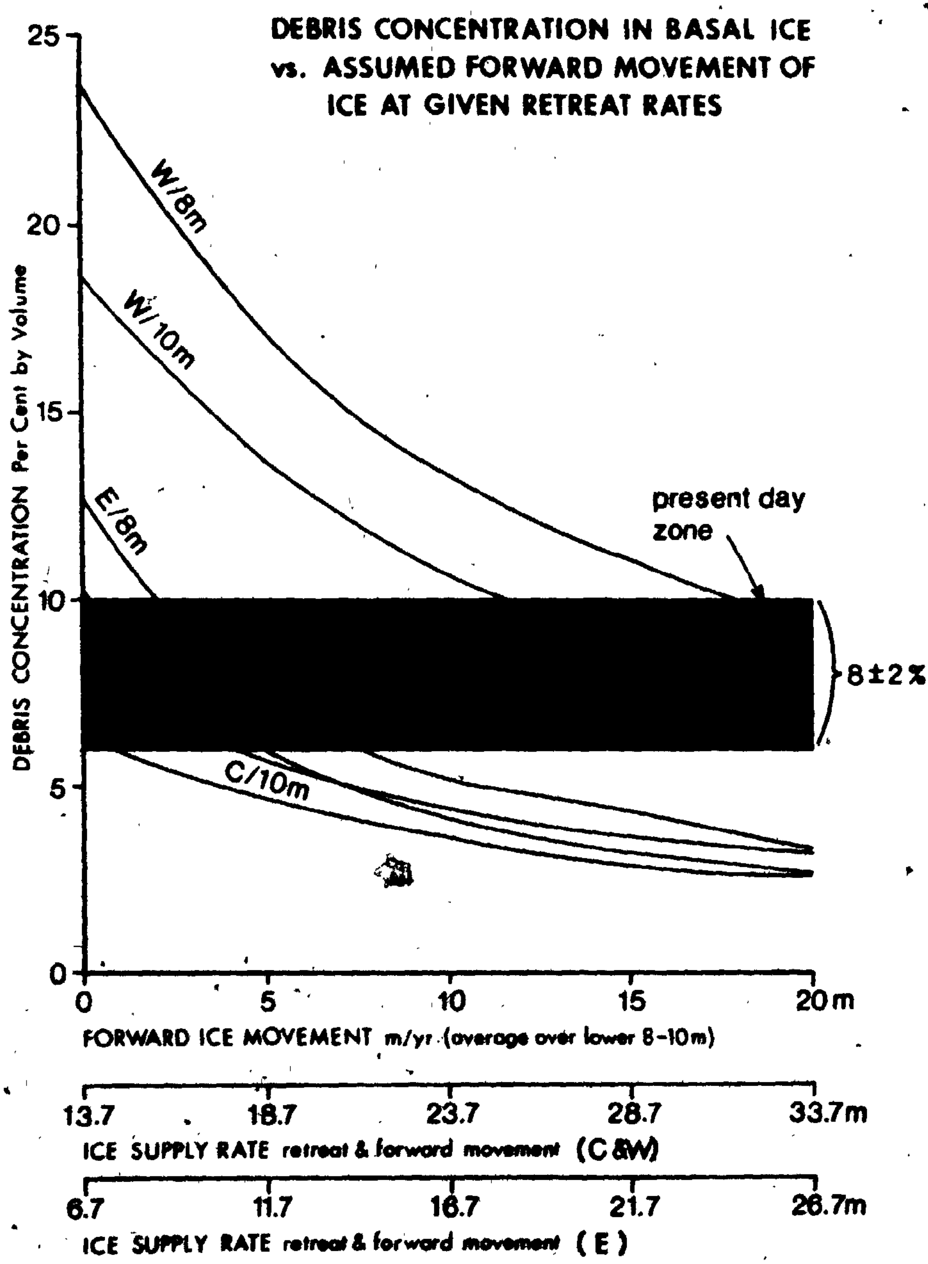

Figure 4-11: Plot of debris concentration (% by volume) in the basal 8 and 10 m of ice versus forward ice movement averaged over the lower 8-10 m of ice. For the sections C, E and W, the ice supply rate is the sum of the average retreat rates and the forward ice flow rate.

independently is a further case for accepting that the chronology is of the right order. This calculation simply demonstrates that an adequate volume of till can be produced but says nothing of its organization into ridges. Nevertheless, glaciological constraints can be met to produce sublacustrine moraines of an appropriate size in an acceptable time, at a retreat rate which is readily compatible with rates elsewhere in the lake basin.

vi) The process of sublacustrine moraine formation: Thus far the general environment, timing and till volumes have been identified and discussed but the key process has not. Obviously, some till release mechanism operates which concentrates it into a ridge and then the process slows or ceases for a period and begins again to form another ridge. The hydrological-glaciological relationships are shown to be the controlling factors (Barnett & Holdsworth, 1974, Section 7, p. 404) and are summarized as follows.

Each spring rapid run-off of snow melt leads to a rapid rise in the lake level by mid-July. The larger the lake the less rapid this rise is liable to be. In 1968 and 1969 the rise was approximately 1.6 m. In fact, a hyperbolic relationship between lake area and water level rise was proposed by Holdsworth (Barnett & Holdsworth, 1974, Figure 13). From this point Holdsworth examined the relationship between mean moraine height and lake area and found it to be linear with a correlation coefficient of 0.90 (Barnett & Holdsworth, 1974, Figure 16). From these two relationships he went on to examine mean moraine height and water level rise and produced a hyperbolic plot with the form $\bar{h}_m = \frac{5.2}{zw}$ (Barnett & Holdsworth, 1974, Figure 17). From this relationship it appeared worthwhile to seek a physical connection between moraine size and lake level fluctuations.

Undercutting of the ice margin occurs (probably year round) as the water at lake bottom is in the 1°-1.5°C range (Barnett, Forbes & Whytock, 1970). Where the ice is cliffed undercutting (Figure 4-4) leads to collapse which may be gradual or sudden as witnessed on July 4, 1968 (Figure 4-3). Where the ice is ramped (Figure 4-12, approximately half the present margin) which is coincident with the deeper water (ca. 40 m or more), failure is a gradual process culminating in fracture after repeated flexing induced by water level rise and fall. Holdsworth was fortunate to witness just such a fracture on July 17, 1970. This occurred 3½ hours before the lake level maximum for that season (Barnett & Holdsworth, 1974, Figure 12). This peak was higher than those of 1968, 1969 or 1971 by several tens of centimetres and, therefore, the bending forces were correspondingly greater than in each of the other years.

The iceberg which resulted from the July 17 fracutre tipped backwards towards the ice cliff thus demonstrating the presence of substantial undercutting (space to tip into) and a minimum extent for it. It also offered an opportunity to calculate the geometry of the berg which was shown by Holdsworth (1973a) to be, in section, markedly tapered to-

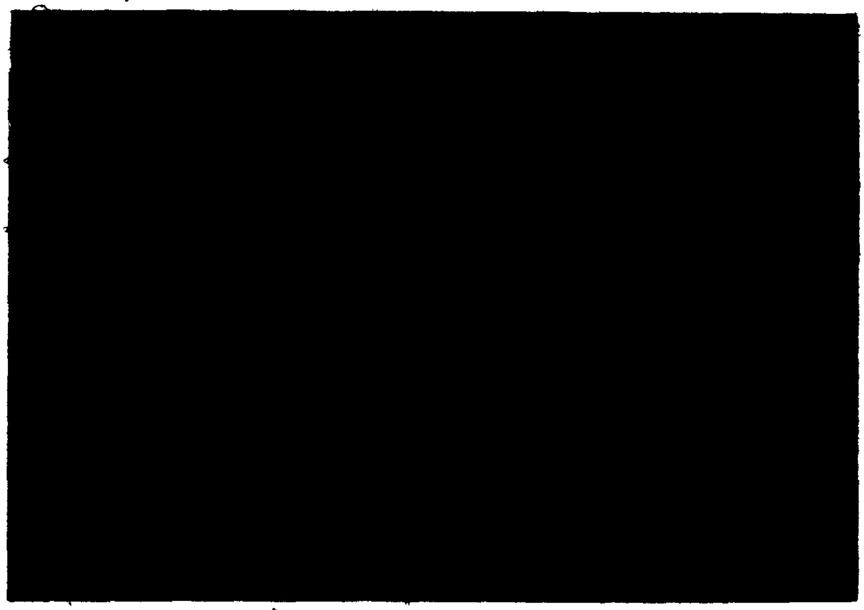

Figure 4-12: Ramp form of ice margin. Tapered in cross-section with substantial undercut. Subjected to bending as lake level rises and falls.

wards the lake and to have an underwater ram (projection) (Figure 4-5). This demonstration of geometry of the iceberg shed some light on the cliff profiling results run by the writer which suggested imperfectly that the ice was protruding further beneath the lake than at the surface.

These findings suggest an environment which is stable for a number of years then is changed by ice ramp fracture to a less stable form which becomes progressively stable again as the cliff form elongates into a ramp form. The tapered underside of the ice margin is compatible with moraine formation both in shape and due to the fact that it is the basal debris-rich ice which is melting to form the underside of the taper. The proximal slope angles of the moraines measured 12.9°-14.1° (Figure 4-10) are similar to those of the underside of the tapered ice.

The current ice margin is characterized by three or four small ramps and not a single one as might be expected for the formation of the long single moraines characteristic of the west end of the 76 m lake basin. The present conditions are at least slightly atypical as the ice front is encountering a crevasse zone and is also in a post-surge readjustment phase. Given the more stable conditions of small annual lake level changes of the mid-to-late phases of the 76 m basin then a single and simple ramp form is probable. Where water depths of less than 40 m occur the ice cliff would be less stable than the ramp form due to slower forward movement of the ice but would still experience undercutting from the warmer lake bottom water and therefore, smaller, generally less well formed moraines may be expected. Such morphological development is certainly the case in the 76 m basin.

The till fabric measurements made on the selected moraine crests only constitute samples which are insufficient in number to allow rigorous analysis of the kind advocated by Andrews (1971), but do yield generally supportive evidence for the model of moraine formation. Clasts with a a:b axis of >2:1 exhibit preferred orientation parallel to ice flow and orthogonal to the moraine crests. This fabric is consistently stronger on the proximal than on the distal slopes. Dips of pebbles also exhibited clustering. On the proximal side an up glacier dip clustering close to the surface slope angle occurs, and on the distal side a down glacier dip predominates with a secondary mode in an up glacier attitude. This dip pattern is similar to one of those found by Andrews & Smithson (1966, Figure 6 - pattern 6) in 10% of the till fabrics in the Isortoq valley. A qualitative conclusion was that the distal slope was less compacted than the proximal.

The inferences which may be drawn from the Generator Lake fabrics are very limited viewed in isolation but together with the other evidence add to the case for the formation of sublacustrine moraines. According to the model at the moment in time when the fully developed ramp of ice breaks off and a new sequence of events starts, slightly accelerated glacial ice movement is predicted in deep water (>40 m). The tempo of this change relative to that of the removal of the broken ramp and that of the wasting of the new margin is critical as to whether or not the accelerated ice flow will be sufficient to advance over the wedge of till (moraine) deposited under the detached ramp and possibly remould the material to give a stronger fabric and greater compaction on the proximal side. Alternatively, the wasting processes are more rapid than the acceleration of glacier motion and net retreat of the margin continues at a decreasing rate allowing a gradual development of the ramp as basal thawing continues and ice motion accelerates. The latter appears more feasible geometrically to account for moraine spacing.

This process and process rate adequately explain the position, morphology, frequency and spatial arrangement of the sublacustrine moraines.

b) Lake bottom sedimentation:

i) General: This response is one related to lacustrine processes exclusive of the direct glacial influence. Fines in the silt and clay range are deposited from suspension onto the lake bottom. They were sampled by Dietz-Lafond grab sampler at one kilometre intervals parallel to the axis of the present lake basin and extending northeastward into the still inundated parts of the 76 m basin. A total of 20 samples were collected by grab sampler from the sites shown on Figure 4-13. In addition a Phleger gravity

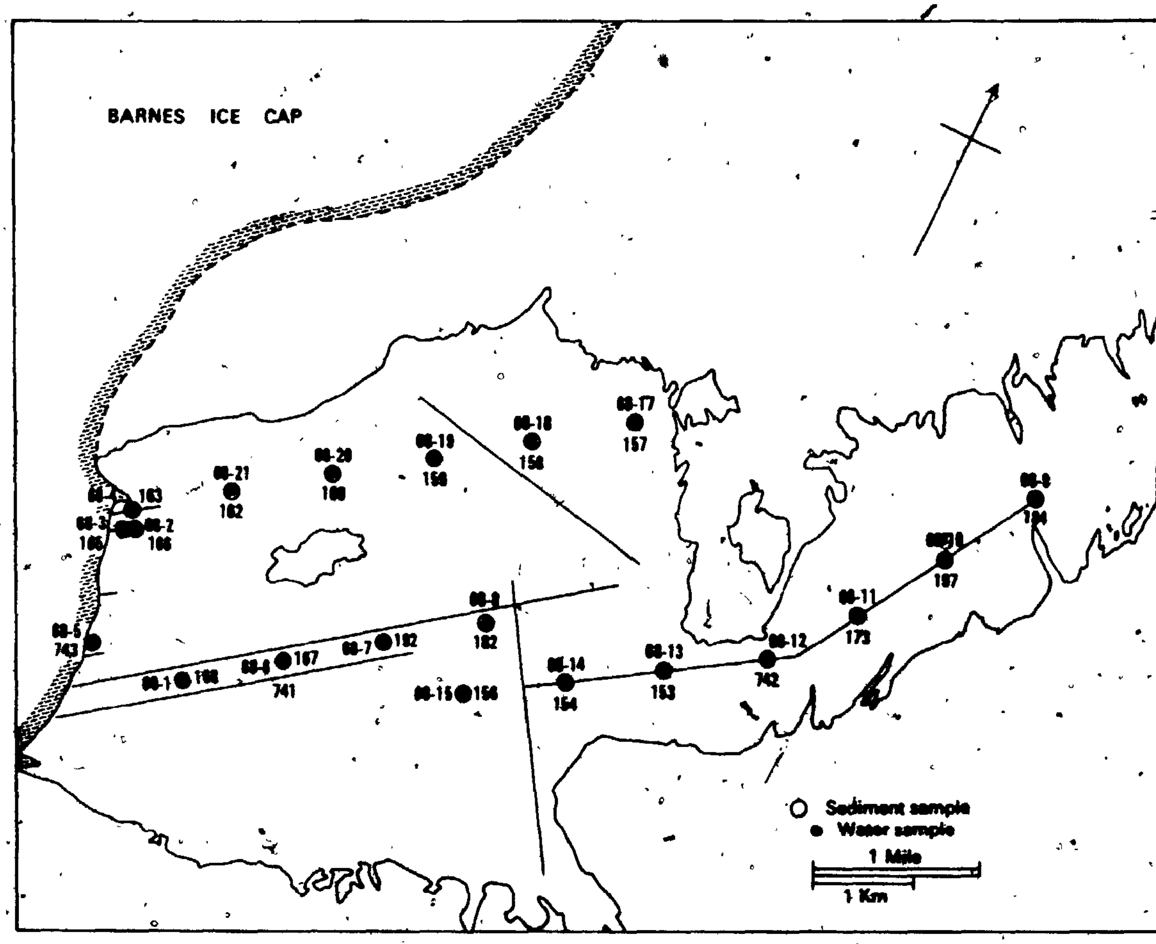

Figure 4-13: Map of Dietz-Lafond grab sample sites. Water samples and lake bottom temperatures also collected at same sites.

corer was used at five of the same sites in an attempt to obtain a stratigraphic record of lacustrine sedimentation. After very limited success this was discontinued as several corer cutting edges were damaged and cores of 10 cm or less were obtained.

ii) Results: The complete grain size analyses are set out in Appendix E. The mean sand-silt-clay ratios for the 20 samples are: 8:30:62. It is clear that the samples with the highest sand and silt fractions were collected from the sites closest to the present ice margin.

At the time of sampling it was anticipated that a deposition rate or rates might have been obtainable from the data, particularly if stratified, in view of the expected metachronous age of the bottom sediment. The later recognition of the lake draining episode, followed by the glacial surge and lake refilling has considerably complicated and diminished that expectation. Fluvial conditions for possibly up to 800 years (ca. 850 B.P. to 50 B.P.) would have enabled previously deposited silts and clays to be eroded from the lake basin from below the present shoreline. Substantial erosion of these lake bottom sediments is not difficult to envisage in the environment where at least one and probably four substantial draw-down events occurred with the lake leaving saturated silts and clays to subaerial processes on moderate slopes. Mass movement plus the annual sheet wash effects of snow melt would act quickly on the unconsolidated fines. Evidence for this evacuation of fines should show up in the stratigraphic record in Flint Lake (Figure 2-1) marking the catastrophic drainage of Generator Lake. It is noted that the silts and clays are almost completely eroded from the zone between the present shoreline and that at 76 m. (Only one fine grained sediment sample BDA-68-16 was collected from a local depression in the exposed zone with a sand-silt-clay ratio of 20:72:8).

If extensive erosion of silts and clays did take place from 850 B.P. to 50 B.P. then lake bottom sediment sampled may well have been deposited during the period 1900-1968 A.D. Maximum thickness of sediment encountered was approximately 15 cm and was unstratified by visual inspection. It was grey in colour with a hint of iron staining in one or two samples. No ice crystals were observed despite careful search for them. This is an important observation as the Phleger gravity corer yielded very short sediment cores and several badly bent cutting rims resulted from the various drops. The damage to the cutting rims could have been caused by impact on: 1) isolated rocks, 2) frozen ground, 3) bedrock or 4) any combination of the three. The first can be dismissed as no pebble or rock fragments were found in any of the grab samples. The second is improbable in view of the lack of frozen material which the corer would have at least chipped on impact. Therefore, the third is the most probable and yet that suggests that only some 15 cm of sediment occurs over bedrock. The time period for deposition is more than 20 years and possibly 60-70 years.

Such thin accumulation of sediment seems improbable for such a proglacial environment over the timespan of more than 1000 years which have elapsed, without allowing for the fluvial interval postulated to have preceded the glacial surge. Embleton & King (1975, p. 551) in a discussion of varves or couplets as lake floor deposits consider 1-5 cm to be 'the normal range of thickness' but admit to as low as 'less than 1 mm'. Even at the low end a 1000 yr interval should yield nearly one metre of sediment. This indirect evidence derived from samples along a 4 km long profile line strongly supports the concept of a lengthy fluvial interval and recently reestablished lacustrine conditions for the recent history of the lake basin.

iii) Lake bottom sedimentation: Anchor ice sediment rafts: Close to shore the lake waters freeze to the bottom. As the melt progresses in early July the main body of lake ice floats, a shorelead is formed and some nearshore anchor ice remains frozen to the lake bottom with lake water above it. Depending upon the warmth of the season, the anchor ice on the bottom may float free when the surrounding water reaches a sufficient temperature to release it as a raft. In a very cool season, it may thaw only partially and remain as a nucleus for additional anchor ice the following season.

Where anchor ice occurs close to a strong sediment source, accumulation takes place over the anchor ice and any subsequent release and floating of the ice will cause disruption and redistribution of the overlying sediment, particularly if and when the raft capsizes.

This sediment is distinguished from the generally much smaller amounts of sand brought down by the initial phases of snow melt which are deposited on the often snow covered lake ice surface and soon fall into the lead.

Several anchor ice sediment rafts were observed (Figure 4-14) and sampled in both 1968 (July 31-August 1) and 1969 (July 14) and all occurred on the north side of the lake close to the main streams draining from the ice cap. One was observed surfacing. Rafts large enough to beach a boat against and to stand on were observed (ca. 10 m long and 2 m wide with ice at least 25 cm thick). Accumulation of up to 28 cm of coarse sand sediment was measured and sampled; mean sand-fines ratios of 95:5 resulted from three samples. The material which was poorly bedded was observed gradually being washed off the raft margins by wave action but at some point instability would occur and the raft would tip dropping the remaining sediment onto the lake floor. The location of this event (not observed) would depend upon the shape and size of the shorelead and the wind direction, but it is easy to foresee that at least occasionally irregular deposits of coarse sand would occur among the finer grained bottom materials. Volumetrically these deposits may not be particularly significant but as indicators of sediments from a former proglacial lake environment they may be diagnostic of thermal regime and their distribution be diagnostic of the source(s) of the most competent

Figure 4-14: Mid-section of anchor ice sediment raft, August 1968. Note wave-washed 'shoreline' across the sediment and substantial background snow cover. Sediment thickness 20-28 cm, width 1-2 m and length 5-6 m.

proglacial streams: i.e. a plot of coarse sand accumulations may show an irregular fan shaped distribution emanating from one or more locations along a former shoreline - giving some idea of relative strength of sediment sources.

During the cooler 1968 summer much larger accumulations of sand (28 cm vs. 10 cm) were noted on the rafts which floated about two weeks later than in 1969 indicating that this process of sediment transfer and subsequent deposition is more effective in a moderately cool summer than a warm one when the rafts will float much sooner with the probability of less sediment. However, in a warm summer they may drift further in a larger lead before capsizing.

The distinctiveness of this element of the model is the irregular transfer of nearshore sediment to an offshore location and a catastrophic type of deposition of perhaps several centimeters of sand in an environment where persistent deposition rates are measured in millimetres. These circumstances are dependent upon the presence of anchor ice which requires sufficiently cold temperatures to initiate and sufficiently warm temperatures to thaw, abundant sediment available as load and rapid sedimentation. The location and geometry of the response is in turn dependent upon the lake bottom morphology and the wind speed and direction. The landforms, the ephemeral anchor ice sediment raft and the more lasting, but unobserved, sand deposit which could be like a small isolated kame in form, are, if not peculiar to the polar proglacial environment, then they certainly exhibit strong development within it.

Andrews (1963, p. 61) noted several small kames, usually centrally located, in the Rimrock valley and in the adjacent Isortoq valley. They were described simply as isolated knolls of sand (page 71). No dimensions were given. His explanation was somewhat speculative and invoked melt-water penetration to the base of the glacial ice despite the general expectation of a cold ice thermal regime which theoretically precludes water penetration to the base. Anchor ice sediment rafts offer a simple process for emplacement of irregular poorly bedded sand piles not considered by Andrews.

IV <u>Hydrological processes and responses</u>

There are two components to the hydrology: the fluvial and the lacustrine, and both are heavily dependent upon temperature for their efficacy. The fluvial aspect will be examined first.

a) Fluvial processes and responses:

i) General: One aspect of the lake basin which is significant for processes and is characteristic of this high latitude environment is the very high percentage of water which runs off with little percolation possible, because of the seasonally frozen ground, permafrost and proximity of bedrock. Sparse vegetation does little to impede run-off and demands little moisture for evapotranspiration. Despite this as was pointed out in Chapter 2 the drainage density, as plotted on the 1:50 000 topographic map, is extremely low at 0.951 km/km^2. This is indicative of a very youthful environment with a drainage network poorly organized. Internal drainage occurs and numerous small ponds result. Relatively frequent drainage changes are expectable in such a dynamic environment and at least one locality (Figure 4-2 'dd') land-based drainage diverges and follows two separate routes. In such circumstances a single high magnitude drainage event could establish a single routing along the shorter course.

All streams are seasonal only and are short, ranging up to 18 km at maximum (including the supraglacial sections), with generally steep gradients, exhibiting turbulent flow and several stretches with rapids. Streams on the northwest side of the lake are longer and steeper than those on the south and east. The thalwegs of five are shown in Figure 4-15 and in cross profile in Figure 4-16. The seasonal flow hydrographs also differ particularly in warm summers (Figure 4-17). Following the snowmelt flood the southwest side streams reduce to quiet flow occasionally augmented by rainfall but the northwest side streams are very sensitive to temperature and insolation such that a warm sunny day can produce snowmelt magnitude events along the major streams. The responses to these processes are readily visible both in ephemeral and longer term modes. Snowmelt flood entrains considerable load particularly on the northwest side of the lake where the streams are eroding the shear moraine zone of the Barnes Ice Cap. Following snowmelt

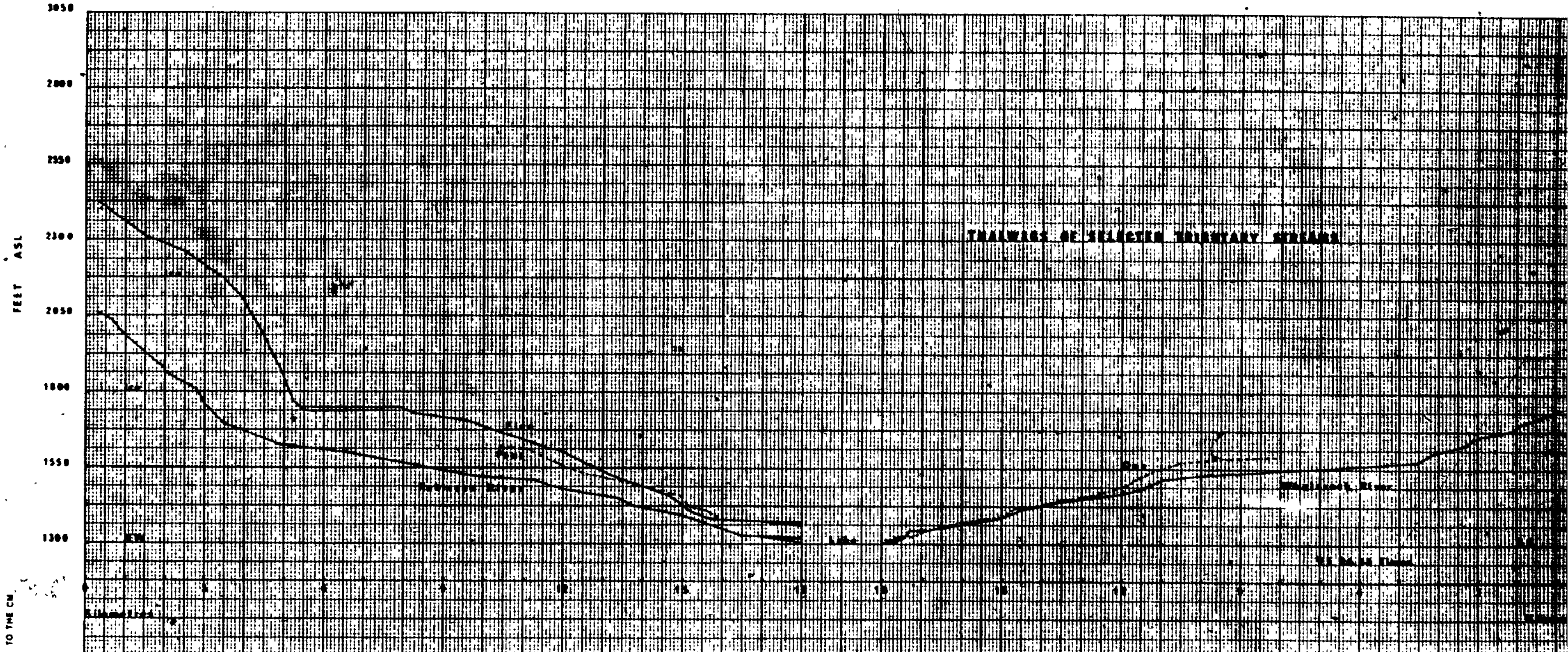

Figure 4-15

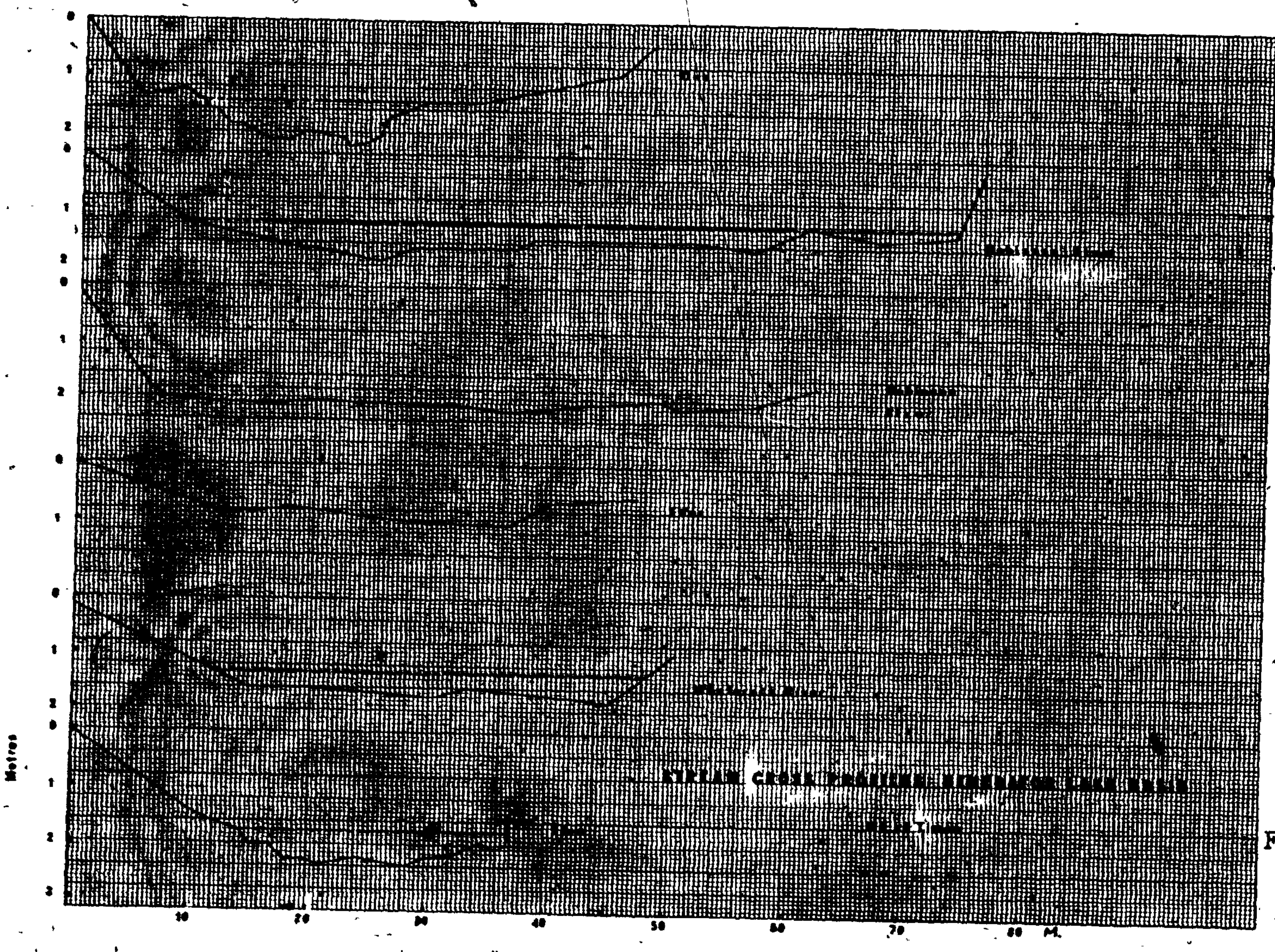

Figure 4-16

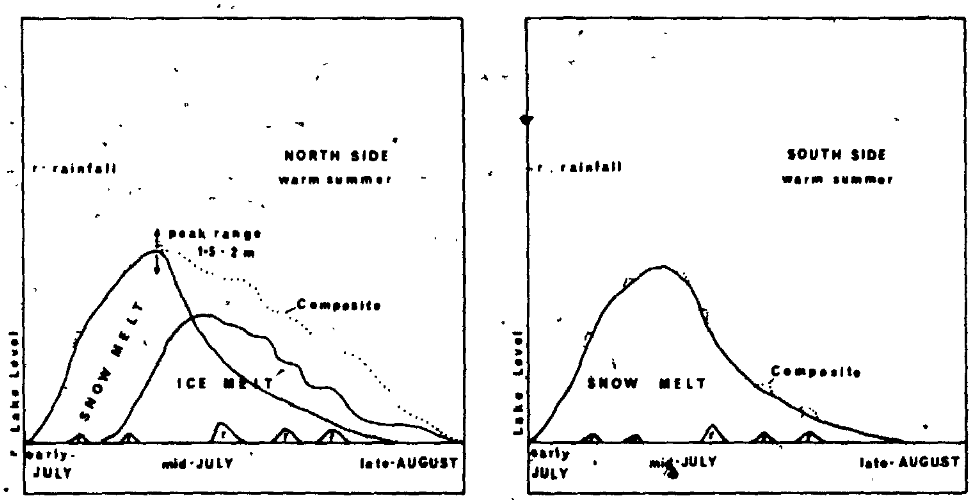

Figure 4-17: Schematic flood hydrographs for north side and south side streams during a warm summer.

considerable sensitivity to temperature and insolation is maintained by the northwest side streams with corresponding variations of entrainment of load. There is one exception to this and that is the stream at point 's' on Stream 5 (Figure 4-2) which has a lake acting as a settling basin which traps almost all the bedload. Visual inspection was made of both that stream plus the 'Robinson' River immediately to the southwest on the same afternoon during a warm sunny period (air temperature approaching 14°C). The former was running in spate but sparkling clean and clear and the other, also in spate, running brown and opaque with heavy suspended load entrained. Such processes result in responses such as the absence of substantial deltas on the former stream (except at the 76 m level) and in the case of the second stream the presence of several deltas from which datable material was taken. The exception of the 76 m level is indicative of the absence of the small lake acting as a settling basin thus requiring that the ice margin extend slightly further east at the period around 3090 B.P.

ii) Quantitative fluvial data:

1) Discharge and stage: In order to quantify some of these observed processes and their responses, samples of discharge, suspended sediment, stream temperature and flow-travel time were measured. These were gathered to obtain sample data only, in order to a) estimate a range of present process intensity, b) establish whether the measured rate of

present processes and the radiocarbon chronology could be made reasonably compatible to result in responses of the magnitude observed, c) to assess in a very preliminary way the thermal contribution of the streams to the lake.

Discharge was measured on three northwest side streams, two southeast side streams and on the overflow channel of Generator Lake (Clyde River). The measurement sites are indicated on Figure 4-2 and the cross profiles on Figure 4-16. Measurements were made as frequently as possible but on an irregular basis. The most measurements were made on the 'Robinson' River where discharge was measured by an Ott current meter on four occasions between July 9, 1969 and August 22, 1969, and stage recorded on 21 occasions (Table 4-3), from which approximate flows can be derived. Other stage and flow rates are presented also on Table 4-3.

In the case of the Generator Lake outlet only one discharge rate was calculated (due to relative inaccessibility) but this stage when combined with lake stage data offers a complete water level record for the 1969 season together with close estimates for 1968 and 1970. From the discharge figure and the stage record estimates of peak discharge and minimum flow can be made for each season. This is possible because of the reservoir effect of Generator Lake dampening out the diurnal fluctuations of the proglacial streams.

2) Suspended sediment data: A depth-integrating DH-48 suspended sediment sampler was used to estimate sediment transport on each of the occasions when discharge was

TABLE 4-3

STAGE RECORDS MONITORED STREAMS

Stream name	Date (1969)	Stage (cm)	Time (Local)	Discharge (cfs)	Mean Temp. water (°F)	Velocity (m/sec)
One	Jul 7	0.0		195.2	--	0.6228
	Jul 8	-3.4	11.50		--	
	Jul 13	-4.5	14.00		-	
	Jul 13	-2.0	15.30	117.29	-	0.6006
	Jul 29	-27.0		13.05	62.0	0.0668
Whalerock River	Jul 10	0.0		231.66	-	0.6793
	Jul 15	-8.0	10.30	165.92	42.3	0.4894
	Jul 29	-13.0		105.44	58.1	0.3110
Four	Jul 12	0.0		9.82	-	0.2782
	Jul 18	-0.3	17.50	11.78	48.1	0.3380
	Aug 5	-7.0		8.03	53.8	0.3285
Five	Jul 18	0.0	13.05	77.69	42.2	0.2681
	Jul 18	1.8	16.20		-	
	Aug 5	3.0	11.30		-	
	Aug 5	6.0	14.30	86.93	44.3	0.2639
Clyde River	Aug 11	0.0		2010.00	37.1	0.9854
	Aug 19	-20.0		1700.00	-	
Robinson River	Jul 9	5.0		355.62	-	0.5664
	Jul 14	0.0	13.30	313.24	42.0	0.5102
	Jul 16	1.2	11.10		-	
	Jul 16	0.8	17.40		-	
	Aug 8	3.0	18.00		42.0	
	Aug 12	-15.5	10.25		40.7	
	Aug 12	-14.0	10.42		41.0	
	Aug 12	1.5	18.20		39.5	
	Aug 13	-10.5	11.05		38.0	
	Aug 13	-5.5	17.24		36.6	
	Aug 16	-19.0	10.40		36.5	
	Aug 16	-5.5	15.00		40.0	
	Aug 18	-20.0	10.50		38.0	
	Aug 18	-9.0	18.05		40.0	
	Aug 20	-19.5	11.45		40.0	
	Aug 20	-8.0	17.18		42.0	
	Aug 21	-20.5	11.00		39.5	
	Aug 21	-6.5	16.43	517.66	42.6	0.9107
	Aug 22	-21.0	10.55		37.1	
	Aug 22	-20.0	12.08	245.66	-	0.5278
	Aug 22	-9.0	16.50		42.0	

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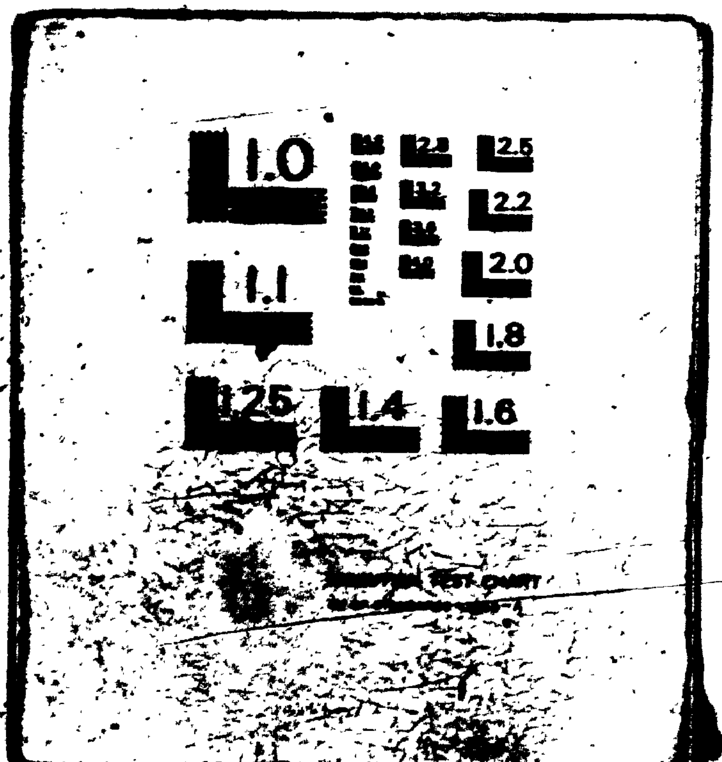

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measured. The data are set out in Table 4-4. It is readily apparent from this table that the amount of sediment in transit on the northside streams is considerably greater than for those to the south. The effect of Generator Lake as a settling basin is apparent from the low value at the outlet.

This part of the sampling program was to compare amounts of sediment in transit with the volume of the raised deltas on the 'Robinson' River and their apparent period of deposition from the radiocarbon chronology and evaluate the figures accordingly. Obviously a number of uncertainties are inherent but any agreement on the order of magnitude should be considered very encouraging.

The volume of the deltas along the 'Robinson' River (Figure 3-4; located by radiocarbon data annotations on Figure 4-2) was calculated in part using photogrammetric data derived by Mrs. G. Mizerovsky and in part from field measurements. A planar basal surface for the deltaic sediment is assumed and the dissected portions of the deltas were 'filled in' for the purpose of some calculations.

Two approaches are followed - first to derive rates of sedimentation using the radiocarbon chronology and the delta volumes and secondly using a mean sedimentation rate and the delta volumes to generate a chronology. In both cases annual flow regimes of 70 and 80 days are assumed.

Delta volumes of sediment are set out in Table 4-5. Values are rounded to the nearest 500 m^3 and include

TABLE 4-4

SAMPLE SUSPENDED SEDIMENT DATA FROM DH-48 SEDIMENT SAMPLER

Sample No.	Stream	Date (1969)	Discharge (cfs)	Sediment (g./m³)	Location	Comment
1 F	one	July 7/69	195.20	21.92	south	
5 F	one	July 13/69	117.29	9.21	south	
11 F	one	July 29/69	13.05	1.38	south	
2 F	Robinson River	July 9/69	355.62	156.54	north	
6 F	Robinson River	July 14/69	313.24	66.77	north	streaming flow
15 F	Robinson River	Aug. 21/69	517.66	45.68	north	shooting flow
16 F	Robinson River	Aug. 22/69	245.66	3.44	north	streaming flow
3F	Whalerock River	July 10/69	231.66	14.00	south	
7 F	Whalerock River	July 15/69	165.92	1.35	south	
10 F	Whalerock River	July 29/69	105.44	1.95	south	
4 F	four	July 12/69	9.82	10.04	north	not fed by ice cap
9 F	four	July 18/69	11.78	6.29	north	not fed by ice cap
13 F	four	Aug. 5/69	8.03	15.66	north	not fed by ice cap
8 F	five	July 18/69	77.69	2.60	north	lake sediment trap upstream
12 F	five	Aug. 5/69	86.93	0.83	north	lake sediment trap upstream
14 F	Clyde River	Aug. 11/69	2010.00	4.13	south	overflow channel

TABLE 4-5

DELTA SEDIMENT VOLUMES AND CHRONOLOGY

Delta	Height (m)	Sediment Volume remaining (m^3)	Approx. Original Volume (m^3)
1	76	490 000	2 000 000
2	55	20 500	28 000
3	23	50 000	116 000
4	12	180 000	404 500
5	3	51 000	70 000
Present	0	138 000*	N/A
		929 500	

* Lower order of accuracy as submerged.

Total sediment deltas 2 - 5 incl. 301 500 m^3:2600-1240 BP

1360 yr.

Delta	% of sediment in system	Time (yr)	Time BP
2	6.8	90	(2600-2510)
3	16.6	225	(2510-2285)
4	59.7	810	(2285-1475)
5	16.9	230	(1475-1245)
		1355	

301 500 m^3 sediment present in system (= minimum value)

221.69 m^3/yr.

Based on 80 day flow season = 2.77 m^3/day = 85.44 g./sec.

(minimum mean value).

2 000 000 m^3 estimated total original volume of delta 1 and

up to 400 yr for deposition (3000-2600 BP) 5000 m^3/yr.

Based on 80 day flow season = 62.5 m^3/day = 1922.4 g./sec.

(maximum mean value).

Note the agreement in magnitude with current range of values in Table 4-4.

was just over 2.5°C on August 8, 1969 despite the persistence of lake ice.

In the case of the 'Robinson' River one further investigation of process was made to evaluate the response of stream water reaching Generator Lake often at 4°C or more. These temperatures were achieved flowing over a distance of 7 km from the ice cap margin. The temperature of the supraglacial stream was confirmed to be 0°C where it leaves the ice and begins crossing the debris-laden ice zone. The flow-travel time of the water from ice margin to the lake was established by timing the flow of Safranin dye plumes. This was done in six phases to ensure easy recognition of the dye plume at the check points. All measurements were achieved in the space of a few hours and the composite elapsed time was 2 hr. 16 min. for a 7 km flow. This was during a cool cloudy period and so the increases in temperature of 0.36°C/km and 1.1°C/hr. are attributed primarily to friction (Morisawa, 1968, p. 36). In this case irregular geometry was exerting influence to warm the water entering the partially ice covered lake.

From these few measurements some inferences may be made about the processes operating. Thermal input into Generator Lake is asymmetrical in degree although perhaps not in amount, with warmer water arriving from south and east side streams but more water from north side streams. Some of this heat will contribute to melting the lake ice cover. Some of the water at or about 4°C at maximum density should delta 4, for example. A case can be made for this apparent change in process intensity to account for the larger response. At the time of initiation of the 76 m delta the stream was truly ice marginal, which through time has become progressively less so, as the ice margin retreated and other substantial streams now drain to the west of the 'Robinson' River. Therefore, a former greater stream volume with a longer reach in the debris-rich ice margin could reasonably have given rise to substantially greater sedimentation rates.

Next, sedimentation rates are examined quantitatively, albeit with some necessary assumptions about 1) the ratio of suspended load to bedload, 2) an overall uniformity of rate for the deposition of deltas 2 to 5 inclusive. Taking the 301 500 m^3 of sediment present in these deltas, an average eighty day flow season, a minimum rate of deposition of 2.77 m^3/day is derived, which is 85.44 g./sec. (density of 2.67). This figure is easily sustainable from suspended load only (3 of the 4 measurements on the 'Robinson' River are well above that figure and one well below). The measured range for suspended load only was from 1762.6 g./sec. to 25.1 g./sec. with a mean of 762.6 g./sec.

In the case of the 76 m delta the rate of deposition based on a time span of 400 years (80 days per year) and a volume of two million cubic metres offer a guide to a maximum rate of sedimentation. This calculation produces figures of 62.5 m^3/day which is 1922.4 g./sec. This figure only just exceeds the high of 1762.6 g./sec. measured on the present stream but that latter figure is for suspended load only and for a diminished stream with lesser available load than at time of deposition. No data are available on bedload but some broadly comparable conditions have been carefully documented by Church (1972) from a proglacial situation at the northwest end of the Barnes Ice Cap and from Ekalugad Fiord some 125 km to the southeast. In both locations Church (1972, p. 61) found bedload to range up to 90% of total sediment load, with a low figure of 13% at a site where 'total sediment transport was almost nil' (page 61). A lengthy discourse on bedload is unjustified in this context but as the writer has visited both sites examined by Church, some contrasts should be made. The sandurs studied by Church contain a much wider range of size and much higher percentage of large calibre material than is present in the deltaic sediments in question. Therefore, the distinction between bedload and suspended load is easier to make in the former case, but in the case of Generator Lake the bulk of the transported load is in the sand size range.

In conclusion measured sediment transport should be increased by a minimum of 13% but probably not of the order measured in the sandur regime. Given the broad and generally shallow course of the 'Robinson' River it seems likely that some of the sediment sampled as 'suspended' could be considered equally well as bedload in the saltation mode

(Church, 1972, p. 55) particularly at high flow rates.

The essential findings of the sample results are that the chronology and sedimentary regime are compatible in tempo and magnitude. The current processes are capable of producing the former responses in the time available, taking into account the changing geometry of the lake basin. A case can be made for a diminution of both discharge and load following the recession of the lake from the 76 m shoreline.

The change was achieved by the 'Robinson' River becoming progressively less ice marginal as streams to the west became established. It can be shown from the current suspended sediment data (Table 4-4) that the less ice marginal the stream the less sediment transfer for similar flow rates. Hence both changes of volume and load are indicated from less melt from the ice cap along the 'Robinson' River. Following the recession from the 76 m level no marked change in process intensity is indicated by the results.

3) Stream temperatures: As the thermal regime of Generator Lake is important in evaluating the behaviour of the polar ice damming the lake, and as the average summer condition of Generator Lake is substantially ice-covered, thus minimizing solar heating, the thermal contribution of stream water was examined. These measurements were pursued following some casual and surprisingly high temperature readings (4°C) at the shallow lake margin waters only a few

metres from the main lake ice.

During all subsequent visits to the stream monitoring sites temperatures were taken at the bank and during current metering, at intervals right across each stream. This latter procedure on the 'Robinson' River revealed consistently higher temperatures on the west side than on the east side, ranging up to 1°C-2°C. The difference was traced upstream to the confluence of the two main tributaries just above the 76 m shoreline and was not attributable to east-side marginal snowbanks as was first thought. From these initial measurements additional temperature data were gathered on other streams including monitoring a tiny stream at camp on a twelve hourly basis. This stream became small enough to be crossed by a single stride following snowmelt, and the stream and air temperature data are included on Table 4-6. Other stream temperatures are listed in Table 4-3. From these data it is apparent that a) stream temperatures are sensitive to air temperature but inversely with size - the smallest streams are most responsive to temperature fluctuation. b) Some surprisingly warm water enters Generator Lake. For example, the 'Whalerock' River when flowing at 105 cfs had a temperature of 14°C on July 29, 1969. c) The south and east side streams are generally warmer but smaller than the north and west side streams during the post-snowmelt period. d) Some of the heat introduced by the streams is lost out of the Clyde River overflow channel, as the mean temperature of outflow

TABLE 4-6

AIR, STREAM AND LAKE MARGIN TEMPERATURES GENERATOR LAKE, 1969 (°F)

Date	Air	Stream	Lake Margin
July 5	38		32
6	37		32
6	42		34
7	39		33
7	40		34
8	38.2		32.5
8	50		33.5
9	42.8		33.5
9	43		33.5
10	37		33.5
10	36.5		34
11	35.7		34
11	39.5		34.5
12	38.3		33.5
12	45		35.5
13	41.7		34
13	46.3	41	36
14	41.5	40	35.5
14	54	46.5	38
15	44.7	41	39.5
15	42.5	42	38
16	38	40.5	37
16	39.5	40.5	38
17	36.5	39	37.5
17	45.2	43	39
18	40.4	39	37
18	38.5	46	36.2
19	40.1	40.5	36.5
19	42.5	46	38
20	39.2	42	35.5
20	40.2	42	36
21	41.1	40	38
21	39.9	46	41
22	37.6	40	39
22	47.6	51	47
23	39.2	40.5	38.5
23	53	50.5	42
24	50.5	48	40
24	56	55	48
25	47.3	48.5	43.5
25	52.1	56	45
26	43.4	48	44.5
26	50	57	51
27	53	50	42
27	55	59	46.5
28	46.9	50	46
28	61.2	61.5	56
29	48.9	50.5	48
29	52.3	56	53.5
30	49.3	49.5	49.5
30	52	54	52.5
31	48.1	49.5	46.5
31	55	55.5	45.5
Aug. 1	45.7	47.5	43.5
1	49	52	49.5
2	47.8	47.5	47
2	49.4	51.5	51.5
3	48.8	47.5	47.5
3	53	53	50
4	49.1	48	49
4	50.9	51	50
5	49.5	46.5	48
5	48.3	52	50
6	47.8	47	49
6	52.3	51	50.5
7	46.6	47	50
7	49.6	53	52
8	49.1	49	50
8	57.7	57.5	51
9	50.6	50	48.5
9	59	59	51.5
10	49.2	50.5	47.5
10	55.2	56	52.5
11	46.8	46.5	50
11	55.5	56	55.5
12	46.5	47	50
12	50.9	55	54
13	45	47	50
13	49	48.5	50
14	47	45.5	48.5
14	47.8	46	50
15	45.8	43.5	49
15	48	46	49.5
16	40.6	40	45.5
16	43	42	46
17	38.7	36.5	42
17	42.3	43.5	45.5
18	40	38	43.5
18	48.2	46.5	46
19	39.8	39	43.5
19	45.4	46	46.5
20	36.7	36.5	43.5
20	50	48.5	49
21	42.9	39	44
21	48.3	46.5	44
22	34.3	36.5	41
22	44.6	44	45
23	37.8	36	42
23	43	44.5	45
24	38.1	37	42
24	42.7	42	44
25	39.6	38	42
25	42.3	42.5	44
26	36	34	42
26	43.7	44.5	44.5
27	37.2	36	40
27	40.9	43	43.5
28	34.9	36.5	40.5
28	37.9	39	41
29	36.5	35	39
29	32.1	34	40
30	31.8	32	40
30	33.5	35.5	40
31	31.7	31.5	38
31	32.7	34	40

was just over 2.5°C on August 8, 1969 despite the persistence of lake ice.

In the case of the 'Robinson' River one further investigation of process was made to evaluate the response of stream water reaching Generator Lake often at 4°C or more. These temperatures were achieved flowing over a distance of 7 km from the ice cap margin. The temperature of the supraglacial stream was confirmed to be 0°C where it leaves the ice and begins crossing the debris-laden ice zone. The flow-travel time of the water from ice margin to the lake was established by timing the flow of Safranin dye plumes. This was done in six phases to ensure easy recognition of the dye plume at the check points. All measurements were achieved in the space of a few hours and the composite elapsed time was 2 hr. 16 min. for a 7 km flow. This was during a cool cloudy period and so the increases in temperature of 0.36°C/km and 1.1°C/hr. are attributed primarily to friction (Morisawa, 1968, p. 36). In this case irregular geometry was exerting influence to warm the water entering the partially ice covered lake.

From these few measurements some inferences may be made about the processes operating. Thermal input into Generator Lake is asymmetrical in degree although perhaps not in amount, with warmer water arriving from south and east side streams but more water from north side streams. Some of this heat will contribute to melting the lake ice cover. Some of the water at or about 4°C at maximum density should sink to the lake bottom against the glacial ice, where it would cause melting of the margin. However, none of the lake-bottom bathythermograph or reversing thermometer readings taken in 1968 (Barnett, Forbes & Whytock, 1970) reached values in excess of 2°C. Table 2-3 indicates that in July 1968 the regional mean air temperature was 6°F (3.3°C) below normal and in July 1969 it was 3°F (1.7°C) above normal. The 9°F (5°C) difference between the two seasons was marked as the minimum ice cover in 1968 was estimated at in excess of 60% in contrast with complete disappearance in August 1969. Although the lake tends to an isothermal condition (when it becomes a closed system in winter), as shown on the bathythermograms in Barnett, Forbes & Whytock (1970, Section III), the processes promoting this condition are not sufficiently rapid to prevent some energy loss as surface flow of warmer water leaving by the Clyde River.

The geomorphic significance of these thermal processes are much more tangible when considering shoreline processes in the next major section.

b) Lacustrine processes and responses:

i) General: Generator Lake is the dominant hydrological feature of the drainage basin. It is a response by both size and shape to the ice margin and the basin geometry. It is fed by fluvial processes which have been shown to be asymmetric in volume, sediment input and amount of thermal input. It is not suggested as a tenable hypthesis that all drainage basins are symmetrical in such inputs; obviously they need not be. However, the degree of asymmetry of input here is such that differences of responses are demonstrably of significant proportions in this particular environment.

The occasions when Generator Lake is totally ice free are few and brief. It was so in late August 1966, 1969 and 1971. The exact duration of lacustrine conditions is unknown but in all probability two to three weeks a year would be a maximum. In contrast, in 1965 the marginal lead was almost non-existent well into August, and in 1967, 1968 and 1970 a substantial ice cover persisted throughout the season.

The winter ice cover of the lake reaches almost two metres with snow cover up to one metre thick spread unevenly across it. Holdsworth demonstrated, by drilling the ice and levelling the water surface, that the winter water level reaches a constant minimum value (Barnett & Holdsworth, 1974, p. 394). Examination of the cross-profile of the Clyde River outlet (Figure 4-16) and the relevant stage of lake level (Figure 4-18) on the same day (August 11, 1969) indicates that the minimum lake level is the threshold of the overflow channel which demonstrates that outflow must cease in winter and that the lake margin must freeze to the bottom at that location. As these conditions persist for at least six months of each year and in all probability for as long as eight months, it is not surprising to have found a near isothermal condition prevailing in June and July of 1968 with warmer (1^{o}-$2^{o}C$) and denser water at the bottom of

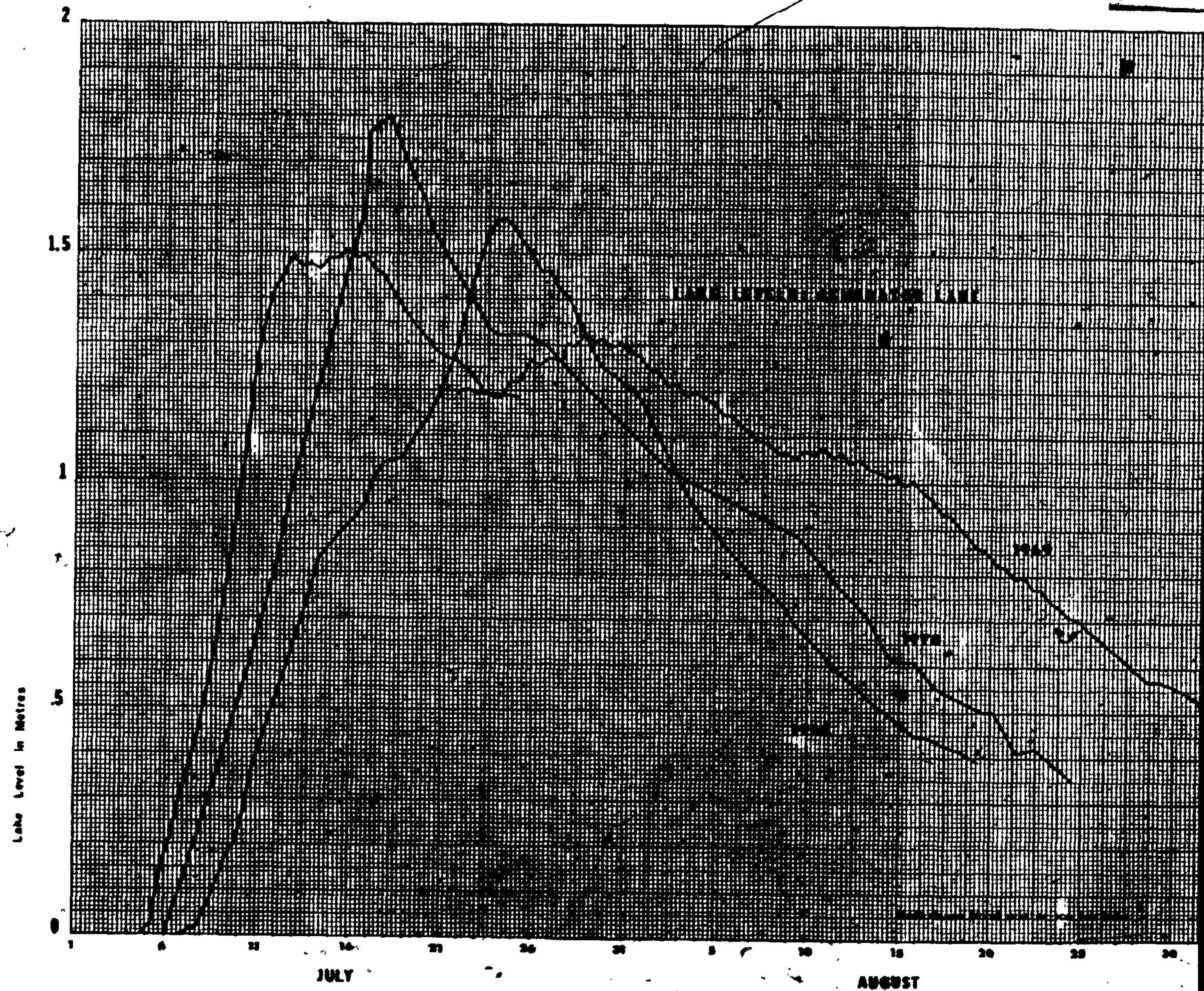

Figure 4-18: Lake levels 1968-1970, Generator Lake.

the lake. Such a low-energy environment lacks dynamic processes and the responses are equally low key - a light silt and clay cover on the basin bottom.

There are two circumstances when the lake environment becomes more dynamic - when the spring influx of meltwater thaws the lake ice margin and allows it to float freely and the whole of the lake ice melts allowing substantial wave action to develop across the lake. In the case of the former, water pours into the lake and raises its level quickly and substantially. It was not until 1970 that the geomorphic significance of this process was fully appreciated. The rise caused appreciable stress on the ramped portion of the ice front by exerting a bending moment. This was graphically illustrated to Holdsworth when he observed and heard a major calving of an ice ramp just prior to the flood crest in 1970. The failure response was attributed to the stress induced by a particularly rapid rise in lake level during the preceding two days, July 15 and 16, and culminating in failure on July 17. In such circumstances the increasing buoyant force was changing faster than the ice could adjust without fracture. Such a response is thought to have been a recurring one throughout the development of the lake basin and to have signalled a distinct discontinuity in the formation of a sublacustrine moraine. In the second case of dynamic lake processes wave action works on both the ice cliffs and the shoreline as well as introducing some turbulence into the normally placid waters of the central part of the lake. This latter case is best illustrated by reference to the shoreline environment in the next major section.

ii) Quantitative lacustrine data:

1) Lake levels: Lake level rise in late June to early July was monitored for four consecutive seasons, 1968-1971. In the last two years it was by Dr. G. Holdsworth. A graphical plot is presented in Figure 4-18. One common attribute of each season is a very rapid rise in lake level and a more modest rate of decrease following a generally sharp crest. The exception was the 1971 season which exhibited an apparently atypical early rise which then persisted at a low peak value for most of July. The peak was of the order 20% lower than the other three years of record although water volume (area under the curve) was generally more than in the other years (Barnett & Holdsworth, 1974, Figure 12). The response to such a process would be to exert a longer-term, lower-intensity stress on the ramped section of the ice cliff leading to accommodation by the ice rather than fracture. Persistence of any ramp forms would result. In this instance the response was primarily attributable to meteorological conditions and their temporal variation over the snowmelt period.

Closer examination of the data on water in storage up to August 5 in Generator Lake, set out in Table 4-7, shows that 1968 was a low volume year in average daily storage, but the relative position of the high volume years 1971

TABLE 4-7

RUN-OFF WATER IN STORAGE GENERATOR LAKE (42 km^2)

Year	Peak Storage (metres)	Days To Aug. 31	Mean Daily Storage (cm)	Area Under Run-Off Curve (cm^2)**	Mean Daily Storage To Aug. 5 (days)	Mean Daily Storage To Aug. 5 (cm)	Area Under Run-Off Curve to Aug. 5 (cm^2)**	Mean Daily Storage As Basin Run-Off (540 km^2) (cm)
1968	1.58	58	67.7*	157*	29	101.7	118	5.27
1969	1.51	60	99.2	238	31	121.0	150	7.72
1970	1.80	59	81.4*	192*	30	114.2	137	6.33
1971	1.26	63	90.9*	229*	37	115.5	171	7.07

** 1 cm^2 - 4.2×10^6 m^3 (42 km^2 x 0.1 m)

* includes estimating linear decline in water level for final 12 days in 1968
7 days in 1970
25 days in 1971

and 1969 change depending upon criteria used in calculations. For example, using the area under the curve (Figure 4-18 and Table 4-7) 1971 is ahead of 1969 up to August 5 by a ratio of 171:150 indicating greater aggregate storage during snowmelt, but when mean daily storage is calculated 1969 is ahead 48.4 cm to 46.2 cm, which is attributable to the exceptionally sunny August of 1969 when marked ablation of the ice margin took place.

The responses to the process of rapid rise of lake level are therefore: a) development of a shorelead, b) buoyant stress on any existing ice ramps, c) if fracture occurs, an underwater projection or ram on the new ice cliff (Barnett & Holdsworth, 1974, Figure 18), d) an iceberg or bergs of considerable magnitude free to float in the lake to the extent of open water; e.g. 0.4-0.5 x $10^6 m^3$ in 1970 (Barnett & Holdsworth, 1974, p. 399), e) large free floating icebergs would then have potential for creating local ice pushed ridges at the lake shoreline, and/or impede lake drainage by jamming the Clyde River outlet.

It has been possible to quantify the relationship between the lake level rise (Z_w) and the area ($\bar{A}$) of the lake thus: $Z_w . \bar{A} = C$; water level rise multiplied by area is equal to a constant or $Z_w = \frac{70}{\bar{A}}$ which is an idealized hyperbolic relationship (Barnett & Holdsworth, 1974, Figure 13). Therefore, the smaller the lake the more effective lacustrine processes operating are in destroying ice ramps due to rapid and large rises in lake level. This process diminishes the potential size of sublacustrine moraines by floating the source material (the ramp) and destroying the environment of deposition. Conversely the larger and wider the lake, the less stress is exerted on the ramps due to small and slower rises in water level and the more potential there is for both strong and longer sublacustrine moraine development beneath the ramps.

2) Overflow discharge: As inflow exceeds outflow in the early part of the snowmelt season lake level rises represent water in temporary storage above the Clyde River threshold. It is apparent from the figure of over five centimetres per day (Table 4-7) in each year for basin run-off in storage, that this volume is a response to the geometry of the outlet channel as it does not represent throughput of water. If it had been so, 305.7 cm of water (more than ten times the estimated annual precipitation) would have drained in 1968, the lowest run-off year. Reversing the calculation and assuming annual water-equivalent precipitation at between 12 and 25 cm the daily precipitation throughput in a 60 day melt season would be between 0.2 and 0.42 cm/day, thus indicating that basin run-off is retained between 12 and 26 days in the lake itself. This estimate is based on data from 1968 (5.27 ÷ 0.2 and 5.27 ÷ 0.42), a season in which annual snowmelt constituted the main hydrological event, with little appreciable melt of the ice cap.

Taking the mean flow rate through the Clyde River overflow for 1968 and 1969 (the two seasons measured personally) and the lowest and highest values recorded of the four available, another assessment of the quantitative values is possible. A mean flow taken over 60 days at the measured velocity on August 11, 1969 (0.9854 m/sec.) produces a throughput of water of $135.95 \times 10^6 m^3$ and $146.19 \times 10^6 m^3$ in 1968 and 1969. When divided by the basin area of 540 km^2 these figures produce 25 cm and 27 cm respectively. Although of an expectable order of magnitude the 27 cm figure must be considered low as 49.5 cm of water was still flowing out on August 31 and a linear projection of lake level decline predicted another 15 days of diminishing flow. In fact, assuming no further inflow and a continuing discharge at 1000 cfs it would take 8.4 days to drain the lake to the threshold. This minimal additional flow represents a further 3.8 cm of run-off from the basin, making a seasonal total of 30.8 cm at a minimum, an increase of 23% above 1968 levels.

3) Additional responses: The responses additional to those listed above of the inflow of water exceeding the capacity of the overflow channel to discharge are:

a) trapping of heat in the bottom of the lake basin which promotes melting of both glacial ice and lake ice,

b) trapping of sediment in the lake basin thus seriously diminishing the amount of sediment leaving the lake. A minimum influent sediment rate for the lake, derived from a mean value for each of the five streams measured, although not strictly numerically comparable, gives a minimum order of magnitude comparison with outflow: 848.6 g./sec. and 236.4 g./sec.; therefore, a minimum of 72% of the suspended sediment and all bedload is trapped in the basin (less than half the influent streams were monitored). c) Downcutting of the lake outlet. Abundant evidence of the erosive power of the overflow is present on the southfacing hillside above the present outlet. Exposed water-washed bedrock, largely free of debris, plus a breach in the Iron Formation outcrop approximately six metres deep occur just upstream of an accumulation of outwash just outside the lake basin above the Clyde River.

V <u>Shoreline processes and responses</u>

a) General: The shoreline is the locus of some of the most dynamic processes and most interesting responses in the lake basin, as it is the interface not only between the land and the water but between the glacial ice and the water and also land, water and ice. The movement of this process boundary, both gradually and more catastrophically when the lake drained, has offered a rare opportunity to document not only processes but to estimate their rates over the past 4500 years.

Four aspects will be examined in this section: ice pushed ridges, deltaic sedimentation, beach pebble morphology and ice contact deposits.

b) Ice pushed ridges:

i) General: The morphological features considered here were created by lake ice. They are discontinuous along the shoreline and are rarely more than three metres high. Therefore they have little in common with features bearing the same name discussed by Rutten (1960) and Scheidegger (1970, p. 47) which are much larger (50-200 m high) and longer (100 km) and are attributed to glacial ice advance over terrain with a permafrost condition. A third type of ice pushed ridge is that created by sea ice and appears to be intermediate in size between the two; this latter type was documented from Alaska by Hume & Schalk (1964). Dionne (1971) recorded estuarine ice push from the St. Lawrence and riverine ice push from the subarctic Grande Rivière in Quebec (Dionne, 1976).

The term ice pushed ridge (Nichols, 1953) describes both process and response and there is no doubt that lake ice, in this case, is the pushing agent, but two differing processes have been documented in the literature. The two are: thrusting due to rapid thermal expansion of an essentially complete ice cover (Hobbs, 1911; Zumberge & Wilson, 1953) and secondly, rafting of ice (Tyrrell, 1910) which requires a shorelead, probably the lake ice broken into two or more pans, a strong or persistent wind and a gently shelving shoreline. In both cases loose materials or easily disturbed bedrock in close proximity to the shoreline are also essential elements of the process-response system. However, all the above circumstances even in favourable combinations do not guarantee preservation of ice pushed ridges as other than seasonal features because wave action can subsequently destroy both types during the open water season. Therefore, the occurrence of fossil and perennial lake shore ridges requires the absence of sufficiently destructive processes to account for the lasting features. The two initiating processes are seasonally mutually exclusive but are less so spatially. However, at Generator Lake thermal thrust would have to take place at the winter low-water stage and thus any ridges so-formed would be vulnerable to any subsequent wave action at a higher lake stage.

In order to build lasting ice pushed ridges it is necessary to remove or restrict the destructional effects of summer waves from the landform. This separation of the destructive process from the earlier response may be achieved in several ways - by draining the lake, by insufficient lead development to allow significant wave formation, i.e. by a very short, cool summer season, by seasonal fluctuation of lake level such that the ice push is at or near the crest of the cycle and any wave generation is during the declining phase, by creation of ice pushed ridges only in high intensity-low frequency climatic conditions of wind and wave thus pushing them beyond the reach of high frequency low intensity processes. Obviously, several combinations of these circumstances could account for the occurrence of perennial ice pushed ridges.

Most references testify to incidental observations of ice pushed ridges often occasioned by striking responses created either during or ~~just~~ prior to visiting a locality, e.g. Peterson (1965). The study by Zumberge & Wilson (1953) was an exception and it clearly documented the thermal expansion thrust process in quantitative terms. No quantitative study of wind driven ice pushed ridges is known for the lacustrine environment, but that by Mansikkaniemi (1970) for the marine environment is detailed enough (and includes experimental work) to have applicable aspects for lake study. Recently an entire conference was devoted to the subject and the collected papers (some of which are cited separately) appear in No. 1-2 of Volume XXX of Revue de Géographie de Montréal.

ii) Ice push at Generator Lake: Ward (1959) recorded observations made in 1950, all of which were of ice floes thrusting ashore with '4 ft. thick ice' ending up resting 3.6 m above lake level, and 'large boulders and sediment' pushed three to six metres inland. Ward and Goldthwait discussed and considered thermal expansion as a possible explanation for existing 'ice ramparts' but concluded that no evidence for this process was apparent in May of 1950 (Ward, 1959). Only a single doubtful instance of thermal thrust, (locally deformed marginal ice prior to the melt) was observed by the writer along the north shore of the lake at the narrows, but no significant geomorphological work was apparent. Hobbs (1911) stressed the prerequisite condition of rapid transmission to the ice of changing temperature and concluded (page 158) that in Greenland slow transmission was the norm due, in large part, to snow cover. Therefore it is concluded that ice floe thrusting is the dominant process and that thermal thrust is a minor process at best and is not geomorphically significant in the lake basin.

During the summer of 1969 which was sufficiently warm to eventually remove all the lake ice by August 26, careful observation was made of localities where and when free-floating ice impinged on the shoreline. The most critical period was August 13 to 18 when a sizeable lead, obvious surface candling of the lake ice and strong winds combined to break up the ice into numerous pans and fragments. They first rafted gently ashore in the northeast but then impinged more vigorously on the southwest shore after a wind switch to northeast. Although wind circulation was the most vigorous of the summer during this period (Figure 4-19) it was a summer of gentle winds when no fewer than 10 of the twice daily one minute run-of-wind observations were calm. This was in strong contrast to the estimated 80-90 km/hr. winds of July 11, 1968 which had been strong enough to destroy a large tent, but achieved no ice push as the lead was insignificant in width at that time.

Hence, in 1969, while the amount of geomorphological work achieved by the lake ice was obvious, its magnitude was less than would occur with stronger wind circulation. Two categories of feature were formed - ephemeral ridges

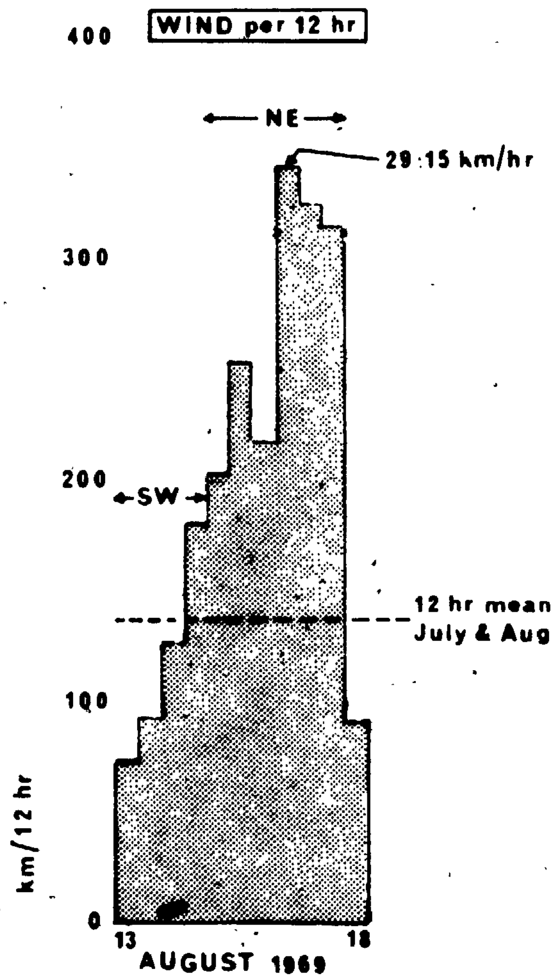

Figure 4-19: Run-of-wind August 13 to 18, 1969.

and grooves (Figures 4-20, -21) of up to 10 cm relief and secondly more lasting modification of the shoreline morphology (Figure 4-22). This latter is very similar to Ward's description of 'boulders and sediment pushed ashore' rather than the boulder ramparts characteristic of other parts of the shoreline (Figure 4-23). Having concluded that extensive open water is only a low-to-moderate frequency event under prevailing conditions, and having observed two seasons with high percentages of open water (1966 and 1969) but also having failed to observe substantial ice push, creating ridges with a probability of more than a season-or-so life expectancy, some conclusions may be drawn: a) ice pushed ridges are low frequency responses to high magnitude processes, b) the process of formation is wind-driven ice floes, c) climatic and geometric conditions suitable for formation occurred around both the 76 m lake and the present one, d) the size, location and length of the ice pushed ridges around the 76 m lake are such that long fetch is a significant geometric variable in their formation, e) considerable thrusting force is necessary to move the size of boulders observed to be components of the ridges.

iii) Quantitative data on ice pushed ridges: Data were gathered from 30 sites, 19 from the present shoreline and the remainder from fossil ridge sites. These data are set out in Table 4-8 and include for each site the mean dimensions of the ten largest boulders observed, the mean volume and the largest single dimension (a-axis),

TABLE 4-8

ICE PUSHED RIDGES DATA

Site No.	Shore (Old) (Present)	Height (ft.	/ m)	Fetch (km)	(degrees)	Orientation (degrees)	Slope (degrees)	Mean Size 10 Largest (cm) Max.'a' Axis (cm)
1	O	6	1.83	9.02	220	160-340	2.25	168x96x37.5 245
2	O	9	2.95	9.02	245	005-185	2.75	139x96.5x34 185
3	O	8	2.44	8.21	260	020-200	4.0	172x111.5x34.5 240
4	O	12	3.66	7.08	075	015-195	4.0	161x100x39 235
5	O	8	2.44	5.15	120	060-240	4.5	146.5x107.5x39.5 165
6	P	7	2.13	2.09	320	050-230	-	121.5x75.5x47 160
7	P	6	1.83	1.77	140	080-260	-	~~104x73x39~~ 120
8	O	7	2.13	4.03	270	030-210	4.0	193x119.5x39 250
9	O	10	3.05	9.66	240	180-360	2.75	153x111x34.5 195
10	O	11	3.35	10.29	230	170-350	3.5	206x136x54.5 390
11	O	10	3.05	4.64	290	050-230	5.0	149x78x27 220
12	P	6	1.83	0.58	240	180-360	-	92x54.5x36 150
13	P	7	2.13	1.59	275	035-215	-	97.5x71.5x43 130
13a	P	7	2.13	1.47	345	105-285	-	110x71x52 155
14	P	8	2.44	4.51	060	150-330	-	103x77.5x46.5 120
15	P	7	2.13	0.89	100	040-220	-	92x70.5x42.5 200
16	P		1.52	1.53	105	045-225	-	[illegible] 110
17	P	4	1.22	1.96	070	010-190	-	89.5x58x38.5 125
18	P	8	2.44	1.96	100	040-220	-	149x106.5x51.5 215
19	3 m*	6	1.83	1.77	095	035-215	-	96x53x34.5 145
20	P	9	2.95	4.03	095	035-215	-	122x81.5x50 165
21	P	7	2.13	3.60	190	130-310	-	111.5x63.5x44.5 145
22	P	9	2.95	6.00	200	140-320	-	121.5x78.5x44.5 150
23	P	10	3.05	[illegible]	220	160-340	-	119.5x73.5x37.5 180
24	P	7	2.13	3.30	255	025-205	-	121.5x74x44.5 170
25	P	5	1.52	6.19	020	140-320	-	85.5x64x33.5 105
26	P	4.5	1.37	2.02	325	085-165	-	86.5x55.5x30.5 160
27	P	8.5	2.59	5.42	225	165-345	-	98.5x70.5x47.5 130
28	P	4.5	1.37	4.05	135	075-255	-	67.5x53.5x33.5 105
29	O	5	1.52	6.56	190	120-300	-	133x76.5x42 180
Mean	O	8.6	2.6					162x103x38
	P	6.82	2.1					104x70x44

* 3 m above present lake

Figure 4-20: Ephemeral ice pushed ridges, south shore Generator Lake, August 1969.

Figure 4-21: Ephemeral grooves caused by ice push of large boulder on pebble beach, south shore Generator Lake, August 1969.

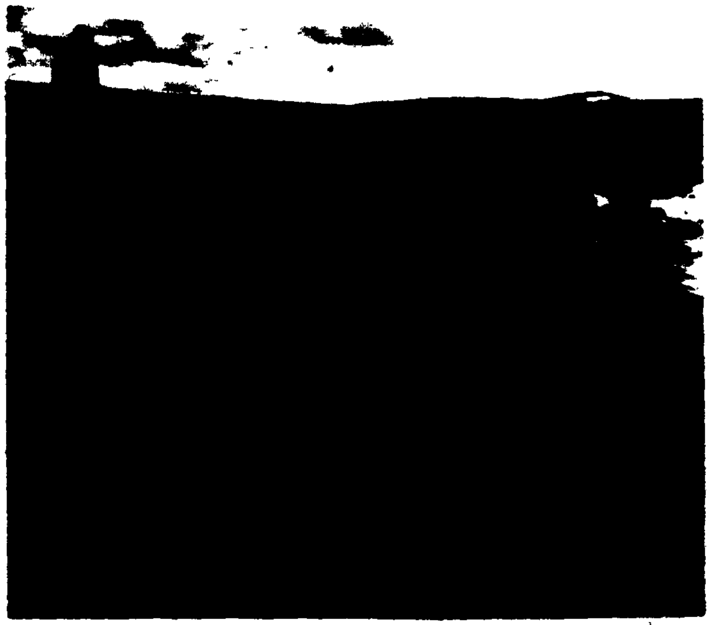

Figure 4-22: Ice push leading to shoreline modification. Note the unmodified shore in the background. South shore with till cover.

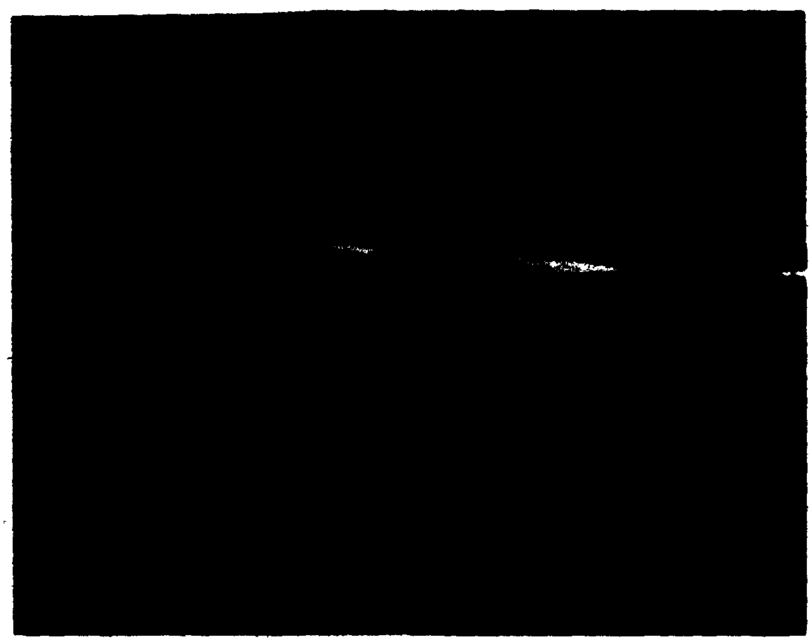

Figure 4-23: Ice pushed ridge in boulder rampart form. North side Generator Lake. Tape (5 cm across) for scale.

orientation of the ridge, height of ridge, fetch length and direction (calculated as maximum potential open water without regard to dominant wind direction) and, for the raised features the inclination of the slope below the ridge. The highest ridge is 12 ft. (3.66 m) high and occurs on the 76 m shoreline, as does the largest boulder (390 cm long) and the largest mean volume 3.627 m^3. The larger features are associated with the larger lake. An attempt was made to analyse the variables contributing to the formation of ice pushed ridges. Fetch appeared to be a variable of some considerable importance but no marked preferential orientation of ridges is apparent. A simple linear regression of fetch with height was calculated for the ridges around the present lake. The result was an insignificant correlation coefficient (r - 0.38; r^2 - 0.144); by deleting two of the values derived from very close to the present ice margin a much better least squares fit was achieved (r - 0.64; r^2 - 0.41) for the remaining 15 sites, but obviously fetch is either not the single dominant variable or it does not exhibit a linear relationship with ridge height. This uncertainty is confirmed when the data from the 76 m shoreline are subjected to the same analysis because another insignificant correlation results.

Examining the distribution of height values a strong modal value of 7 ft. (2.1 m) occurs based on 19 present shoreline measurements with a mean value of 6.82 ft. (2.1 m). The distribution of values for the 76 m shoreline shows two weak modes at 8 and 10 ft. (2.4 m and 3.0 m) based on 10 measurements, with a mean value of 8.6 ft. (2.6 m). This analysis leads to the conclusion that the bigger the lake the higher the ridges. The continuity and length of ice pushed ridges, although not measured, was considerably greater on the 76 m shoreline then on the present shoreline, and is estimated to be of the order of two to three times the length (Figure 4-24).

With this indication of the larger ridges being related to the larger lake a further analysis of the variables thought to be significant in the formation of ice pushed ridges was by use of a stepwise multiple regression. The independent variables considered were slope, fetch, total length a+b+c axis means, mean of 'b' axis and maximum 'a' axis; the last three values were from the ten largest boulders measured at any one site. The dependent variable was height. The program continued operating at low significance levels ranking variables by explanation of variance (Table 4-9). Combinations of variables yielded a degree of explanation of 49-63% but probably the most significant result was that the ranking varied between 'b' mean and fetch and slope angle (for the 76 m shoreline ridges), but the latter remained close to the top of the list. From these analyses it appears obvious that at least one additional significant variable is or was operating.

From the data gathered and inferences from climatic records it seems unquestionable that ice pushed ridges of

TABLE 4-9

STEPWISE MULTIPLE LINEAR REGRESSION EXAMPLES

76 M SHORELINE

Site No.	Height Dep.Var.1	Fetch Ind.Var.2	Slope Ind.Var.3	T abc Ind.Var.4	'b' Mean Ind.Var.5	Abs a Ind.Var.6
1	6.0	5.6	2.25	301.5	96.0	245
2	9.0	5.6	2.75	269.5	96.5	185
3	8.0	5.1	4.00	318.0	111.5	240
4	12.0	4.4	4.00	300.0	100.0	235
5	8.0	3.2	4.50	293.5	107.5	165
8	7.0	2.5	4.00	351.5	119.5	280
9	10.0	6.0	2.75	298.5	111.0	195
10	11.0	6.4	3.50	396.5	136.0	390
11	10.0	2.9	5.00	254.0	78.0	220
Mean/(N-9)	9.0	4.62	3.64	309.22	106.22	239.44
Standard Deviation	1.94	1.454	0.90	42.74	16.39	66.35

Step No.	Variable Entered	Multiple R	RSQ	Increase In RSQ	F Value To Enter	Number Of Independent Variables Included
1	3	.2504	.0627	.0627	.4682	1
2	2	.6832	.4667	.4040	4.5456	2
3	4	.6937	.4812	.0144	.1392	3
4	6	.7033	.4947	.0135	.1069	4
5	5	.7488	.5607	.0661	.4512	5

Degree of explanation achieved 56% with slope and fetch contributing more than 40% of explanation of the variation in ridge height.

PRESENT LAKE LEVEL

Site No.	Height Dep.Var.1	Fetch Ind.Var.2	T abc Ind.Var.3	'b' Mean Ind.Var.4	Abs a Ind.Var.5
6	7.0	1.30	244.0	75.5	180
7	6.0	1.10	216.0	73.0	120
12	6.0	0.36	182.5	54.5	150
13	7.0	0.99	212.0	71.5	130
13a	7.0	0.91	233.0	71.0	155
14	8.0	2.80	227.0	77.5	120
15	7.0	0.55	205.0	70.5	200
16	5.0	0.95	194.5	63.5	110
17	4.0	1.22	186.0	58.0	125
18	8.0	1.22	307.0	106.5	215
20	9.0	1.26	253.5	81.5	165
21	7.0	2.25	219.5	63.5	145
22	9.0	3.79	244.5	78.5	150
23	10.0	3.63	230.5	73.5	180
24	7.0	2.05	240.0	74.0	170
25	5.0	3.87	183.0	64.0	105
26	4.5	1.26	172.5	55.5	160
27	8.5	3.39	216.5	70.5	130
28	4.5	2.53	154.5	53.5	105
Mean/(N-19)	6.82	1.86	222.18	70.32	148.16
Standard Deviation	1.70	1.15	41.60	12.13	31.72

Step No.	Variable Entered		Multiple R	RSQ	Increase In RSQ	F Value To Enter	Number Of Independent Variables Included
1	'b' mean	4	.6657	.4432	.4432	13.5305	1
2	fetch	2	.7356	.5456	.1024	3.6062	2
3	Abs a	5	.7855	.6171	.0715	2.7998	3
4	T abc	3	.7950	.6320	.0150	.5697	4

Degree of explanation achieved 63% with the 'b' mean contributing 44% of the explanation of variation in ridge height, with fetch adding a further 10%.

Figure 4-24: Long ice pushed ridge on south shore at the 76 m level.

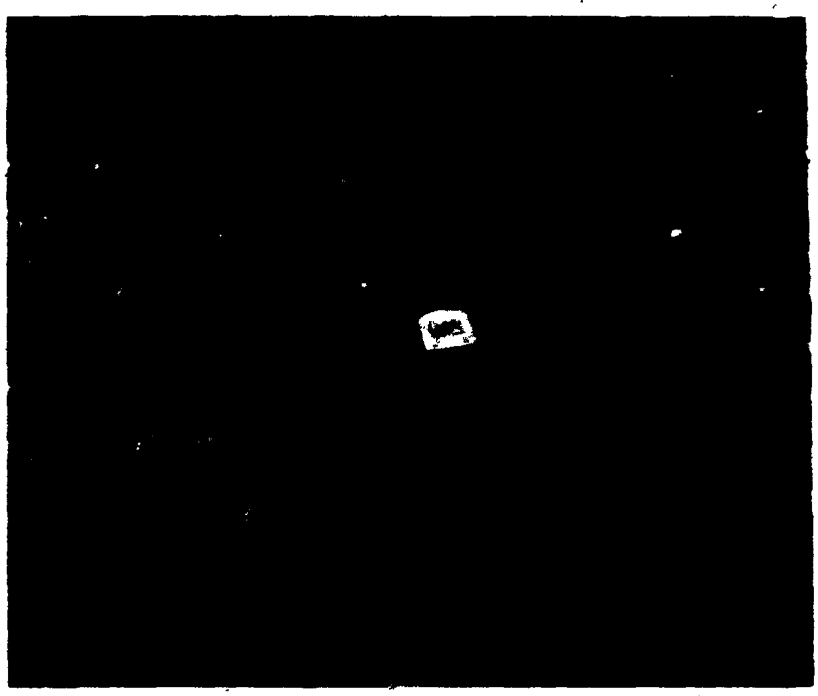

Figure 4-25: Pebble beach showing some rounding of pebbles.

substantial form are the result of high intensity, low frequency processes. Under such conditions it seems highly probable that wind speed was the variable which it is not possible to include in the analysis. Although 'fetch' is a partial surrogate for wind speed, the critical minutes of ice push could well come from a direction considerably off the line of maximum fetch.

Mansikkaniemi (1970) examined ice push on a seashore in southeastern Finland both quantitatively and experimentally. Some of his results are applied to the Generator Lake basin to derive a minimum force required to move blocks of the magnitude measured. Mansikkaniemi (1970) used a naturally flat exposure of rapakivi (igneous) rock, calculated the volume and weight of the test stones and measured the force required to drag them across the surface by dynamometer. Three different sides of each rock were used as the dragging surface. He found that the volume could be adequately approximated by the formula $v = 4/3\ \pi\ r^3$ where $r = \frac{a+b+c}{6}$ and that the average force required was 80% of the weight of the stone for an even and dry rapakivi surface. Additionally, he noted that the degree of immersion in water facilitated thrust as did a slimy surface for small stones. In the laboratory it was shown that well rounded cobbles could be moved at 50-60% thrust (expressed as percentage of weight of block) and 'very unrounded' blocks required 100% and more thrust (Mansikkaniemi, 1970, p. 22). In the field blocks weighing 1300 kg were easily moved but one weighing

7800 kg was not (page 26). At the time of writing Mansikkaniemi (1970, p. 25) had blocks 5000-15 000 kg in sight (in the field) to test for what he considered may be the upper limit of thrust in that marine environment.

At Generator Lake historical conditions of major thrusting are interpreted as having been on moderately smooth igneous rock having an upward slope of between 2-5°, being wet but not slimy and with little actual immersion of the blocks forming the ice pushed ridges. Differences between the Finnish situation and Generator Lake are that in the latter case additional thrust would be required to overcome the slope, whereas the wet surface would ease the thrust requirements. However, the interaction of the various blocks comprising the ridge would be an additional deterrent to thrusting. This condition was not specifically addressed by Mansikkaniemi in the experimental case but he recorded 'block walls' and 'block pavements' as common responses to ice thrusting. Given these constraints it is proposed to present data using the 80% weight approximation for thrust as a best estimate but with a greater probability of the 100% value being valid rather than values below 80%.

The thrust values and site numbers are set out in Table 4-10. It will be seen that the largest block (site 10) required, according to formula, a thrust of 12 321 kg but the next largest (site 8) only 4490 kg. These values exceed those measured in southeastern Finland by a considerable amount but clearly Mansikkaniemi (1970, p. 25) deliberately

TABLE 4-10

ICE PUSHED RIDGES:

BLOCK SIZE AND CALCULATED MINIMUM THRUST VALUES

Site No.	Shore (O/P)	Height (ft.)	Mean Axis Length a b c (cm)	Largest Block (cm)	Volume (m³) Largest	Volume (m³) Mean	Weight (kg) Largest	Weight (kg) Mean	Minimum Thrust (80 % wt, kg) Largest	Minimum Thrust (80 % wt, kg) Mean
1	O	6	164x96x37.5	240x140x[illegible]	1.888	0.532	5098	1436	4073	1150
2	O	9	139x96.5x34	160x150x60	0.982	0.380	2651	1026	2121	821
3	O	8	172x111.5x34.5	240x145x40	1.489	0.624	4020	1685	3216	1348
4	O	12	161x100x39	215x160x55	1.542	0.524	4163	1415	3330	1132
5	O	8	146.5x107.5x39.5	160x145x25	0.697	0.490	1882	1323	1506	1054
6	P	7	121.5x75.5x47	150x110x50	0.762	0.282	2057	761	1646	609
7	P	6	104x73x39	115x115x40	0.382	0.195	1031	527	825	422
8	O	7	193x119.5x39	280x155x40	2.079	0.842	5613	2273	4490	1818
9	O	10	153x111x34.5	165x155x25	0.796	0.516	2149	1393	1719	1114
10	O	11	206x136x54.5	390x200x75	5.704	1.209	15401	3264	12321	2611
11	O	10	149x75x27	220x105x40	0.943	0.318	2546	859	2037	687
12	P	6	92x54.5x36	150x60x55	0.361	0.118	975	319	780	255
13	P	7	97.5x71.5x43	130x90x55	0.403	0.185	1088	500	870	400
13a	P	7	110x71x52	155x100x75	0.697	0.245	1882	662	1506	530
14	P	8	103x77.5x46.5	115x95x60	0.341	0.227	921	613	737	490
15	P	7	92x70.5x42.5	200x125x75	1.241	0.167	3351	451	2681	361
16	P	5	90.5x63.5x40.5	95x95x60	0.303	0.143	818	386	654	309
17	P	4	89.5x53x33.5	125x95x75	0.493	0.124	1345	335	1076	268
18	P	8	149x106.5x51.5	215x135x75	1.489	0.561	4020	1515	3216	1212
19	3 m*	6	96x53x34.5	145x45x45	0.252	0.120	680	324	544	259
20	P	9	122x31.5x50	165x120x70	0.868	0.316	2344	853	1875	682
21	P	7	111.5x63.5x44.5	145x85x60	0.473	0.205	1277	554	1022	443
22	P	9	121.5x78.5x44.5	125x105x50	0.426	0.283	1150	764	920	611
23	P	10	119.5x73.5x37.5	180x135x50	0.943	0.238	2547	643	2038	514
24	P	7	121.5x74x44.5	165x85x70	0.636	0.268	1717	724	1374	579
25	P	5	85.5x64x33.5	105x105x40	0.303	0.119	818	321	654	257
26	P	4.5	86.5x55.5x30.5	160x75x40	0.403	0.100	1088	270	870	216
27	P	8.5	98.5x70.5x47.5	130x120x60	0.697	0.197	1882	532	1506	426
28	P	4.5	67.5x53.5x33.5	105x65x45	0.193	0.072	521	194	417	155
29	O	5	133x76.5x42	170x75x50	0.498	0.309	1345	834	1076	667

* 3 m above present lake

chose the smaller blocks for his experimental work but anticipated thrusts up to 15 000 kg based on field observation of large blocks. Several of the thrust values required for the movement of blocks on the present lake shoreline are below 1000 kg (as low as 417 kg) which are in the same range (by size and required thrust) as those measured on the Gulf of Finland.

As concluded in the hydrological section, the optimal probability period for ice push is well into the declining phase of lake level and so, even if a block which is being pushed begins the process partially submerged, and therefore is more easily moved, it must be quickly thrust up above lake level, thereby rapidly losing any buoyant advantage. Blocks rest up to three metres above the probable lake level at the time of formation. Therefore, the thrust values derived may not be readily diminished by invoking major assistance from the buoyancy effect. In fact, just such an advantage has to be considered from daily changes of 30-50 cm (Mansikkaniemi, 1970, p. 24) when assessing the Gulf of Finland example.

From the data one may conclude that both the current and former thrust values at Generator Lake are not an order of magnitude less and in fact, may be approximately equivalent to those currently experienced in the Gulf of Finland despite the considerably different size of water bodies involved. Longer fetch and potentially larger masses of ice have not led to demonstrably greater thrust in the Gulf of

Finland. Such similar thrust values support the conclusion that ice pushed ridges are low frequency, high magnitude events at Generator Lake.

Consideration of the shoreline geometry and the morphology of the longest and largest ice pushed ridges preclude any possibility of the high value thrusts being attributed to icebergs calved from the ice margin as any such bergs with sufficient freeboard to gain major thrust from the wind would ground well offshore and would tend to form arcuate rather than linear responses.

In addition the data suggest that any close-to-shore man-made structures planned for either an ice-bound small lake or seashore with a two to five degree slope should be capable of withstanding thrusts of at least the magnitude inferred from the above data. In fact, it appears probable that the values derived should be considered minimum values only.

The geometry of the basin, the shore slope and fetch coupled with the geometry of the boulders play significant roles in the formation of ice pushed ridges. However, given the geometry on which the process must operate, the dynamics of the climatic variable are critical, as it is necessary to have strong winds during a period of substantial open water with the lake ice not only free to move but of sufficient internal strength to do the work and not just disintegrate into thousands of candles on impact with the shore.

iv) Conclusion: The formation of ice pushed ridges is a good example of the intimate relationship between processes and geometry giving rise to a distinctive response. In terms of the qualitative model ice pushed ridges are not a unique response attributable to the polar or subpolar proglacial environment. Their distinctive quality is their good degree of development and preservation which is a reflection of their environment. The low frequency of the constructive thrusting processes must at least equal or exceed the frequency of destructive wave processes, in addition to which the tempo and effectiveness of the wave processes must be much lower than for ice push. Other aspects being equal, it is the local climate, strongly influenced by the ice cap, which controls the effectiveness of the processes and hence the degree of development of the response.

The predictive qualities derived from the model are limited but are as follows: ice pushed ridges are most likely to occur on shore slopes of less than five degrees carrying loose boulders or loosely jointed bedrock. Their size will be greater the longer the fetch and the larger the size of boulders available up to an unknown limiting value of at least 390 cm. Winds in excess of 30 km/hr. and probably considerably in excess of that figure (perhaps 80 km/hr.?) are required in conjunction with free floating ice pans of sufficient integrity to drive ashore without disintegrating into a shower of candles.

c) Beach pebble morphology:

i) General: Pebble morphology and the process of rounding of shoreline pebbles was investigated. The basic assumption was that all beach pebbles began with minimal evidence of rounding by glacial or other processes and consisted of reasonably uniform granite-gneiss at the period of initiation of the lake and therefore the simple hypothesis was that the degree of rounding of a pebble increased with time and could be measured. Thus the time span indicated by the chronology would yield a rate or rates for rounding of pebbles for the two shorelines.

It was recognized also that variation in rounding was liable to be more complex than a simple linear relationship with time. Other variables considered included the initial size of the pebbles, fetch, shore slope and nature of the bedrock.

The dominant process in pebble rounding was assumed to be attrition due to wave action. As control, pebbles from adjacent terrain above the beach (and below in the case of the 76 m beach) were measured for comparison.

ii) Quantitative data on pebble morphology: As indicated in the methods section of Chapter 1, the method used for measuring pebbles was that advocated by Cailleux (Blenk, 1960) which involves measuring the 'a' axis ('b' and 'c' axes also measured) and the minimum radius of curvature in the principal plane. The formula used is R (roundness) = $\frac{2r}{a} \times 1000$ where r is the minimum radius of curvature in the principal plane and 'a' is the length of the 'a' axis. This radius of curvature was measured by visual approximation using a transparent acrylic sheet with arcs finely scribed on it so the rock particle could be viewed through it. This enabled a 'best fit' for the curvature of the pebbles.

Sites (Figure 4-2) were chosen by the pragmatic factor of the presence of large numbers of beach pebbles (Figure 4-25) as opposed to stretches of shoreline where cobbles and boulders were common. Fifty pebbles were collected at each site by random process (the collector reached behind himself and collected by touch but moving constantly to prevent gathering a cluster of pebbles). Pebble rejections were limited to a very few which showed recent fracture which would have obviously biased the results in favour of minimal rounding. Rejections by size were without measurement but were of cobble and boulder size rocks. The samples were labelled and returned to camp where measurements were conducted in the relative comfort of the tent thus avoiding measuring with cold and/or wet hands. Some samples were measured by two observers to check on operator variance. In all cases a complete 50 pebble sample was measured by only one observer. The 'a', 'b' and 'c' axis data which were identified visually, were measured with little absolute variation and excellent relative variation between the two observers. However, there was less agreement on the more subjective value of roundness based on the minimum radius of curvature in the principal plane in terms of absolute values. This was despite the improvement of the technique (the arcs on transparent sheet) but one observer produced somewhat higher values for each sample set remeasured (3). There appear to be appropriate comments on this result: a) the materials to be measured ranged from an "absence of rounding" through to a relatively low degree of rounding (Figure 4-25), each of which demands a greater degree of subjectivity than would measurements in the high range of roundness, b) the second, and obviously related, aspect is that the two observers may not have measured the same 'corner' of the rock in some of the more angular samples.

These reservations about the quality of the data refer only to the roundness measure and so the data on length, sphericity, and flatness are not subject to the same concern. Nevertheless despite the degree of uncertainty, evaluation of the data will be undertaken with the mean of the two measured roundness values accepted for the three pairs of test samples.

To put the values in perspective, a perfectly round pebble would have an R-value of 1000 and yet the highest value recorded was 291; the lowest value possible is somewhat more arbitrary for a 50 mm 'a' axis and a radius curvature of 1 mm, the R-value would be 40 but the R-value would increase to 80 if the 'a' axis was only 25 mm with the same (minimum) radius of curvature. There is a simple exponential relationship between length 'a' and roundness R with the form $R = \frac{2000\ r}{a}$ when or if 'r' is taken as constant. However, a brief examination of the variation of roundness with length was run on the non-beach samples which yielded a linear relationship of the form L = 138.5 - 0.45 R (Figure 4-26A) with a correlation coefficient of -0.897 ($r^2 = 0.805$) thus suggesting that for that environment and general pebble size, that given the length, the approximate roundness of non-beach pebbles could be predicted. This, admittedly, is a predictive capacity of distinctly limited utility. It could be used where degree of rounding by wave action was suspected to be marginally detectable. In such a case the predicted value could serve as a general threshold value above which rounding could be measured.

Bluck (1969) examined the relationship between length and roundness of six lithologies and found the highest degrees of roundness shown by beach pebbles was in the 64-128 mm size range. In passing it should be noted that one of the earliest investigators of pebble morphology, Wadel (1932), advocated using a standard pebble length of 70 mm. The overall mean length of the 39 sample means from Generator Lake is 69.57 mm, a fortuitously close agreement. Twenty of the 39 sample means lie within the 64-128 mm range, one just above and 18 lie just below 64 mm (smallest mean was 44.4 mm). Therefore the samples were generally in a size range shown by some studies to be most favourable to detecting the process of rounding.

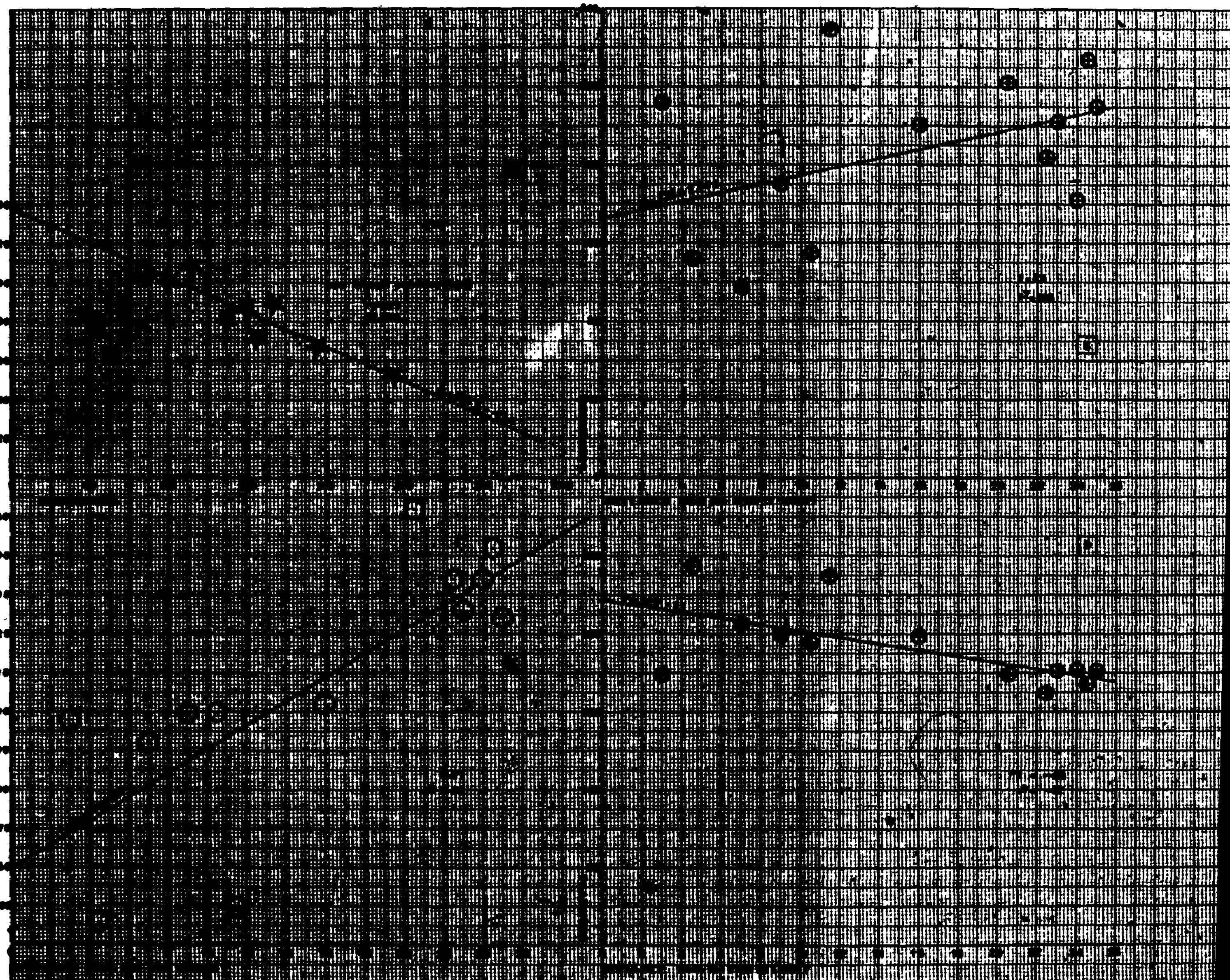

Figure 4-26: Pebble morphology relationships and correlation coefficients.

Formulae used: roundness (R) = $\frac{2r}{a} \times 1000$

sphericity (S) = $\sqrt[3]{bc/a^2}$

flatness (F) = $a+b \times 100/2c$

Rounding was detectable as shown in Table 4-11 and Figure 4-25 which indicates that the mean roundness of all beach sample means exceeded that of all non-beach sample means by an R-value of 61.

The hypothesis that pebble rounding at the present lake level is a simple function of time is testable by assuming that the retreat of the ice margin was approximately regular. There is little evidence counter to this assumption even after the major event of the lake draining and reestablishment was recognized: this event had the effect of removing additional beach processes, for an unknown duration but up to several hundred years, while the lake drained. Then the processes were reestablished simultaneously along the shoreline for the period following the surge of the ice cap. Thus the beach material has been subjected to wave action of decreasing duration towards the ice cliffs with a superimposed relatively brief period (ca. 50 years) of potential wave action of equal duration throughout the present basin.

A simple linear regression (least squares fit) of pebble roundness on distance from the ice cliffs (Figure 4-26B) was adopted and found to be very weak in explaining the roundness of the 13 samples measured. The regression $R = 133.72 + 2.137\ D$ produced a correlation coefficient of only 0.43 ($r^2 = 0.185$). However, when the same samples are tested for variation of sphericity with distance (Figure 4-26C) the regression $S = 0.660 + 0.003\ D$ produces a

TABLE 4-11

PEBBLE ROUNDING: BEACH VS. NON-BEACH SAMPLES

Location	No. of Sample Sites	No. of Pebbles	R Sum	R Mean	R Attributable To Wave action
Beach	31	1550	5817	187.6	61
Non-Beach	8	400	1013	126.6	-

correlation coefficient of 0.77 ($r^2 = 0.598$). Furthermore, the data on variation of flatness with distance (Figure 4-26D) produce a negative correlation coefficient of -0.65 ($r^2 = 0.42$) from a regression equation of $F = 221.98 - 2.034\ D$. One may conclude from these data that sphericity is the most sensitive of the three measures, if the approximation of a linear relationship is correct, and secondly that flatness is also a much more sensitive measure than roundness. Nevertheless rounding is demonstrable to vary with time but obviously not in a strongly linear relationship. Pebble geometry is demonstrably modified by beach processes despite the relatively minute percentage of the year that the environment is active. No matter how dimly this result is perceived from the quantitative data, it is an important one for it demonstrates that in this strongly periglacial environment a few weeks annually (at most) of lacustrine processes leave their mark. This leads to the conclusion that up to 45 weeks of freezing temperatures, with doubtless several spring and fall freeze-thaw cycles in the micro-environments between pebbles and cobbles, do not lead to rapid fracture and splitting of pebbles, thus masking or obliterating the modifying effects of beach processes.

In addition, variation in rounding with fetch was examined and again a simple linear relationship was assumed with the hypothesis that the longer the fetch the greater the roundness expected. The resulting regression had the form $R = 194.25 - 3.494$ Fe (correlation coefficient $r = 0.345$; $r^2 = 0.119$). Before drawing any conclusions, variation in both sphericity and flatness with fetch was examined with the following results: $S = 0.752 - 0.006$ Fe (correlation coefficient $r = 0.752$; $r^2 = 0.565$), $F = 161.05 + 3.749$ Fe (correlation coefficient $r = 0.589$; $r^2 = 0.347$). Again sphericity appears to be the most senitive measure of the three with a similar degree of statistical explanation achieved by a linear relationship. Flatness again offered the better fit of points to the line than did roundness, but in this case the formerly negative relationship for flatness with distance has become positive, suggesting that the relationship with fetch, at least, is spurious.

The role of fetch in the 76 m basin is difficult to isolate due to the open shape of that lake. As the lake became larger, the maximum fetch became greater for almost all sample sites from the very youngest to the oldest. Taking fetch as the dominant variable and examining the variation in roundness, sphericity and flatness of the sample pebbles and assuming a linear relationship no meaningful correlation resulted. The highest value was for roundness which, however, only explained statistically 38.6% of the measured variation ($r^2 = 0.386$).

Taking sphericity as the most successful measure of linear variation with both distance from the ice margin (time) and with fetch, it was tested for linear variation with values derived from the product of distance and fetch. The result was not statistically significant.

iv) Conclusion: The relative sensitivity of the three measures with distance only remains the same for the two shoreline environments; sphericity, flatness and then roundness a long way back. However, only sphericity maintains an acceptable level of explanation for each environment (Blalock, 1960, p. 299, indicates 0.7 for 'r' as a general lower limit), presupposing a generally linear relationship. Each measure indicats a weaker relationship on the 76 m shoreline than for the present. It is doubtful whether firm quantitative significance can be attached to such data but the trend indicated is definitely in harmony with what one would expect for an environment where the processes changed (about 2500 years ago) from those modifying shape toward sphericity and roundness, to periglacial processes which lead to disintegration into angular forms. Not only does this suggest a trend but it also offers some insight into the tempo of periglacial processes which modify pebble shape - they are not as rapid as the lacustrine processes which modify pebble shape. Without a chronological framework such insights would not be sustainable.

d) Proglacial lake terrace:

i) General: This term was introduced in Barnett (1967, p. 178) to describe a substantial linear accumulation of bedded sands found just below (2-4 m), and associated with, the 76 m shoreline largely on the south side of

Generator Lake. Somewhat lesser but similar accumulations on the north side of the lake were also recognized, and all are readily visible on air photographs as discontinuous light sandy patches immediately below the strandline.

The size, abrupt boundaries, and the distribution of the poorly-bedded sands demand that the source of the material was the ice margin where a vigorous stream entered the former lake. The bedding generally dips in a direction bisecting the angle between the line of the ice cliffs and the shoreline. The top surface is planar but not quite horizontal (however, the shoreline is horizontal, Barnett, 1967, p. 185) declining slightly in elevation away from the ice. The top surface also carries a few scattered boulders and it is deeply dissected by snowmelt gullies which in most localities have cut right through the sand to the underlying till surface. The largest unit of this terrace is sufficient, without any modification, to land an Otter aircraft fitted with oversize tires.

ii) Stratigraphic relationship with sublacustrine moraines: The stratigraphic relationship was established between the sands and sublacustrine moraines at two locations on the south side of the lake (Barnett, 1967, p. 180) (Figure 4-27) which clearly indicated that the sands overlie till. This distinct relationship is just as would be expected, in keeping with the current state of knowledge about the process of formation of the moraines, but was less certainly expectable at that time when the Andrews &

Figure 4-27: Poorly sorted sands of the proglacial lake terrace overlying the till of a sublacustrine moraine.

(GSC Photo 203028 A)

Smithson (1966) theory of moraine formation at the ice front would allow for simultaneous deposition of both glaciofluvial sands and till in close proximity to the ice margin. The process of moraine formation outlined in Barnett & Holdsworth (1974) and summarized earlier in this chapter identified the locus of till deposition as beneath the tapered ice ramp. In this case the glaciofluvial ice-marginal drainage would deposit its load just beyond the till ridge which, when later exposed, would receive a capping of sands distinct from the underlying till.

A similar morphological setting occurs in the present lake (Figure 4-28) where the ice margin, the lake and the shore environments meet. However, in this case the geometry is changed by the presence of the surge moraine as a subaerial ridge which submerges gradually as its alignment swings from east-west to north-south across the basin. At present ice marginal drainage is for the most part flowing on the proximal side of the surge moraine but a parallel stream flows on the distal side of the moraine. Both are flowing in environments rich in loose, sand-size or smaller sediment and both are building deltas of which the one on the proximal side is certainly an ice-contact feature. As the ice-margin retreats (Figure 4-28) rapid deposition by aggradation creates an elongated ice marginal terrace of the general form of that observed near the 76 m level. The processes operating are, therefore, concluded to be rapid sedimentation, generally in a narrow deltaic form due to

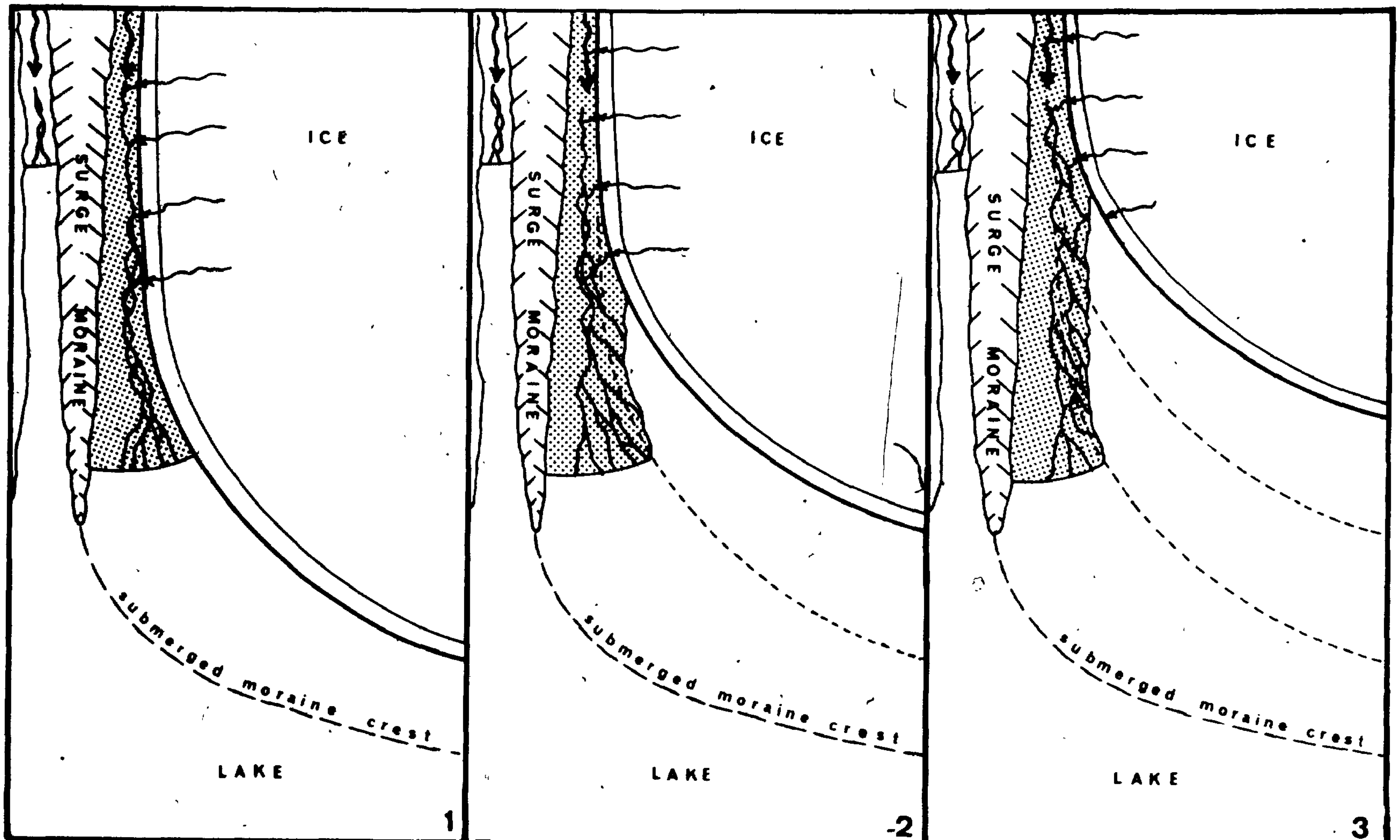

Figure 4-28: Three phase evolution of proglacial lake terrace on the proximal side of the surge moraine.

the retreating locus of the apex of the delta resulting in a narrow elongate terrace form.

iii) Quantitative data on the proglacial lake terrace: Six short vertical channel samples of sand were collected from the raised terrace from locations 100 m apart. In five of these localities a channel sample from beach material immediately above (topographically) was collected for comparative purposes. The intent of this sampling was to examine whether the degree of sorting of the two sand deposits was different in order to show a different depositonal history or whether perhaps the terrace sand was all part of an eroded beach. In addition a sample from the present beach and three more from the active surface of the proglacial terrace forming in the present lake were collected. Both Trask Sorting and Logarithmic coefficients were calculated (Krumbein & Pettijohn, 1938, p. 232) and yielded a strong similarity of the arithmetic measure for sorting of the beach samples (old and new). Equally encouraging was the degree of variability which characterized the sorting of the samples from the two terrace environments (Table 4-12). The collective differences are shown graphically in Figure 4-29 which shows, by shading, the consistently poorer degree of sorting of the terrace samples.

Ives (1962, p. 198) reported 'kame deltas' from an abandoned proglacial lake shoreline at the northwest margin of the Barnes Ice Cap. These features appear to be similar

TABLE 4-12

Sorting coefficients of grain size analyses

Morphological form	Laboratory number	Trask sorting coefficient	Logarithmic coefficient
"Proglacial lake terrace"	GBL 2001	1.95	0.290
*"Proglacial lake terrace"	GBL 2002	2.92	0.465
*"Proglacial lake terrace"	GBL 2007	1.78	0.250
*"Proglacial lake terrace"	GBL 2009	2.13	0.328
*"Proglacial lake terrace"	GBL 2011	1.87	0.272
*"Proglacial lake terrace"	GBL 2013	6.48	0.812
			Mean 0.403
*Abandoned beach (south shore)	GBL 2003	1.52	0.182
*Abandoned beach (south shore)	GBL 2006	1.58	0.199
*Abandoned beach (south shore)	GBL 2010	1.38	0.140
*Abandoned beach (south shore)	GBL 2012	1.81	0.256
*Abandoned beach (south shore)	GBL 2014	1.69	0.228
			Mean 0.201
Present beach (south shore)	GBL 2005	1.54	0.188
Kame delta (present lake)	GBL 2023	3.87	0.588
Kame delta (present lake)	GBL 2024	1.97	0.294
Kame delta (present lake)	GBL 2025	1.52	0.182
			Mean 0.355

Samples marked * are paired, *i.e.*, beach sample taken directly above terrace sample at approximately 100-metre intervals from west to east.

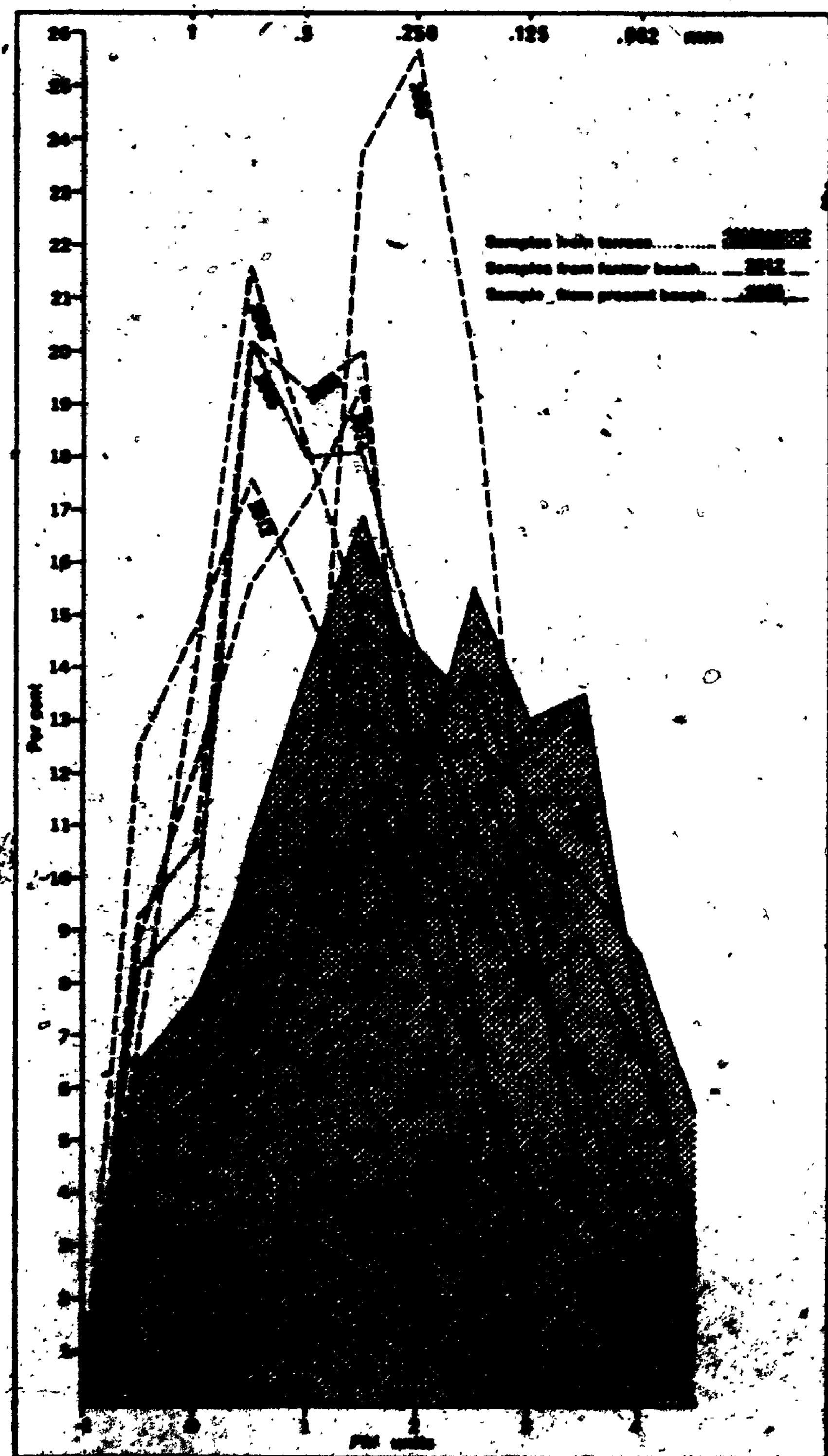

Figure 4-29: Grain size distribution showing contrast between beach and terrace (shaded) samples. Particles finer than 44 microns, have been omitted. The maximum emission occurs in No. 2013 where it amounts to 7.8% of the sample. (After Barnett, 1967).

in origin to the Generator Lake terrace although perhaps more deltaic in morphology. If such is the case, then similar processes have produced features of different form highlighted by the terms 'delta' and 'terrace'. The controlling variable on morphology is, therefore, the rate of retreat of the ice margin.

iv) Conclusion: The occurrence of a proglacial lake terrace in Generator Lake is not unique but once again, as in other aspects, it represents a development which is close to one end of a spectrum of responses. The conditions are such that the cold ice ensures a high percentage of surface meltwater run-off which finds a readily available load at the shear zone of the ice margin. This meltwater with often a considerable sediment load, deposits the material at the ice margin in a slightly offshore location. This location is a reflection of the locus of initial breakup of the lake ice where a lead develops between a narrow landfast ice foot and the main lake ice sheet. The good preservation of the relatively poorly sorted, deltaic bedded landform is a function of the low frequency of significant waves which would, in a more open lake, have sorted the material and moved much more of it into the beach zone with a consequent blurring of the morphological definition of the proglacial lake terrace. Hence the cold ice and the local climate have led to the occurrence of a well-developed landform which would be either poorly developed or reworked in a more dynamic lacustrine environment.

e) Deltaic sedimentation:

i) General: Sandy deltas are being built into the present lake basin wherever streams enter. The asymmetry of both inflow and load, already demonstrated for the 76 m shoreline environment and discussed in the present hydrological regime, is visible in the magnitude of the present deltas. The special case of the ice-contact delta forming the proglacial lake terrace discussed above is one of the more vigorously growing. Larger deltas occur on the north side of the lake than on the south. Deltaic processes operating are unusual only in their strong seasonality. The deltas are subject to seasonal fluctuation of the lake level which changes the locus of the active processes. Initial deposition is in an aggradational environment as lake level rises strongly and is then replaced by degradation in the upper part of the delta as lake level recedes. Thus, a delta is built both outward and upward for the first three weeks to a month of each season and then is progressively exposed and incised by the stream(s) as lake levels drop. Prograding takes place in part at the expense of the fresh materials deposited earlier near the apex of the delta.

ii) Quantitative data on deltas: As the present deltas are under lake ice and then under water at least for part of the season, limited effort was devoted to their examination. In order to obtain some idea of the geomorphic response to the rapid seasonal change in lake level, profiles were run from the apex of two deltas (located at DP1 and DP2 on Figure 4-2) outward into the lake as far as wading safely would permit. The resulting profiles are shown in Figure 4-30. For comparative purposes a profile was run across a raised delta at the 76 m level (located at DP3 on Figure 4-2) and plotted on the same figure. It is apparent that the micro-relief of the raised delta top surface is greater than that of either of the present ones. That in itself is entirely expectable as the present ones are subject to annual rejuvenation whereas the raised one is subject to erosion by meltwater, nivation and to some small degree at least, by deflation. However, the range of relief of almost seven metres, which could be considered the dissected topset of the raised delta, is in contrast with one metre or so relief of the modern deltas. Recalling the model of the behaviour of the lake, where water level rise is a function of area, then the prediction from that relationship alone would suggest a greater range of elevation for the modern topset slope than for the former delta topset.

One obvious possible explanation is that the profiles for the modern lake are incomplete and that considerably more relief would have been measured before reaching the delta foreslopes if the profiles could have been extended safely. The opinion formed at the time was that the beginning of the foreslope was in fact reached and also, as the length of all the profiles is similar, it suggests that

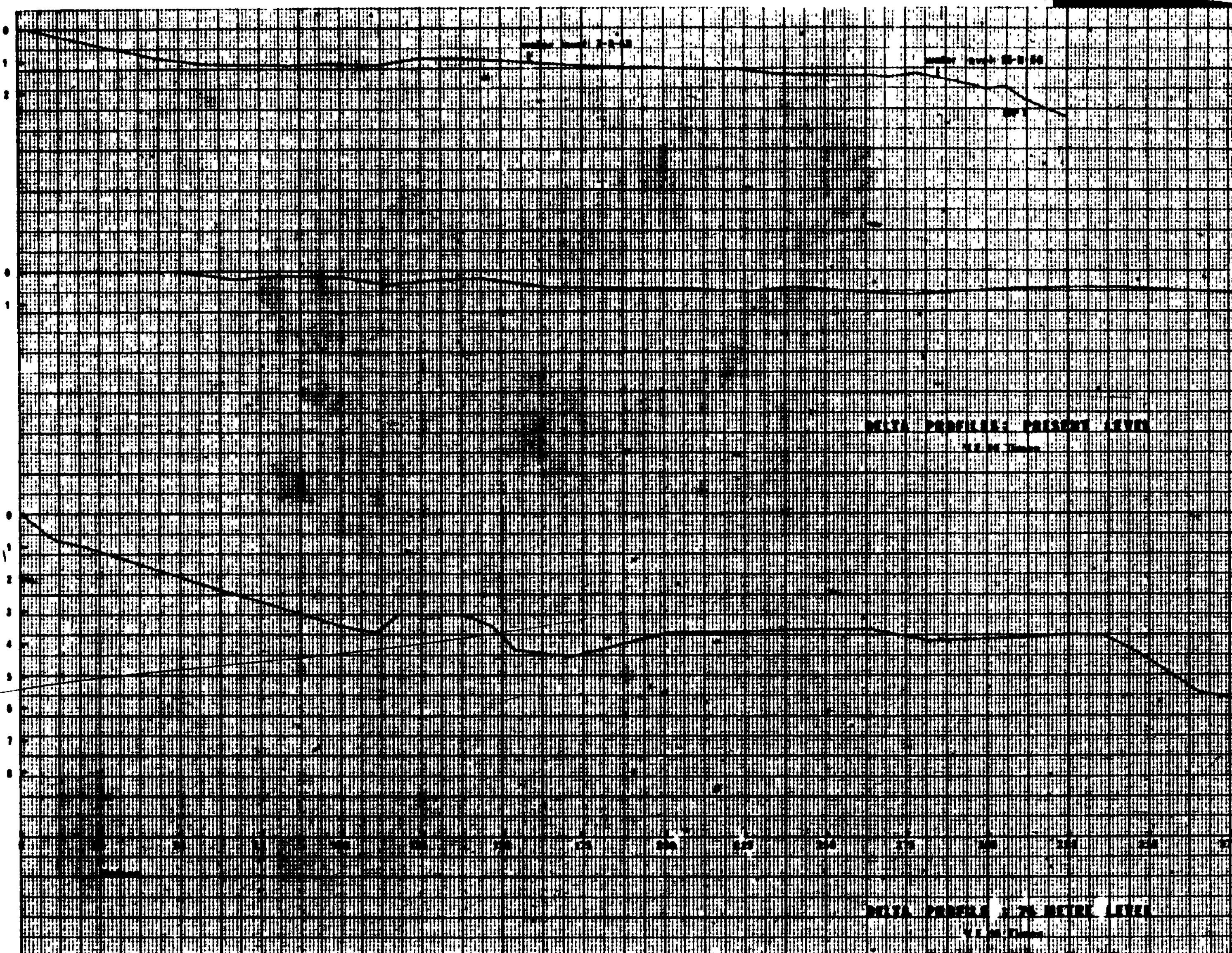

Figure 4-30

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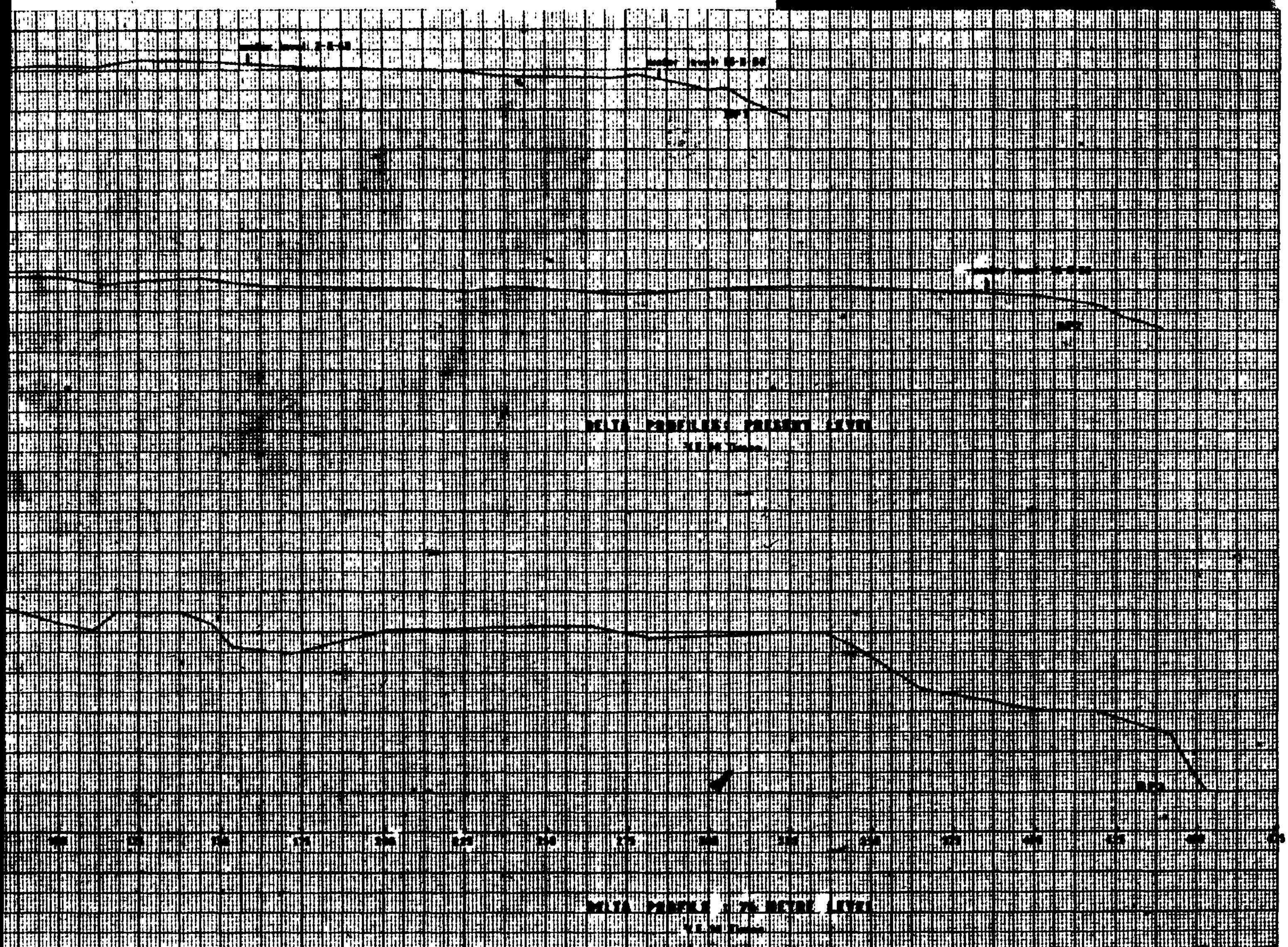

Figure 4-30

that is not an unreasonable supposition. Additional evidence for the smaller relief range in which shoreline processes are operating at present lake level is provided by profiles (Figure 4-31) run across the modern shoreline at four localities (shown on Figure 4-2 as SP1, SP2, SP3 and SP4) on the north side of the lake. Each shows a range of elevation of the shorezone of the order of two metres. A comparative profile for the 76 m shoreline is shown in Figure 4-32. It must be recognized that for every elevation between 76 m and the present shore lake level must have passed through at least once, and as all lake level adjustments must have been summer season events, then continuous deltaic deposition must have occurred at all levels along the 'Robinson' River, for example. This circumstance of ephemeral changes is not in itself adequate to explain the relief of the 76 m delta, but may represent a partial explanation for the morphology of the lower facet between 350 m and 440 m across the profile (Figure 4-30) which could represent a topset surface at a slightly lower elevation than 76 m.

iii) Conclusion: Single deltas in Generator Lake are not remarkably different from those in any other lake in an extreme climate, but collectively the flights of deltas have considerable importance. Together they represent a distinct record of key events which has been very useful for unravelling some of the complexities of the evolution of the lake basin. They indicate stability of lake level

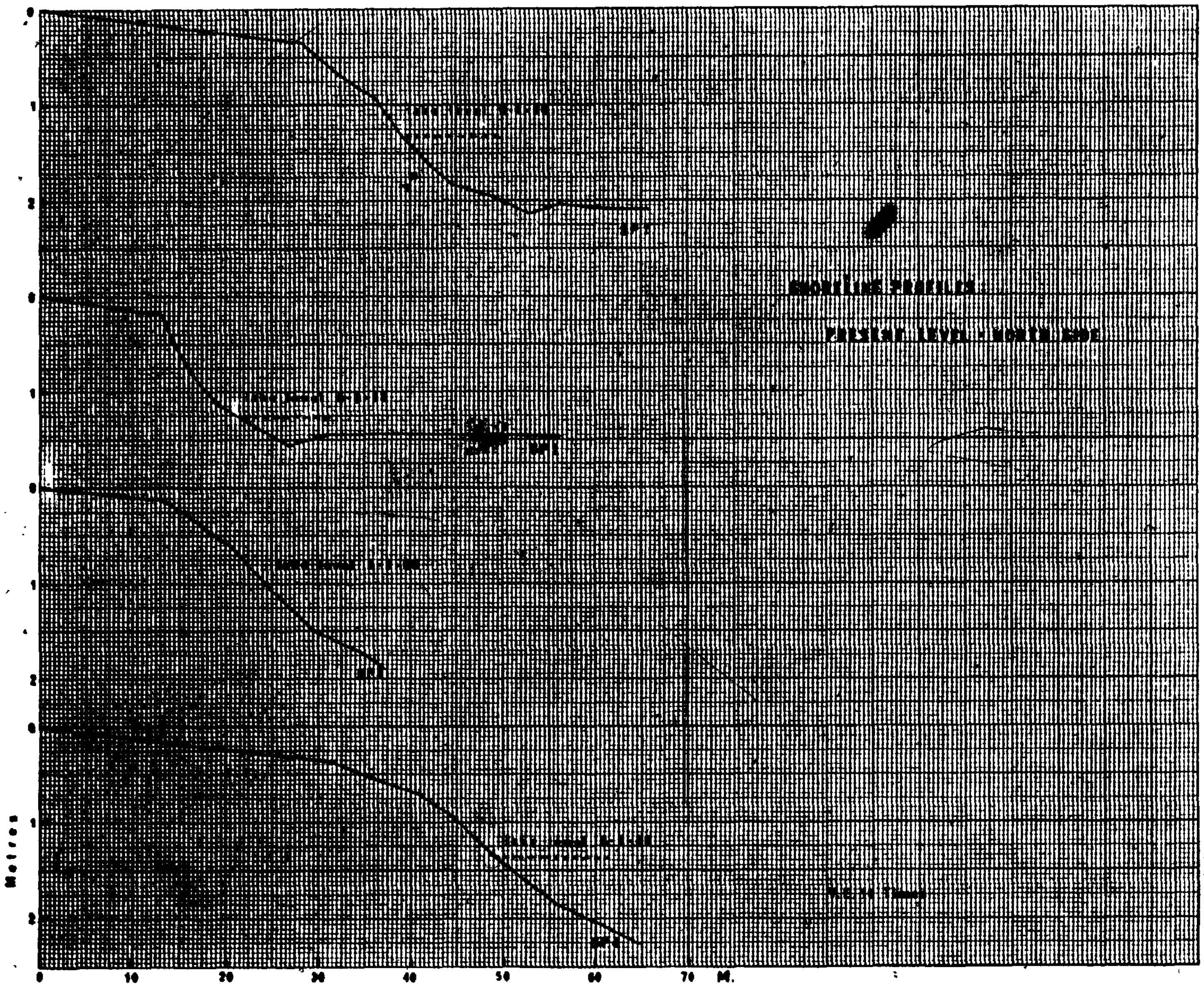

Figure 4-31

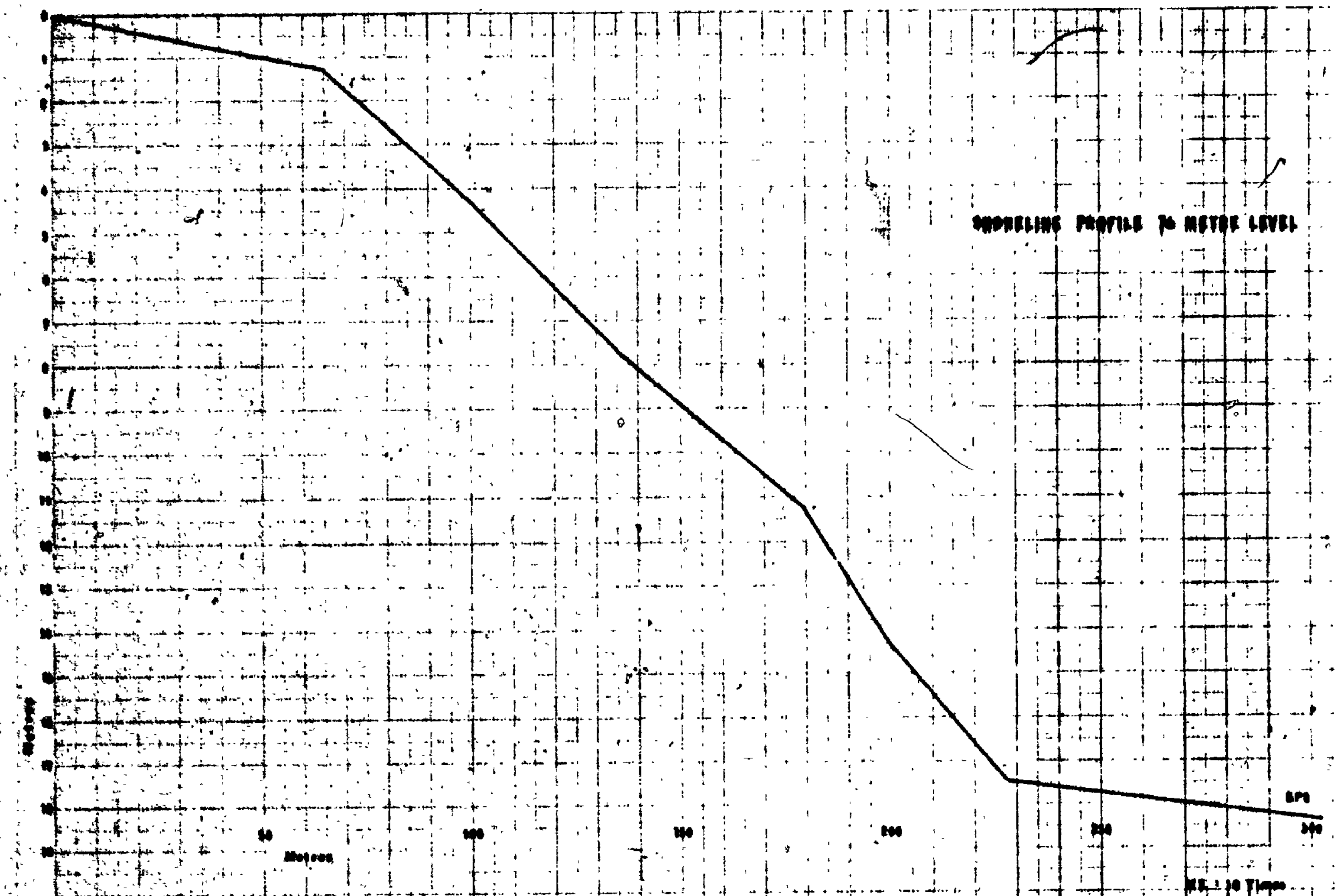

Figure 4-32

when glaciologically that was unanticipated; they represent a source of datable material which is significant when lichenometry was shown to be of debatable absolute value; they provide a sedimentological record of process and process intensity over the period of lake level adjustment, as well as a check on the coherence of the chronology derived from them; their very existence and size is due to the presence of the ice cap, the polar nature of which promoted surface run-off rather than en- or subglacial drainage which in turn, if operating, would have diminished the size of the deltas.

VI Other subaerial processes

In order to complete consideration of the range of processes operating this brief section is included to cover aeolian action, sheetflow, nivation, freeze-thaw and shallow slope failures. Some activity in all these processes was observed but in each case the responses were minor in magnitude and not particularly distinctive to the environment.

Wind is responsible for snow distribution and keeps moraine crests free of snow. The consequent low moisture contents may result in diminished effectiveness of frost action, a factor influencing till fabric as a means of studying primary processes. Fine grained mineral material is also moved about by wind and this is easily detectable as a light dusting on the surface of the Barnes Ice Cap, but has no other perceptible influence in the basin.

Sheetflow is generally visible in some low slope areas during snowmelt and is definitely potentially active if heavy rain falls, but the observed movement in response to this process is minor in both degree and duration.

The effects of nivation were observed as hollows cut in the steep margins of raised deltas but only amount to minor etching of the surface even after periods in excess of 2000 years.

Freeze-thaw action is demonstrable as high centre polygonal ground is visible between the ice cap and the northern shoreline of the 76 m lake. The geometry of boulders that are split is attributed to the freeze-thaw process but it will be recalled that beach rounding has diminished little in the 2500 years that this process has had to work on the partially rounded pebbles.

Shallow slope failures were noted in three localities and although obviously not widespread features, nor large in dimension (the largest was estimated at 75 m^2), they may be attributed to processes which concentrated the fine fraction of sediment, in this instance local patches of relatively boulder-free till. Failure was then dependent upon slope, a shallow frost table and saturated or supersaturated till.

Thus it is concluded that these processes, in sum, are operating at a tempo which does not markedly affect the morphology of the lake basin nor contribute even subtle etching of the responses in a manner which can be considered diagnostic of the proglacial lake environment.

CHAPTER 5

ASSESSING THE MODEL

Assessment of the model of the proglacial environment of Generator Lake involves examination of five aspects or themes, the last one of which will be developed in this chapter. The other four have been running through the earlier chapters. The five aspects are:

1) development of internal coherence and order in the available data, that is the establishment of a logical framework for the presentation of the information,

2) development of both a spatial dimension for and interaction of the processes and responses within the basic geomorphic unit of the drainage basin. A case is made for a distinctive suite of landforms, which individually are not unique to the type of environment but which collectively have the stamp of a sub-polar ice proglacial lake basin,

3) development of quantitative relationships, which allow prediction of future developments. To varying degrees such predictions are deemed possible,

4) demonstration of the potential utility of the qualitative model and the quantitative relationships as a basis for study of other sub-polar proglacial environments,

5) use of the model and the relationships established to facilitate better understanding of paleogeomorphic

relationships in other glaciated areas of Canada, particularly on the Shield. This last assessment is conceptually important, as too often those who seek to make paleogeomorphic interpretations have spent (for obvious reasons of accessibility and cost) either little or no time examining the processes operating at the margin of an ice cap.

In the course of development of the model a number of hypotheses were developed or in some cases merely implied. As a first step in assessment, the conclusions may be stated as follows, with discussion of some conclusions in succeeding sections:

a) a sub-polar ice mass with a cold-based margin produces distinctive landforms (sublacustrine moraines, proglacial lake terrace, a glacial ramp form with thermal undercutting) which are well developed in, but are not individually exclusive to that environment,

b) the landforms produced in a sub-polar proglacial lake basin represent a distinctive suite of landforms (equivalent in concept to a suite of fossils characterizing a bedrock formation) and as such can be considered collectively diagnostic of such an environment. They include anchor ice sediment rafts, ice pushed ridges and shoreline striae,

c) sublacustrine moraines ($\simeq$ cross-valley moraines) are not annual features but take from 5 to 23 years to form with a mean rate of 10 ± 1 years,

d) geometry as a fundamental attribute of the nature and location of geomorphic features has been previously underemphasized. A response or landform is not predictable from a given process without consideration of the properties of the surface acted upon. The regional slope will determine whether ice will form a lake or not. Ice pans drifting ashore will have variable geomorphic effectiveness depending upon shore slope,

e) tempo or rate of process activity should be considered in paleogeomorphic interpretations both as a diagnostic aid to process change and also by reversing the procedure and using assumed tempo to check apparent chronology,

f) lichenometry is a viable relative chronological control, but it does not provide reliable 'absolute' chronological control,

g) single radiocarbon dates, no matter how carefully obtained, should not be considered completely acceptable unless other substantiation is obtained,

h) a series of carefully established radiocarbon dates with one or two 'maverick' dates may be deemed acceptable when other substantial geomorphic and stratigraphic evidence supports the chronology proposed on the basis of that group of dates,

i) increasing recognition of a sub-polar category of ice sheet requires careful evaluation of evidence for such glacial conditions in paleogeomorphic interpretations of the former ice cover of Canada.

The model as a logical framework

The qualitative model takes a drainage basin - a logical and well-defined subdivision of the landscape - as its basic unit. The model is depicted diagrammatically (Figure 4-1) as a circle, representing the watershed or perimeter of the unit. The Generator Lake drainage basin differs in two important respects from a non-glacial basin. It has a significant portion of glacial ice contributing to the hydrology and as such the basin has a relatively poorly defined watershed across the ice surface. Furthermore, the basin differs from some glaciated basins in that the ice margin is partly frozen to the bed and lies across the regional slope thus impounding water and causing it to rise upslope and overflow across the low point in the watershed, however, 7.8% of the basin area lies below the level of hydrological outflow.

Having thus defined the geographical limits of the model the components of process, geometry and response were established in that order because a given process can produce a variety of responses depending upon the geometry of the setting. This is fundamental to the whole thesis as the Barnes Ice Cap if sitting atop a topographic high and all other major variables remaining constant would not have impounded run-off nor produced most of the features documented.

As is appropriate to a model, not all processes and responses observed were included in the model, for example, fluvial processes such as stream braiding or meandering were considered only implicitly and responses such as patterned ground considered not significant in the context of testing a hypothesis for distinctiveness for the proglacial environment.

The process segment of the diagram was set up to allow for interaction between processes, with few which could be shown to be more effective and sensitive than the fluvial process on the ice cap side of the lake in response to increased temperatures. In addition to such interactions, shown in the diagram by radial linkages, each process has a seasonal variation, some of which are extreme. Perhaps none is more so than wave action which may effectively be absent in some of the coldest summers, but with the right combination of wind and water it may achieve considerable response in a very brief time span. These seasonal variations have been considered within the concept of tempo, although this term has also included consideration of variation of rate of activity between years and over longer time spans.

The explicit recognition of geometry within the model has compelled consideration of its role in each response recognised. This has been achieved at three different levels: the regional level (the ice creating the lake by moving up the regional slope), the local level (the creation of ice pushed ridges on 2-5° slopes) and at the micromorphology level (involving modification of pebble morphology). The role of geometry has been demonstrated not only in the foregoing three fundamental conditions but also recognized in the more ephemeral circumstances where the geometry of the immediate past has an influence on the current response. For example, ice floes may drift gently ashore in a rising wind and act as protection for the shoreline from the effects of more vigorous floes arriving as wave energy builds up.

The response segment of the model diagram includes the main landforms recognized within the basin but as such they are not simply a catalogue listing but have been attributed to more than one process and more than one influence of geometry wherever appropriate.

In these ways the structure inherent in the model has brought order to the components. The use of a centrally pivoted pointer in the diagram allows the concept of time to be superimposed; one revolution of the pointer constitutes one year. Thus, conceptually, there is a constant updating of the geometry with the new season's processes operating on the landscape modified by last year's processes. In fact, this concept is completely flexible allowing a revolution of the pointer at any speed appropriate to the tempo of the process being considered - an hour, day or week.

The model structure is, therefore, applicable to any proglacial lake environment with the details probably changing slightly to accommodate the particular emphasis and blend of components prevailing, but the process-geometry-response-time relationship could stand as a basis for ordering that information.

The suite of landforms - the basis for distinctiveness

The paleontologist may take a group of specimens of a single fossil type and make an environmental assessment of the conditions under which the organisms lived. Depending upon the state of preservation of the specimens and the tolerance of the species for a wide or narrow range of conditions, the interpretation by the paleontologist will be correspondingly general or more specific. It is highly unlikely that a detailed interpretation will be possible on the basis of a single fossil type. However, the paleontologist who is able to study the range of fossil types in a bedrock stratum or a formation is able to take the collective environmental limitations of the species and make a much more specifically limited and, therefore, accurate interpretation of former living conditions.

An analagous position has been adopted in this dissertation: that the landforms must be considered as a suite. The occurrence of a fossil ice pushed ridge enables a geomorphologist to deduce certain minimum conditions prevailing not necessarily including a glacial ice dam. Similarly, a perched delta enables a geomorphologist to conclude a body of water was formerly maintained at that elevation, which in turn requires an impeding of drainage. Furthermore, closely-spaced multiple moraine ridges are known to have occurred in a variety of environments and so are not necessarily closely diagnostic of specific environmental conditions if examined in isolation.

Observation of the process of anchor ice sediment rafting with the immediate short term response of 10-30 cm of sand several square metres in area drifting about the lake, led to the inferred conclusion that a substantial dumping of sands in the offshore lake bottom sediments is to be expected. It is not known how common this process may be, but the observation that it was apparently more effective in the cool summer of 1968 than the warmer one of 1969 suggests that it may be optimally developed in the proglacial environment, and the occurrence of kames or other small sand deposits in other areas may require further evaluation.

However, despite the need to examine the suite of responses, it would be unrealistic to avoid some specific assessment of the dominant feature in the basin, the sublacustrine moraines, in part due to the time devoted to them, and in part due to the occurrence of a family of similar forms in paleogeomorphic environments of northern latitudes. A careful assessment of both geomorphic and glaciological processes, together with establishing the nature of geometric constraints and time available, enabled the elucidation of the development of sublacustrine moraines, in a manner which accommodates the field evidence available.

The processes which explain the sublacustrine moraines of Generator Lake do not necessarily explain similar forms elsewhere, for despite similarity of morphology, other aspects such as marine as opposed to fresh water and the form and thermal condition of the ice margin, would complicate their explanation, at least in detail.

Despite the validity of pursuing the processes involved in creating one landform, the relationships between such responses is an equally valid pursuit. When one examines the whole basin and the spatial relations of the features within it and considers the processes involved in producing each of the landforms comprising the suite, then the assemblage becomes much more distinctive in composition. In this context, the imprint of the sub-polar ice becomes clearly discernible, without being wholly convincing in the case of any one response. Individually, the degree of development of the features may be attributable to the proximity of cold-based marginal ice but such an attribution is difficult to identify clearly. Thus a lake basin with sublacustrine moraines, ice pushed ridges, proglacial lake terraces and/or kame-deltas and with evidence of anchor ice sediment rafts among lake bottom fines can be much more closely identified with the sub-polar proglacial environment than any one feature alone.

Quantitative relationships

A variable degree of success in establishing quantitative relationships was achieved. In the sense of numerical values, the establishment of a radiocarbon chronology is one of the major successes. This chronology superceded the lichenometrically-based provisional one and is more soundly based even when considering only the radiocarbon evidence. When consideration is extended to include the compatability of the geomorphic and glaciological evidence then this new chronology is even more strongly supported. In this context it is worth noting that the chronology, which turned out to be somewhat more complex than initially assumed, was responsible for triggering additional insights into the geomorphic evolution of the basin, thereby explaining previously puzzling responses. Perhaps, the most notable of these was the absence of substantial sublacustrine moraine development in the present lake basin, despite apparently adequate time and suitable conditions.

A second major quantitative result was the establishment of the impossibility of the sublacustrine moraines being annual features or more specifically responses to some annual process such as the influx of spring meltwater. Ten to 20 years was established as a typical time for development of each moraine in the 76 m basin. The size of an individual moraine was thus also demonstrated to be a function of time.

As part of the demonstration of the process of moraine formation it was shown that the rate of till release necessary to create the fossil moraines ranged between 6.9 $m^3/m/yr$ and 25.5 $m^3/m/yr$ which was equivalent to an average till thickness over a 500 m wide band of between 62 cm and 1.86 m. It was also shown that these amounts could in turn be released from the ice by assuming rates of forward movement of the ice compatible with or less than those presently prevailing.

It was also established that net retreat rates showed a trend to decrease with increase in lake area, an empirical result substantiated by a theoretical relationship developed by Holdsworth.

The hydrological regime of the lake was established as one which exhibits a rapid rise and more gradual decrease from a sharp peak value (three of the four years of record) 1.5-1.8 m above a winter minimum consistent with the threshold of the lake at the Clyde River outlet. The rate and magnitude of lake level rise was shown to be critical in the process of fracturing the tapered ice ramps beneath which sublacustrine moraines form. A hyperbolic relationship between maximum lake level rise and lake area was established by Holdsworth (Barnett & Holdsworth, 1974, Figure 13). This relationship is important not only in assisting explanation for the increasingly stable conditions in the 76 m lake suggested by the increasingly large moraines towards the western end of it, but also in allowing prediction for the size of those yet to be formed as, and if, the ice margin continues to retreat in the present basin.

Quantitative data on the sediment transfer by streams like the 'Robinson' River can only be considered in order-of-magnitude terms, but the values derived from the very limited sampling program are shown to be in general accord with the historical values derived from the perched delta volumes and the radiocarbon chronology.

Similarly the quantitative aspects of pebble morphology generally in the 50-100 mm a-axis range, are such that only qualitative conclusions may be drawn, although the major one is, in fact, quite an important finding. That is that shoreline processes despite their short effective time span can be demonstrated to be more rapid and efficient than frost shatter in modifying pebble shape between 50-100 mm. This was indicated not only on the present active shore zone but also on the 76 m shoreline. In this latter situation a return to periglacial (subaerial) processes for 2500 years has not demonstrably diminished the shoreline modification of pebbles which was achieved over a period of the preceding 1500 years or less. Roundness is either less sensitive or more difficult to measure effectively, or both, than sphericity, and probably than flatness also. Yet, despite the difficulty of establishing a high degree of statistical explanation for variation in roundness with either distance from the ice margin, or fetch, or these two variables in combination, it is clear that a trend toward increasing roundness, increasing sphericity and decreasing flatness can be demonstrated to occur in some complex way with time (distance from the ice margin) and fetch. The complexity may be simply a reflection of the vagaries of ice floe distribution on the rare occasions when wave action takes place, with floes protecting some beach areas and not others.

It is remarkable that such trends are discernible in an environment, and at a latitude, in which it is generally accepted that periglacial processes are active, and limited observation would suggest very modest shoreline processes.

The quantitative aspects of ice pushed ridges give some clear indication of the relevant variables. Fetch is a dominant variable in determining the size and location of ice pushed ridges. Shore slope at less than 5° and greater than 2° are pre-requisites as well as materials suitable to be pushed. The larger the calibre of the material the higher the ice pushed ridge. Obviously, an upper limiting value must occur, but a boulder 390 cm long was measured in one of the ridges. Nevertheless, that upper limiting value will be also a variable dependent upon fetch. Wind speed and direction is also critical but the least predictable of the major variables because of the conclusion that major ice push is a low frequency, high magnitude event. In such a case the track of the particular storm would be the deciding factor on wind direction for the critical minutes of the ice rafting ashore.

The climatic data indicate the extreme variability in precipitation characteristic of a desert (low precipitation) environment. In addition an important concept to appreciate is that a relatively small seasonal variation in mean summer temperatures produces remarkably different responses and substantially affects the tempo of geomorphic processes because of the general proximity of temperatures to the freezing point. At least some of these locally significant but small variations may be attributed to the 'cold' source of the ice cap. The difference between a geomorphologically active 'warm' summer and a quiescent 'cool' summer is relatively small in terms of mean temperature. Obviously the mean summer condition fluctuates about a critical threshold value, as some of both types of summer conditions were experienced in the few years of observations. To attribute a specific predictive capacity to temperature data, in terms of the geomorphic tempo of the basin, is difficult but a mean temperature for July and August combined of about 44°F (7°C) would definitely produce an 'active' summer. A mean temperature of 39° to 40°F (4°C) would, in all probability, produce a rather quiescent season with subdued geomorphic responses.

<u>The utility of the model for other sub-polar proglacial environments</u>

As the establishment of a model with a unique combination of attributes is at odds with one of the main concepts of modelling - that of seeking salient attributes of a group of similar items to better understand them all - it is necessary to extend the assessment to its applicability to the balance of the group of sub-polar proglacial environments.

The most obvious limitation to this is that the number of similar environments occurring at the present time are few. The most immediate of comparable size are those of Conn and Bieler Lakes just over 100 km north of Generator Lake also on the east side of the Barnes Ice Cap. They are at the high level lake stage as the meltwaters from them spill eastward across the divide at the easternmost end of their respective valleys. Both basins are directly comparable with Generator Lake at an earlier phase in its history and the model would fit them well. However, the delicate balance of the thermal control on the tempo of geomorphic processes is such that available photographic and occasional personal reports (Glaciology Division personnel, Dept. of Environment) suggest that those two lake basins lie across the temperature threshold in apparently perpetual 'cool' summers with minimal leads developed around the lakes.

Between Bieler Lake and Generator Lake and just north of Gee Lake a small lake has drained recently (?) revealing at least some similar features such as small sublacustrine moraines and irregular sand deposits among bedded lake bottom fines. A willow stem dated from the lake bottom yielded a date of 160±200 B.P. (GSC 1422).

The Penny Ice Cap on southern Baffin Island and other ice masses on Bylot, Ellesmere, Axel Heiberg and Devon Islands all impede drainage and create proglacial lake basins of generally rather small size, but in each of these environments the basic geometry is one of much greater relief. Such differences of relief and size are of course accommodated in the model by the geometry segment.

The great ice sheets covering much of Greenland and Antarctica dominate these areas and proglacial lakes are generally quite small, e.g. Lake Tuto in northwest Greenland. Nevertheless, such basins would be quite compatible with the model, although their geomorphic diversity of process and response would probably be somewhat less than at Generator Lake.

The utility of the model for interpreting paleogeomorphic environments

A brief glance at the Glacial Map of Canada (1964) indicates enormous areas of Canada which have been subjected to flooding by various proglacial lakes. Often when such areas are investigated in detail, the history of such a region reveals that the lacustrine episode is composed of a series of lakes of variable duration. Together these reflect the changing location of the ice margin and, therefore, the different escape routes for the meltwaters. The larger areas depicted on the map cover the south and western sectors of the Laurentide Ice Sheet. However, it is on the northern and northeastern edges of the ice mass that the more analagous paleogeomorphic environments are to be expected. Such environments which are generally only poorly known are not of the magnitude of the better known Glacial Lake Agassiz environments trapped against the regional slope. The northern ones are considerably smaller where the local topography led to the impounding of meltwaters from the ice. It is because of the former extensive areas of terrain which have been covered by proglacial lakes that a study of an existing one is significant for assisting with paleogeomorphic interpretations. The particularly significant aspect of Generator Lake is that within one basin features of the active environment and of a recently relict environment occur in close proximity and furthermore, the ice is in fact thought to be one of the last remnants of the Laurentide Ice Sheet.

It would be particularly interesting to study a latitudinal range of proglacial lake basins, all on the Canadian Shield, and of similar size with view to testing the hypothesis that as the latitude decreased the geomorphic influence of sub-polar ice decreased and glaciofluvial landforms increased in both number and development. Such a hypothesis could be tested effectively only by examining the suites of landforms developed in each basin, because of the subtle nature of the influences. In this way the concept of polar, sub-polar and temperate ice masses could be tested to see if abrupt or gradational changes occur in the paleogeomorphic responses. Insights which may be gleaned from such an examination could have spin-off benefits for examining non-lacustrine environments if greater assurance as to the properties and nature of the former ice mass could be established. The implications and need for such information are set out in a review paper by Schytt (1974) which outlines the issues of 'ice temperature and melt water' and 'ice temperature and movement'. Without clearer understanding of these two aspects of glaciology the paleogeomorphic interpretation of the former proglacial lake environments of the northern edge of the Laurentide Ice Sheet is difficult. A case is made for a careful evaluation of proglacial lake environments as a probable source of clues to aid understanding of the larger question of what happened at the base of the Laurentide Ice Sheet.

Refining the model

The conceptual model is flexible in that the items identified on the arcs are movable as to location within each sector of the framework and additional items could well be added to the response sector if another lake basin exhibited different responses. In other areas different combinations of processes may be perceived as leading to the same geomorphic response. However, the framework of this particular conceptual model is not flexible as it melds the four essential components of the environment into one coherent diagram in such a way as to demand consideration of the influence of each of the components.

Refinements could certainly be made particularly in the area of 'tempo' as the identification of the key variable - probably mean July-August temperature - has a major influence on the amount of geomorphic activity in the basin. The influence of each 1°C mean air temperature rise over some threshold value in the 5-6°C range approaches a geometric progression of influence on the tempo of geomorphic -

processes. There appears to be a similar type of negative relationship below that threshold value.

Quantitatively there is considerable room for refinement, particularly as the quantitative aspects measured were essentially exploratory sampling enabling more incisive questions to be posed.

One of the intriguing questions posed which would require work outside the lake basin is the date of the first draining of the present level of Generator Lake via Flyway Lake to Foxe Basin. That single event must have yielded a perceptible response downstream particularly as the little evidence available suggests that whatever fine lake bottom sediment there was in the earlier present-level lake was flushed out during the draining episode plus the ensuing fluvial episode. Additional materials for radiocarbon dating would be interesting, particularly if sufficient organic matter could be found in several pond bottoms at different elevations between the 76 m and present levels of the lake. Such material would not be abundant, an assumption based on the general low frequency and productivity of vegetation in the basin.

Refinement of pebble morphology relationships would require either additional insights into the form of possible interrelationships of fetch, wind speed, slope, pebble-size and time, or if morphological modification is, in large measure, a random process dependent upon location of ice floes, then a laboratory study of the sphericity producing and rounding processes would seem the more promising line of investigation. Obviously the technique of measuring roundness requires additional refinement to enable reproducibility of results by different careful investigators.

It would be desirable to observe ice push in action in a strong wind to evaluate the process first hand and to establish the rate of wind speed, fetch, material size and ice integrity and to establish additionally if ice floes also exhibit a protective role under the same conditions.

In order to make substantial progress with quantifying fluvial processes an intensive sampling program would be required to document a full season's flow in both a high and low activity year for both north and south side streams. This kind of program would require a meteorological observing program to measure precipitation and to equate variation in flow on the ice cap side to temperature change and insolation.

Refinement of the sublacustrine moraine model would not be simple or short term - an obvious test would be to wait until additional moraines are formed and accessible by echo sounder and then compare their size and morphology with the predicted values for that particular lake area. Additional volume measurements of till within the basal ice are desirable but to obtain representative data of high quality would require large volumes of ice to be melted out, which would be a logistically difficult operation.

SUMMARY: The contribution to knowledge

A contribution has been made in three categories of knowledge - new data on the geomorphology of the Generator Lake basin and its interpretation in the regional setting of central Baffin Island, innovative or improved field techniques and philosophical or conceptual development of geomorphology.

The study included the first chronology based on radiocarbon dated material from within the lake basin and it is believed the first on an extensive series of organic detritus in an active proglacial lake basin. The lifespan of the lake was shown to be somewhat longer than previously anticipated. This chronology is compatible with the regional setting which shows that the net retreat of the east side of the ice cap has been very slow over the last 8000 radiocarbon years - of the order of 8 m/yr as a long term average. Earlier judgement on the chronology was influenced by recent relatively dynamic changes in the lake basin which gave the impression that such dynamism could have created the lake in the order of 1000 radiocarbon years. A more careful evaluation indicates that the chronology is indeed of the order of 4000 years and that the dynamic events were perturbations on the tempo of the longer term process of deglaciation.

A workable model for the creation of multiple sub-lacustrine moraines was developed which included a glaciological component which was primarily the work of

G. Holdsworth; it is the first time that both process and chronology have been spelled out in detail in any of several attempts to explain multiple sub-aqueous morainic ridges. In this instance it has even been possible to predict the size of moraines which have been demonstrated by echo sounding.

Field techniques developed included the use of a ship's echo sounder to operate through lake ice cover by using a bonding medium for the transducers and which was very successful but somewhat slow. In a second application the transducers were placed in a vertical mode thus transmitting impulses horizontally against the ice margin; in this instance the technique was perhaps more successful than initially assumed as the graphical plot was erratic and diffuse but became more intelligible when the ram shape of the front of an ice ramp became known.

The use of a series of finely scribed arcs on a transparent sheet was an improvement in the technique of measuring pebble roundness as it allowed visual assessment of the best fit of the radius of curvature. The potential success of this procedure was, however, masked by an additional problem. The decision as to just which is the area of a pebble exhibiting the minimum radius of curvature from samples showing relatively little sign of rounding is difficult to make consistently between observers.

In the philosophical and conceptual development field one contribution was in the formulation of a particular diagrammatic form of presentation of the process-response model. This circular diagram emphasized the unity of the lake basin and the inclusion of the geometry segment revived emphasis on an early essential component of geography which is fundamental to the understanding of the linkage between processes and responses. Thus a concise structural concept allowed the ordering of data into a whole framework which emphasized that collectively the landforms are part of a distinctive suite. These landforms are shown to represent a particular collective response to phases of a proglacial lake which is dammed by a sub-polar ice mass. The conceptual equivalent is that of the fossil assemblage of a bedrock formation which collectively is more meaningful than the individual fossil species.

Each of these contributions was then assessed collectively to include the potential which they may have for future studies.

APPENDIX A: RADIOCARBON SAMPLE DATA

Lab No.	Field No.	Initial Date	Corrected Date	Material
GSC-1087	BDA-68-C-5	3650±140	3690±140	plant detritus

No base treatment; one 3 day count; mixed with dead gas 15 gms.

Site: 2.6 m below delta surface. 30 cm from local surface.

Lab No.	Field No.	Initial Date	Corrected Date	Material
GSC-1168	BDA-68-C-1	4600±290	N/A	plant detritus

No base treatment; one 3 day count; mixed with dead gas 4.5 gms. Photo 3-9-1968

Site: 3.3 m below delta surface. 23 cm below local surface.

Lab No.	Field No.	Initial Date	Corrected Date	Material
GSC-1177	BDA-68-C-4	1560±140	1660±140	plant detritus

No base treatment; one 3 day count; mixed with dead gas 20 gms.

Site: single bedding plane in foreset beds.

Lab No.	Field No.	Initial Date	Corrected Date	Material
GSC-1239	BDA-68-C-3	1240±210	1270±210	plant detritus

No base treatment; one 4 day count; mixed with dead gas 8.8 gms. Photo 5-3-1968

Site: single bedding plane foreset beds 3 m above lake.

Lab No.	Field No.	Initial Date	Corrected Date	Material
GSC-1244	BDA-69-C-5	3690±250	3730±250	plant detritus

No base treatment; one 3 day count; mixed with dead gas 40 gms. Photos 10-11-69; 10-12-69

Lab. No.	Field No.	Initial Date	Corrected Date	Material

Site: 80 cm from delta surface; 50 cm local depth
2 cm thick unit.

GSC-1276	BDA-69-C-3	3080±170	3090±170	plant detritus

No base treatment; one 3 day count; mixed with dead gas 10.5 gms. Hand picked by tweezers.

Site: upper third of delta; 20-30 cm from local surface.

GSC-1304	BDA-69-C-1	2480±150	2520±150	plant detritus

No base treatment; two 1 day counts; mixed with dead gas. 150 gms. Hand picked by tweezers.

Photo 3-11-68

Site: 61 m west of site GSC-1168

GSC-1315	BDA-69-C-4	2600±150	2620±150	plant detritus

No base treatment; two 1 day counts; mixed with dead gas. Weight not recorded. Hand picked by tweezers.

Site: 2 m below delta surface; 90 cm in from face; depth below local surface 75 cm.

GSC-1325	BDA-69-C-6	1480±160	1530±160	plant detritus

No base treatment; two 1 day counts; mixed with dead gas. 25.5 gms. Hand picked by tweezers.

Site: 130 cm in from face; 110 cm from local surface.
Coarse sands with mudball inclusions.

GSC-1422	BDA-68-C-6	190±200	160±200	willow stem

Base acid and distilled water; one 1 day count.

Lab. No.	Field No.	Initial Date	Corrected Date	Material

5.4 gms.

Site: Partially embedded in lake bottom sediment.

GSC-1621	BDA-69-C-5	2240±390	N/A	plant detritus

Treated with HCl then 3 distilled water washes, one 5 day count; mixed with dead gas.

1.2 gms.

Site: 50 cm local depth; 80 cm in from surface, from 2 cm thick layer.

GSC-1622	BDA-69-C-6	2180±240		plant detritus

Treated with HCl then 3 distilled water washes; one 4 day count; mixed with dead gas.

3.0 gms. Photo 13-3-69

APPENDIX B: PALYNOLOGICAL REPORTS

Palynological Report No. 71-1

Date: Jan. 13/71

Locality: Barnes Ice Cap, Baffin Island, NWT.

Lat.: 69°34'N Long.: 71°58'W

NTS: 27 C.

Submitted by: D. M. Barnett (collected by G. Holdsworth for submitter)

Field No.: GH-1

Lab No.: PL-70-52

Description of Sample: Conical mound of silt. Very low organic content.

Results and Interpretation: Pollen grains and spores of the following taxa were identified on one slide only. Figures are actual counts not percentages.

Pollen

Pinus sp.	14
Alnus	6
Ericaceae	9
Artemisia	5
Ambrosia	2
Tubuliflorae	1
Gramineae	5
Cyperaceae	5
Unidentified	6

Spores

Sphagnum	12
Moss	3

Long distance transport by wind is probably a major reason for the occurrence of all the taxa listed above because of the probable mode of formation of dirt cones. The apparent absence of Betula pollen is a puzzle as it is a common constituent of most Arctic spectra. Redeposition of grains from an older deposit by the glacier followed finally by wind transport is a possibility although the above spectra does not compare with known spectra from organic deposits in the vicinity of the Barnes Ice Cap.

R. J. Mott,
Quaternary Paleoecology
Laboratory

Palynological Report No. 71-9

Date: June 1/71

Locality: Generator Lake, Baffin Island, NWT.

Lat.: 69°34' to 69°44'N Long.: 71°32' to 71°55'W

NTS: 27 G.

Submitted by: D. M. Barnett

Field No.: BDA-68-1, -9, -18, -c-2, -c-5

Lab No.: PL - 71 - 4 and 5

Description of Sample: Three clayey, lake bottom samples and two silty deltaic sediment samples.

Results and Interpretation:

	BDA-68-1	-9	-18	-c-2	-c-5
*Picea	0.4	-	-	-	1.1
*Pinus	-	0.6	1.0	1.7	4.4
*Betula	95.8	97.0	93.0	90.1	93.3
*Alnus	3.2	7.3	4.0	8.3	1.1
*Salix	0.4	0.6	-	-	-
*Undetermined Betuloid type	0.4	1.2	2.0	-	-
Ericaceae	12.7	19.2	10.0	15.5	33.3
Gramineae	3.5	7.3	-	3.9	2.2
Ambrosieae	-	-	-	0.6	-
Artemisia	-	1.1	1.0	-	2.2
Caryophyllaceae	-	-	-	0.6	1.1
Ranunculaceae	-	0.6	-	-	--
Rumex (?)	-	0.6	-	4.4	8.9
Thalictrum	-	-	1.0	-	-
Polygonaceae	-	-	-	0.6	-
Lycopodium annotinum	3.9	9.6	3.0	6.1	8.9
Lycopodium selago	0.7	1.1	1.0	-	4.4
Pteridophyta	-	-	1.0	0.6	2.2
Polypodiaceae	3.2	5.1	6.0	3.9	6.7
Cyperaceae	4.2	4.5	3.0	5.0	3.3
Sphagnum	1.1	1.7	6.0	1.1	1.1
Bryophyta	0.7	-	-	-	-
Unidentified	1.1	1.1	3.0	0.6	4.4
Unidentified algal spore or cyst	-	-	-	-	67.8

Percentages are based on arboreal pollen (*) equalling 100 percent.

The pollen and spore assemblages in the surface samples from Generator Lake and from samples of deltaic sediment deposited at former shorelines of the lake are very similar to each other but do not compare with assemblages found in other surface samples collected anywhere in the high Arctic. The assemblages are similar to those found in the Isortoq beds by Terasmae(1) differing only in the smaller amount of <u>Salix</u> pollen present.

Long distance transport of exotic elements from farther south can be ruled out because most of the exotic types usually found in surface samples are not present. Local overrepresentation of the most abundant genus, <u>Betula</u>, is not possible because this genus is not present in the local vegetation. The only plausible explanation seems to be reworking by the ice of older deposits containing assemblages similar to those in the Isortoq beds and redeposition in the present lake as well as in the lakes which occupied the same basin in the past.

Absolute pollen abundance was determined for one sample (BDA-68-18) and a content of about 4200 fossil grains per gram of dry sediment was obtained. This is a very low abundance when compared with results obtained in other environments. Values in excess of 50,000 grains per ml. of wet sediment are generally found in recent lake bottom sediments in southern Canada. The effect of this small amount of pollen on the radiocarbon dates is probably small but if reworking of an older organic deposit is the source of these grains then presumably other, and possibly coarser,* organic matter is also involved which could affect the dates.

1. Terasmae, J., Webber, P. J. and Andrews, J. T. 1966. A study of late-Quaternary plant-bearing beds in north-central Baffin Island, Canada: Arctic, V. 19, No. 4, pp. 296-318.
2. Davis, Margaret B. 1969. Climatic changes in southern Connecticut recorded by pollen deposition at Rogers Lake: Ecology, V. 50, No. 3, pp. 409-422.
3. Kirkland, Douglas W. 1967. Method of calculating absolute spore and pollen frequency: Oklahoma Geology Notes, V. 27, No. 5, pp. 98-100.
4. Darrell, James H. II and Hart, George F. Environmental determinations using absolute miospore frequency: Mississippi River Delta: Unpublished manuscript, 14 pp.

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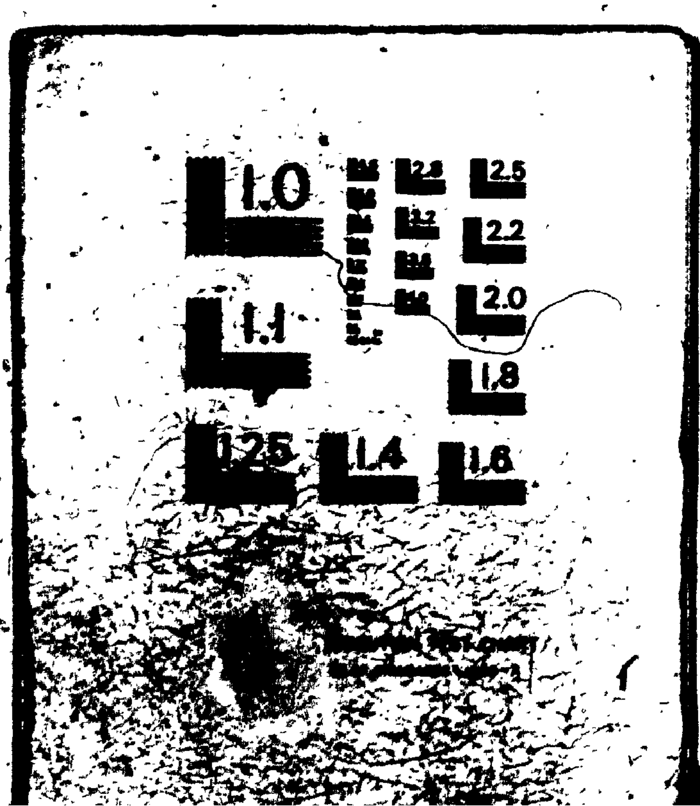
* but none were identified as <u>Betula</u> in the sample submitted. (DMB).

4



Palynological Report No. 71-11

Date: Nov. 5, 71

Locality: Generator Lake, Baffin Island, NWT.

Lat.: see below   Long.: see below

NTS: 27 G.

Submitted by: D. M. Barnett

Field No.: below

Lab No.: below

Description of Sample: Detrital and pond vegetation.

Results and Interpretation:

	BDA 69-C-6 Lat. 69°41'N Long. 71°46'N PL-71-66		GH 3 Lat. 69°36'N Long. 71°50'W PL-71-64	
	1	2	A	B
Pinus	1.2	0.7	2.9	1.9
Betula	92.0	53.0	3.3	3.1
Alnus	4.9	2.8	3.8	1.2
Salix	1.8	1.0	28.2	1.6
Ericaceae	35.6	20.5	7.7	2.7
Gramineae	3.1	1.8	38.3	77.8
Ambrosieae	1.2	0.7	0.9	0.4
Artemisia	-	-	0.5	1.2
Chenopodiineae	+	+	0.5	+
Rosaceae	0.6	0.4	-	-
Rumex type	0.6	0.4	0.5	2.8
Lycopodium annot.	14.1	8.1	-	-
Lycopodium selago	-	-	0.5	0.8
Pteridophyta	1.8	1.1	-	0.4
Polypodiaceae	4.9	2.8	-	-
Cyperaceae	8.0	4.6	6.2	3.5
Sphagnum	-	-	2.9	2.3
Unidentified	3.8	2.1	3.8	0.4

1 Percentages based on arboreal pollen equalling 100%
2, A, B Percentages based on total pollen equalling 100 %

Sample BDA-69-C-6 has been calculated two ways using a different base for the percentages. Column 1 is based on total arboreal pollen equalling 100 and gives percentages which can be compared with a previous palynological report (No. 71-9) and with percentages found in the Isortoq beds by Terasmae, Webber and Andrews (1). As can be seen from the percentages the spectrum is very similar to spectra obtained previously from samples from Generator Lake and to the spectra found in the Isortoq beds.

Column 2 shows percentages based on total pollen equalling 100 percent so that a comparison can be made with results obtained from two surface samples from small ponds near Generator Lake which are calculated in the same manner and are shown in columns headed A and B. To calculate percentages based on total arboreal pollen equalling 100 percent for these two would result in unrealistic figures because of the small amounts of arboreal pollen present. The percentages show that the surface sample spectra are unlike the spectra found in the other samples in and surrounding Generator Lake.

It appears that the Generator Lake bottom samples and the deltaic sediment samples near the lake are receiving pollen and spores from some source other than what is falling on the area at the present time. The similarity with spectra found in the Isortoq beds suggest reworking of an older deposit possibly still covered by the ice. A second possibility is that the lake is receiving reworked pollen and spores from streams flowing into it. The reworked pollen could be derived from deposits laid down during a time when the climate was better (i.e. during the hypsithermal interval) and the surrounding area supported Betula. These deposits could be upstream from the delta or be the delta sediments themselves. Other hypotheses regarding the origin of the anomalous spectra may also be valid.

Should the source be the delta sediments then the organic matter in the sediments would give valid radiocarbon dates. However, should the source be an older deposit the radiocarbon may contain enough old carbon to produce anomalously old dates with the amount of error depending on the age of the deposit being reworked and the amount of old carbon incorporated into the delta sediments.

1. Terasmae, J., Webber, P. J. and Andrews, J. T. 1966. A study of late-Quaternary plant-bearing beds in north-central Baffin Island, Canada: Arctic, V. 19, No. 4, pp. 296-318.

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APPENDIX C: BRYOLOGICAL REPORTS

Author's note: Minor editing of these reports was necessary to clarify some meanings. The main purpose in including them is to make available the moss species listings. The geomorphological interpretations by my former colleague (a botanist) are quite speculative and occasionally demonstrably untrue, e.g. B.R. 137 - the materials are *in situ*.

Bryological Report No. 18

Date: Nov. 22/69

Material: 55 single moss remains and 3 fragments of Salix

Field No.: BDA-68-C-5

Locality: near Generator Lake, Barnes Ice Cap, Baffin Island

Collector: D. M. Barnett

Radiocarbon data: GSC-1087 3690 ± 140 (corrected date)

Results: Vascular plants. Salix sp. cf. arctica. A dwarf willow.

Mosses: 30 specimens of Calliergon sarmentosum (Wahlenb.) Kindb. Stem fragments, well preserved.

15 specimens of Polytrichum hyperboreum R.Br. Whole specimens or top parts of stems. Two with sexual organs. Well preserved.

2 specimens of Aulacomnium turgidum (Wahlenb.) Schwaegr. Small stem fragment with heavy damage to leaves.

2 specimens of Pogonatum capillare (Mich.) Brid. Nearly whole plants perfectly preserved.

1 specimen of Miniobryum wahlenbergii (Web. et Mohr) Jenn. cf. var. glaciale. Whole lateral branch with leaves. They are obtuse, short but wide with strong nerve, indistinctly dentated at apex. Such forms are common in the Canadian Arctic and grow on barren soils near glaciers. They have no definite taxonomic names: "cf. var. glaciale".

Interpretation: Calliergon sarmentosum - Arctic-Alpine with relict localities on lowlands; known from many Pleistocene localities; common in Canadian Arctic Archipelago. It grows mostly along slowly running streams, at waterfalls, in tundra ponds with seasonally changeable water level, on nunataks, but rarely occurs in wet tundra.

Polytrichum hyperboreum - Arctic element, sometimes named as a variety of a cosmopolitan P. pififerum; common the High Canadian Arctic, where it grows on barren wet clay mostly of a morainic origin as well as on the weathering material including clay. Common on nunataks and near glaciers on fresh morainic substrates.

Aulacomnium turgidum - Arctic-Alpine with a main centre in the Arctic, common on climax, periodically wet, sloping tundra. Remains seem to be transported (damage to leaves). Known from Pleistocene.

Pogonatum capillare - As P. hyperboreum and usually with it on same habitats.

Miniobryum wahlenbergii - A cosmopolitan moss. In the Canadian Arctic produces dwarf forms growing mostly on muddy soils, between stones, rarely among other species. A poliedaphic moss.

All above species well represent a cold, arctic environment. Only A. turgidum can be considered as an introduced species, others grow mostly on rocky and muddy nunataks, moraines, along glacier streams and by lakes.

A negative feature of the sample is the absence of material in which the remains were deposited.

M. Kuc
Terrain Sciences Division

Bryelogical Report No. 19

Date: Nov. 23/69

Material: Single remains of Bryophytes and vascular plants

Field No.: BDA-69-C-3

Locality: Vicinity of Generator Lake, Barnes Ice Cap, Baffin Island

Collector: D. M. Barnett

Radiocarbon data: GSC-1276

Results: Vascular plants; 3 bottom parts of stems and 1 flower head of Luzula sp. - Roots.

Bryophytes: 9 fragments of Pogonatum capillare. Among them 1 stem with sexual organs. Well preserved. (See Report No. 18.)

5 stem fragments of Bryoerytrophyllium recurvirostre (Hedw.) Chem. All specimens heavily damaged.

3 whole specimens of Distichium capillaceum (Hedw.) BSG. Well preserved.

2 stem fragments of D. cf. inclinatum. Leaves short, narrow, plant dwarf, nerve strong.

1 small piece of a stem with some leaves of Encalypta sp. - excl. E.; alpine, vulgaris, affinis, longicollia, brevicolla, streptocarpa, procera; possible E. rhabdocarpa. Specimen heavily damaged.

Interpretation: Distichium capillare - Pan-continental, common in the Arctic, facultative calciphilous plant, poliedaphic, sometimes exclusive (in regions poor in calcareous habitats).

D. inclinatum - Arctic-Alpine. A plant of barren muddy soils, occurring usually with Pogonatum capillare and P. hyperboreum (see Report No. 18).

Bryoerytrophyllium recurviroste - Pan-continental or cosmopolitan common in Arctic on very different substrata. Often in association with the above species.

The sample is difficult to interpret. Luzula, D. inclinatum, Pogonatum capillare and other fossils from this locality listed in Report No. 18 seem to be occupants of the same environments.

M. Kuc
Terrain Sciences Division

Bryological Report No. 31

Date: Dec. 31/69

Material: Moss remains

Field No.: BDA-68-C-4

Locality: Vicinity of Generator Lake, Barnes Ice Cap, Baffin Island

Collector: D. M. Barnett

Radiocarbon data: GSC-1177 1660 ± 140 (corrected date)

Results: Mosses: 1. Psilopilum laevigatum (Wahlenb.) Lindb.

2. Pogonatum capillare (Mich.) Brid.

3. Aulacomnium turgidum (Wahlenb.) Schwaegr.

4. Drepanocladus latifolins (Lindb. et Arn.) Broth.

5. Hygrohypnum polare var. facatum (Bryhn) Broth.

6. Calliergon sarmentosum (Wahlenb.) Kindb.

1, 2, 4 - arctic species.

3, 5, 6 - arctic-alpine species with a distributional centre in the Arctic.

1, 2 - plants growing on barren clay; moraines, between rocks, weathered rocks, on alluvial mud; near glaciers, in higher elevations.

5, 6 - mosses growing by moving water, on rocks; in streams, by waterfalls etc.

3 - a moss of lichen-moss tundra, on seasonally wet habitats producing luxuriant forms.

4 - Arctic swamps especially around Tundra pools, a peat-forming moss.

All above species occur as frequent or common plants in the Canadian Arctic Archipelago. Other specimens belong to terrestrial species of families: Pottiaceae (cf. Bryoerytrophyllum) and Ditrichaceae (cf. Ditrichum). They are too damaged for detailed determination and have not been well preserved. Ecologically, the studied material is heterogenic; terrestrial specimens much more damaged than paludal (halophytic) ones. If they were preserved in alluvial accumulations it means that they have been transported. If they were preserved in slope, morainic or other sediments, it means that they have been deposited close to their habitats.

M. Kuc
Terrain Sciences Division

Bryological Report No. 33

Date: Jan. 2/70

Material: Yellow sediment rich in mica, sand and small plant remains

Field No.: BDA-68-C-1

Locality: Generator Lake, Barnes Ice Cap, Baffin Island

Collector: D. M. Barnett

Radiocarbon date: GSC-1168 4600 ± 290 (uncorrected)

Results: Only small damaged remains 0.25-0.5 mm. Mostly mosses. Components of wet Tundra-Drepanocladus sp. (common), Calliergan sp. (subcommon), Campylium (rare), others are Tundra plants: Pogonatum cf. capillare, Aulacomnium turgidum.

Other plant remains: fragments of wood (twigs) and in the finest fraction micro-algae (rare) - Diatomae and others. Relatively frequent occur skins of insects.

The materials are not in situ, and are floristically heterogeneous.

M. Kuc
Terrain Sciences Division

Bryological Report No. 37

Date: Feb. 3/70

Material: Mosses coated by minerals

Field No.: BDA-69-C-1

Locality: Generator Lake, Baffin Island

Collector: D. M. Barnett

Radiocarbon date: GSC 1304

Results: Minerals, sand grains, mica

Mosses: 1 - Hygrohypnum polare (Lindb.) Broth. Forms similar to typical forms than to other lower taxa. Common.

2 - H. p. var. falcatum (Bryh.) Broth. Frequent.

3 - H. luridum (hedw.) Jenn. Common

4 - Drepanocladus exannulatus (B.S.G.) Warnst. Water forms

5 - Miniobryom wahlenbergii var. glaciale (Brid.) Wijk & Marg. Rare.

6 - Aulacomnium turgidum (Wahlenb.) Schwaegr. Rare

7 - Bartramia ithyphylla Brid. Very rare.

8 - Pogonatum capillare (Mich.) Brid. Rare. Terminal cells of Lammellae very low. This feature is characteristic for specimens growing close to glaciers or extremely severe habitats. Mostly top parts of stems.

9- P. urnigerum var. subintegrifolium (Atn. & Jens.) Möll. Very rare. A glacial variety.

Interpretation: This is glacial stream material. In it there are two kinds of remains: species 1-5 (6) so-called arctic Nereidia or Amphinereidia, growing in running stream water, whole specimens without damage by transport and introduced species, 7 - 9, top parts of stems or terminal stem buds of terrestrial plants. These are accumulated in stream valleys by wind, forming so-called wind-drift sediments which after the disappearing of snow (in early summer) are introduced by water into stream moss growth. The sample is most typical of this phenomenon.

M. Kuc
Terrain Sciences Division

Bryological Report No. 137

Material: Organic detritus

Field No.: BDA-69-C-6 (2)

Locality: Generator Lake, Baffin Island

Collector: D..M. Barnett

Radiocarbon date: GSC 1325 1530 ± 160 (corrected date)

Results: Fossils - Vascular plants: 1. Salix arctica: wood, stem fragments with bark, leaves, capsules.

2. Papaver radicatum - the lower part of the stem with leaf parts.

Mosses: 3. Aulacomnium turgidum, frequent.

4. Calliergon sarmentosum, frequent.
5. Hygrohypnum polare, frequent.
6. Polytrichum alpinum, frequent.
7. P. juniperinum or P. hyperboreum, frequent.
8. Pogonatum capilare, rare.
9. Hygrohypnum var. falcatum, rare.
10. Pohlia wahlenbergii, rare.
11. Polytrichum piliferum, rare or very rare.
12. Rhacomitrium lanuginosum, rare or very rare.
13. Dicranum groenlandicum, rare or very rare.
14. D. elongatum, rare or very rare.
15. Pohlia nutanis, rare or very rare.
16. P. sp., rare or very rare.
17. Probably several more species.

Interpretation: Fossils are preserved *in situ*. Especially xerophiles which occur even in tufts and with small parts of substrates on rhizoides, e.g. Nos. 10, 15, 16.

Ecologically they represent several past environments: xerophiles or meso-xerophiles characterise dry (water deficit), loose ground, Nos. 7, 8, 10, 11, 12, 15, 16; mesophiles grow on more or less moist or relatively (seasonally dry) habitats, Nos. 3, 6, 13, 14; meso-hydrophiles or hydrophiles are characteristic for wet places, No. 4, 5, 9, partly 10. Nos. 5 and 9 usually grow at streams.

The occurrence of the mixture of such kind of ecological types most probably indicate a very low but strongly diversified micro-relief, e.g. the vicinity of shallow streams or ground strongly dissected by frost cracks or wind or other kind of erosion.

If the analyzed material contained remains of Dryas the best name for the material would be Dryas flora. It is not peat, as well as detritus. Until now we have not a proper name for it.

All these species grow on the whole Arctic Archipelago being often common components of growth types of the transitory zone between dry and wet (amphineredic) tundra (connected with running water) or rarely for some types of small tundra pools with very unstable water table.

Same or similar materials are known to me from several samples collected in the eastern Canadian Arctic.

M. Kuc
Terrain Sciences Division

Bryological Report No. 138

Material: Organic detritus

Field No.: BDA-69-C-5 (2)

Locality: Generator Lake, Baffin Island

Collector: D. M. Barnett

Radiocarbon date: GSC 1244 3730 ± 250 (corrected date)

Results: Fossils - Mosses: 1. Aulacomnium turgidum

2. Hygrohypnum polare & var. falcatum

3. Polytrichum alpinum

4. Pogonatum capillare

5. Polytrichum hyperboreum

6. Calliergon sarmentosum

7. Parts of grass-like sheaths.

Hydrophiles dominate over xerophiles.

M. Kuc
Terrain Sciences Division

APPENDIX D: TILL FABRIC DATA

Location/Date: TF1/Jul 22/69 Pit: Proximal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 214° mag. Pit Depth: 62 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	U 45	271	26	D 78	342
2	U 8	278	27	U 6	263
3	D 16	245	28	U 20	262
4	D 27	235	29	U 10	280
5	U 9	285	30	D 22	315
6	U 30	323	31	U 56	032
7	U 18	353	32	U 15	266
8	D 37	262	33	U 2	269
9	U 16	033	34	U 28	275
10	U 22	337	35	U 10	320
11	U 10	268	36	U 7	241
12	D 3	006	37	U 22	294
13	U 13	046	38	U 85	272
14	D 60	263	39	U 15	332
15	U 5	312	40	D 45	198
16	U 3	257	41	U 20	318
17	U 35	269	42	U 23	279
18	U 13	193	43	D 85	282
19	U 29	255	44	U 10	296
20	D 245	215	45	U 12	295
21	D 85	282	46	U 78	305
22	U 10	339	47	U 85	269
23	U 19	236	48	U 15	292
24	D 7	339	49	U 33	015
25	D 85	300	50	D 13	330

Distance between proximal and distal pits: 1.75 m

Location/Date: TF1/Jul. 22/69 Pit: Distal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 214° mag. Pit Depth: 70 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	U 28	174	6	U 42	046
2	D 22	066	7	U 52	062
3	U 16	125	8	D 27	105
4	D 52	089	9	D 03	134
5	D 27	048	10	U 61	048

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
11	D 33	131	31	D 45	142
12	D 09	142	32	D 04	113
13	D 39	010	33	D 12	103
14	U 24	156	34	D 24	070
15	D 22	169	35	U 07	113
16	U 12	170	36	D 21	014
17	D 28	107	37	U 11	081
18	D 34	114	38	0	106
19	D 18	093	39	U 28	026
20	D 10	087	40	D 25	054
21	D 14	090	41	D 26	159
22	D 23	012	42	D 01	016
23	D 05	047	43	U 87	243
24	D 02	109	44	U 54	074
25	U 12	062	45	D 13	133
26	D 36	133	46	D 59	169
27	U 21	088	47	U 07	102
28	D 33	122	48	D 79	148
29	D 32	103	49	D 37	116
30	D 19	106	50	U 11	017

Location/Date: TF2/Jul. 21/69 Pit: Proximal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 212° mag. Pit Depth: 50 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	Up 28	162	19	D 83	135
2	Up 33	082	20	Up 28	056
3	D 32	014	21	Up 25	147
4	D 21	020	22	D 34	209
5	Up 02	144	23	Up 25	164
6	Up 41	092	24	D 27	063
7	D 76	088	25	Up 29	116
8	D 22	139	26	D 24	004
9	Up 40	136	27	Up 30	089
10	Up 44	148	28	D 06	049
11	Up 76	143	29	Up 10	082
12	D 45	104	30	Up 11	131
13	Up 59	076	31	Up 16	122
14	Up 28	129	32	Up 18	111
15	Up 36	122	33	Up 61	002
16	Up 72	009	34	D 16	088
17	Up 40	164	35	D 07	080
18	Up 38	173	36	Up 29	117

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
37	Up 24	096	44	Up 33	089
38	Up 49	165	45	Up 29	105
39	Up 29	100	46	D 26	036
40	Up 32	142	47	D 20	012
41	Up 45	123	48	Up 31	177
42	Up 81	129	49	Up 30	163
43	Up 20	143	50	Up 45	147

Distance between proximal and distal pits: 1.40 m

Location/Date: TF2/Jul. 21/69 Pit: Distal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 212° mag. Pit Depth: 50 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	Up 10	138	26	U 54	120
2	Up 7	129	27	U 8	100
3	D 28	136	28	U 10	120
4	D 30	125	29	D 65	231
5.	D 2	119	30	D 8	102
6	D 3	103	31	U 18	143
7	D 17	116	32	U 30	125
8	D 2	156	33	U 20	144
9	D 4	134	34	U 16	162
10	D 16	113	35	U 20	101
11	D 8	096	36	D 12	169
12	D 7	111	37	U 8	097
13	D 5	242	38	D 3	096
14	U 5	157	39	D 2	088
15	U 8	162	40	D 5	049
16	U 62	134	41	U 2	055
17	U 58	094	42	D 32	110
18	U 80	157	43	D 63	156
19	D 2	110	44	U 60	111
20	U 18	153	45	U 80	093
21	U 40	113	46	U 18	119
22	D 8	104	47	D 2	062
23	U 22	200	48	U 1	215
24	D 27	227	49	D 18	115
25	D 15	095	50	U 16	122

Location/Date: TF3/Jul. 27/69 Pit: Proximal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 213° mag. Pit Depth: 75 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	U 26	124	26	U 18	117
2	U 72	082	27	U 27	074
3	D 14	164	28	U 7	111
4	D 17	066	29	U 37	088
5	U 22	152	30	D 18	148
6	D 22	008	31	U 10	131
7	D 11	039	32	D 14	100
8	U 45	111	33	U 17	155
9	D 46	144	34	U 19	056
10	U 41	136	35	U 68	136
11	U 40	143	36	U 20	018
12	U 67	130	37	D 33	024
13	U 29	098	38	U 34	160
14	U 48	118	39	D 48	146
15	D 58	088	40	U 8	065
16	U 44	118	41	U 11	062
17	U 32	157	42	D 6	104
18	U 32	146	43	D 19	169
19	D 66	137	44	U 18	000
20	U 28	074	45	D 64	142
21	U 36	135	46	U 26	174
22	U 57	136	47	U 68	167
23	D 3	131	48	U 43	132
24	U 10	155	49	D 44	101
25	U 17	158	50	U 7	089

Distance between proximal and distal pits: 1.25 m

Location/Date: TF3/Jul. 27/69 Pit: Distal

Pebble Axis a:b 2:1 No. Pebbles: 50

Moraine Axis: 213° mag. Pit Depth: 80 cm

No.	Dip °	Orientation ° mag.	No.	Dip °	Orientation ° mag.
1	D 21	060	7	D 40	075
2	D 19	091	8	D 18	158
3	D 28	115	9	U 3	137
4	D 10	087	10	U 70	174
5	D 30	128	11	U 26	064
6	U 12	121	12	D 18	088

No.	Dip °	Orientation ° mag.
13	U 21	176
14	U 25	072
15	U 4	043
16	U 11	115
17	U 27	235
18	U 12	158
19	D 32	105
20	D 12	155
21	D 38	184
22	U 8	094
23	D 23	078
24	D 33	053
25	D 34	091
26	D 27	146
27	U 3	107
28	D 26	083
29	D 15	113
30	D 55	091
31	D 13	031
32	D 15	158
33	D 5	115
34	U 19	061
35	U 27	054
36	D 18	126
37	D 8	059
38	D 58	106
39	D 75	139
40	U 20	074
41	D 7	127
42	D 5	138
43	U 5	163
44	D 10	172
45	U 3	140
46	U 23	102
47	U 17	84
48	U 52	177
49	N/A	N/A
50	N/A	N/A

APPENDIX E

SUMMARY OF GRAIN-SIZE ANALYSES (PER FRACTION, CUMULATIVE)

In Weight Percentage of Dry Mineral Matter, Smaller than 2000 Microns; Inclusive of Carbonates and Humus and Expressed in Microns

Sample Number	% > than 2 mm	2000-1410	1410-1000	1000-707	707-500	500-354	354-250	250-177	177-125	125-88	88-63	63-44	44-31.2	31.2-15.6	15.6-7.8	7.8-3.9	3.9-1.95	< 1.95	Wentworth Grade % SAND	% SILT	% CLAY
1	1.2	0.06	0.06	0.10	0.15	0.34	0.30	0.27	0.32	0.23	0.17	0.44	0.30	8.71	14.57	18.99	16.56	38.43			
		0.06	0.12	0.22	0.37	0.71	1.01	1.28	1.60	1.83	2.01	2.44	2.74	11.45	26.02	45.01	61.57	38.43	2.00	43.01	54.99
2	1.7	0.61	0.61	0.88	0.83	0.71	1.02	2.55	9.26	10.04	6.7	10.06	6.20	11.62	7.93	6.40	4.25	20.17			
		0.61	1.22	2.10	2.93	3.69	4.71	7.26	16.52	26.56	33.3	43.43	49.63	61.25	69.18	75.58	79.83	20.17	33.37	42.21	24.42
3	0.7	0.07	0.13	0.13	0.13	0.20	0.94	3.45	11.51	13.37	9.3	14.77	6.36	9.32	6.05	4.15	3.36	16.74			
		0.07	0.20	0.33	0.46	0.66	1.60	5.05	16.56	29.93	39.2	54.02	60.38	69.70	75.75	79.90	83.26	16.74	39.25	46.53	20.10
4	4.2	0.69	1.10	1.54	1.79	1.79	1.72	2.03	6.72	10.31	9.50	16.37	9.85	8.81	4.65	2.43	3.45	15.25			
		0.69	1.79	3.33	5.12	6.91	8.63	10.66	17.38	27.69	37.1	55.56	65.41	74.22	78.87	81.30	84.75	15.25	37.17	44.13	18.70
5	4.5	1.23	1.82	2.34	3.13	3.55	3.30	2.49	2.64	1.33	0.7	1.2	2.44	3.96	4.94	6.17	7.66	50.88			
		1.23	3.10	5.44	8.57	12.12	15.42	17.91	20.55	21.68	22.6	23.95	26.39	30.35	35.29	41.46	49.12	50.88	22.67	18.79	58.54
6	0.3	0.00	0.04	0.04	0.04	0.06	0.06	0.06	0.03	0.05	0.0	0.16	0.82	3.15	7.98	20.50	25.74	41.15			
		0.00	0.04	0.08	0.12	0.18	0.24	0.30	0.33	0.44	0.50	0.66	1.48	4.63	12.61	33.11	58.85	41.15	0.50	32.61	66.89
7	0.1	0.02	0.06	0.07	0.14	0.19	0.19	0.14	0.25	0.14	0.10	0.33	1.42	1.78	4.68	8.75	12.94	68.79			
		0.02	0.08	0.15	0.30	0.49	0.68	0.82	1.07	1.21	1.31	1.64	3.06	4.84	9.52	18.27	31.21	68.79	1.31	16.96	81.73
8	0.2	0.14	0.14	0.16	0.26	0.45	0.59	0.45	0.35	0.14	0.0	0.28	0.41	2.72	4.41	7.59	11.69	70.14			
		0.14	0.28	0.44	0.70	1.15	1.74	2.19	2.54	2.63	2.76	3.04	3.45	6.17	10.58	18.17	29.86	70.14	2.76	15.41	81.83
9	0.2	0.00	0.06	0.02	0.04	0.04	0.04	0.04	0.10	0.16	0.20	0.66	0.89	10.02	11.50	11.57	10.74	53.92			
		0.00	0.06	0.08	0.12	0.16	0.20	0.24	0.34	0.50	0.70	1.36	2.25	12.27	23.77	35.34	46.08	53.92	0.70	34.64	64.66
10	0.0	0.00	0.00	0.02	0.04	0.04	0.04	0.04	0.13	0.29	0.31	0.98	1.87	4.43	9.08	14.06	13.48	55.19			
		0.00	0.00	0.02	0.06	0.10	0.14	0.18	0.31	0.60	0.91	1.89	3.76	8.19	17.27	31.33	44.81	55.19	0.91	30.42	68.67
11	0.0	0.02	0.02	0.02	0.02	0.04	0.06	0.06	0.13	0.13	0.14	0.47	0.75	2.87	3.90	8.05	12.32	71.00			
		0.02	0.04	0.06	0.08	0.12	0.18	0.24	0.37	0.50	0.64	1.11	1.86	4.73	8.63	16.68	29.00	71.00	0.64	16.04	83.32

Sample Number	% > than 2 mm	In Weight Percentage of Dry Mineral Matter, Smaller than 2000 Microns, Inclusive of Carbonates and Humus and Expressed in Microns: 2000-1410	1410-1000	1000-707	707-500	500-354	354-250	250-177	177-125	125-[illegible]	[illegible]-63	63-44	44-31.2	31.2-15.6	15.6-7.8	7.8-3.9	3.9-1.95	<1.95	Wentworth Grade: % SAND	% SILT	% CLAY
12	0.8	0.02	0.06	0.02	0.09	0.13	0.21	0.41	1.33	1.87	1.91	4.19	6.75	9.81	8.24	10.46	10.45	44.00			
		0.02	0.03	0.10	0.19	0.32	0.53	0.94	2.32	4.19	6.10	10.29	17.04	26.85	35.09	45.55	56.00	44.00	6.10	39.45	54.45
13	0.3	0.02	0.02	0.02	0.02	0.04	0.04	0.04	0.10	0.22	0.30	1.14	2.67	5.22	8.57	10.51	11.53	59.54			
		0.02	0.04	0.06	0.08	0.12	0.16	0.20	0.30	0.52	0.82	1.96	4.63	9.85	18.42	28.93	40.46	59.54	0.82	28.11	71.07
14	0.0	0.02	0.03	0.02	0.02	0.02	0.02	0.02	0.04	0.04	0.05	0.30	0.10	2.49	3.43	10.06	14.44	68.90			
		0.02	0.05	0.07	0.09	0.11	0.13	0.15	0.19	0.23	0.28	0.58	0.68	3.17	6.60	16.66	31.10	58.90	0.28	16.38	83.34
15	0.0	0.00	0.00	0.02	0.02	0.04	0.04	0.04	0.05	0.10	0.10	0.38	1.14	2.39	3.31	9.14	12.95	70.28			
		0.00	0.00	0.02	0.04	0.08	0.12	0.16	0.21	0.31	0.41	0.79	1.93	4.32	7.63	16.77	29.72	70.28	0.41	16.36	83.23
16	0.1	0.06	0.06	0.16	0.36	0.64	1.00	1.44	4.25	5.77	6.07	15.08	12.08	21.29	15.72	7.80	2.37	5.63			
		0.06	0.12	0.28	0.64	1.28	2.28	3.72	7.97	13.94	20.03	35.11	47.19	68.48	84.20	92.00	94.37	5.63	20.03	71.97	8.00
17	0.1	0.02	0.06	0.14	0.33	0.49	0.51	0.45	0.70	0.72	0.74	1.75	3.27	6.35	8.32	8.66	9.59	57.91			
		0.02	0.08	0.22	0.55	1.04	1.55	2.00	2.70	3.42	4.16	5.90	9.17	15.52	23.84	32.50	42.09	57.91	4.16	28.34	67.50
18	0.2	0.02	0.04	0.04	0.06	0.10	0.21	0.25	0.48	0.54	0.48	1.14	1.70	5.13	8.98	11.24	10.65	58.94			
		0.02	0.06	0.10	0.16	0.26	0.47	0.72	1.20	1.74	2.22	3.36	5.06	10.19	19.17	30.41	41.06	58.94	2.22	28.19	69.59
19	11.8	0.02	0.04	0.02	0.09	0.13	0.11	0.11	0.25	0.25	0.32	0.76	0.35	4.42	8.24	13.28	12.40	59.21			
		0.02	0.06	0.08	0.17	0.30	0.41	0.52	0.77	1.02	1.34	2.10	2.45	6.87	15.11	28.39	40.79	59.21	1.34	27.05	71.61
20	2.4	0.04	0.11	0.25	0.47	0.68	0.91	0.89	1.16	1.06	0.85	1.17	1.60	2.11	5.28	8.90	11.59	62.93			
		0.04	0.15	0.40	0.87	1.55	2.46	3.35	4.51	5.57	6.42	7.59	9.19	11.30	16.58	25.48	37.07	62.93	6.42	19.06	74.52
21	0.1	0.02	0.09	0.22	0.41	1.21	0.56	0.24	0.20	0.19	0.11	0.26	1.76	6.97	19.20	20.28	14.04	34.24			
		0.02	0.11	0.33	0.74	1.95	2.51	2.75	2.95	3.14	3.25	3.51	5.27	12.24	31.44	51.72	65.76	34.24	3.25	48.47	48.28

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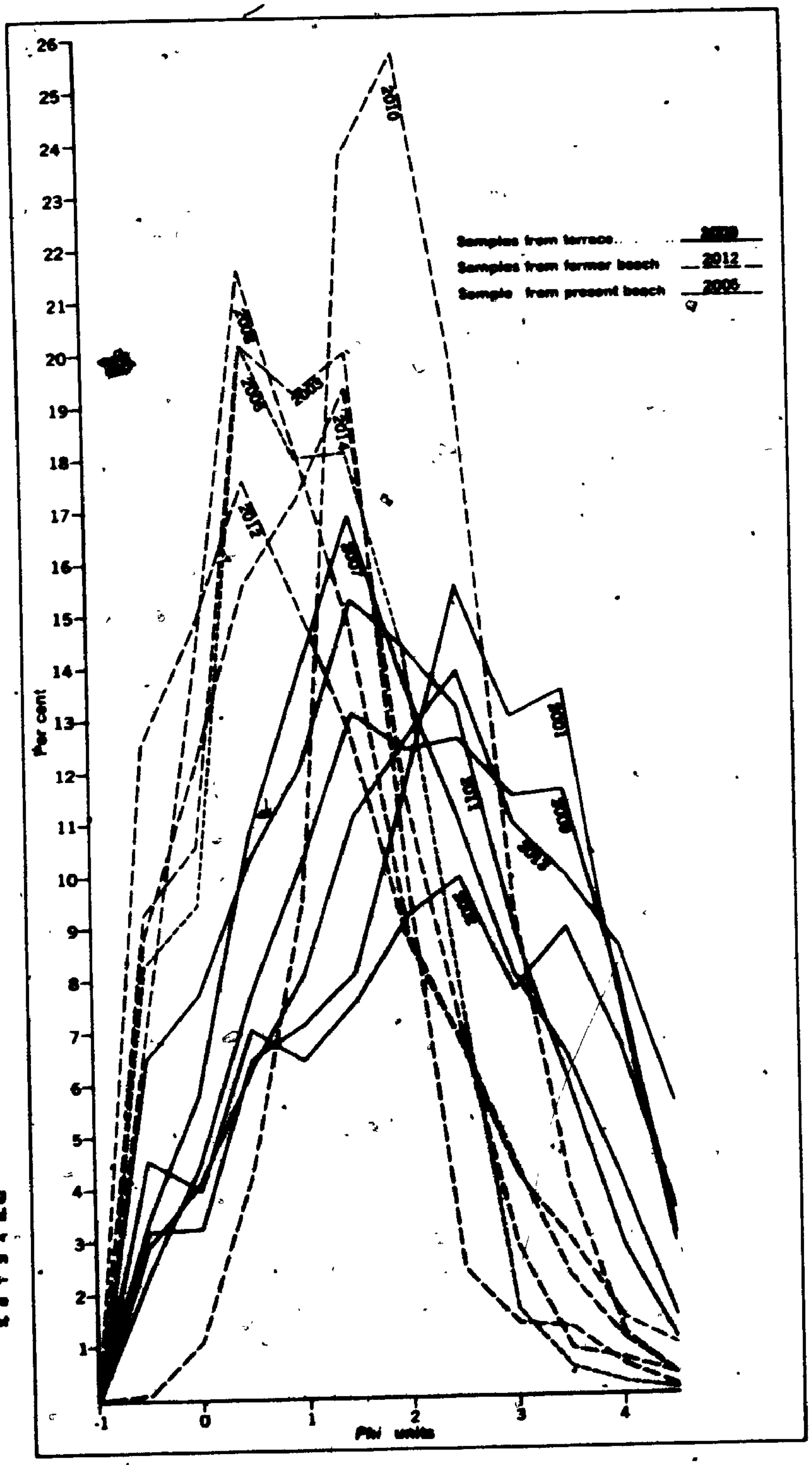

FIGURE 10
Grain sizes in *Phi* units showing contrast between beach and terrace samples. Particles finer than 44 microns have been omitted. The maximum omission occurs in No. 2013 where it amounts to 7.8 per cent of the sample.

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DEVELOPMENT, LANDFORMS AND CHRONOLOGY OF GENERATOR LAKE, BAFFIN ISLAND, N.W.T.*

D. M. Barnett

ABSTRACT

The development of Generator Lake is traced from its beginnings as a proglacial lake ponding against the regional watershed to a lake with an area of some 120 km^2 still dammed by part of the Barnes Ice Cap after a 26-km retreat of the ice margin. Further retreat resulted in the draining of the lake by a new outlet into the Clyde River and a fall in lake level of 75 metres. The trend of net retreat has continued for another 5 km to the present ice cliffs and, since 1948, the average rate of retreat has been 25 metres per year. The present ice cliffs stand more than 17 metres above water level at their highest point. A description is given of the following assemblage of landforms, developed in the proglacial lake environment and exposed when the former lake drained: the shoreline, ice-pushed ridges, deltas, cross-valley moraines and a terrace. The stratigraphic relationship between the last two suggests a subglacial origin for the moraines. A similar set of landforms are developing in the present lake. A critical review of lichenometry indicates that caution is necessary when applying this useful dating technique close to the size limit of a species. A provisional chronology covering the last 1,500 years is proposed, extending from establishment of the higher lake between 1,500 and 1,100 years ago to its draining between 950 and 750 years ago. Careful levelling of the shoreline indicates that it is not measurably deformed.

RÉSUMÉ

L'histoire de l'évolution du lac Generator le montre à ses débuts comme un lac proglaciaire formant retenue des eaux d'écoulement régionales; par la suite il est devenu un lac d'une superficie d'environ 120 km^2 encore endigué par une partie de la calotte glaciaire Barnes, après un recul de 26 km de la rive du glacier. Un recul plus poussé a eu pour résultat l'écoulement des eaux du lac par une nouvelle décharge dans la rivière Clyde et la baisse de 75 m du niveau du lac. Le recul net s'est poursuivi sur 5 km de plus jusqu'aux escarpements de glace actuels; depuis 1948, l'allure moyenne du retrait est de 25 mètres par an. Ces escarpements sont à plus de 17 mètres au-dessus du niveau de l'eau à leur point le plus élevé. L'auteur donne une description de l'assemblage des formes de relief qui se sont développées dans le milieu ambiant du lac proglaciaire et ont été mises à nu lorsque l'ancien lac a drainé la rive, les arêtes poussées par la glace, les deltas, les moraines transversales et une terrasse. La relation stratigraphique entre les deux dernières évoque une origine sous-glaciaire pour les moraines. Un assortiment similaire de formes de relief est en voie de formation dans le lac actuel. Un examen attentif de la «lichénométrie» indique qu'il faut être prudent quand on emploie cette technique de datation et s'en tenir à la grosseur limite d'une espèce. L'auteur propose une chronologie provisoire embrassant les 1,500 dernières années. Elle s'étend depuis l'établissement du lac supérieur, datant de 1,500 à 1,100 ans, jusqu'à son assèchement datant de 950 à 750 ans. Un nivellement soigné de la rive prouve qu'elle n'est pas sensiblement déformée.

MS submitted June, 1967.

*A shorter paper based on some of this work was presented at the meeting of the Canadian Association of Geographers in Sherbrooke, Quebec, May 1966.

INTRODUCTION

This paper is based on observations made during parts of the 1965 and 1966 field seasons. More detailed work and other papers are planned or in preparation, for example, a paper on echo sounding of the present lake elaborating on some of the problems outlined here.

Generator Lake is one of several proglacial lakes located at the margin of the Barnes Ice Cap and lies in north-central Baffin Island on the interior upland termed by Bird (1967, p. 62) the *Baffin Surface*. The ice cap has disrupted the drainage of the area and impounded several lakes such as Conn and Bieler in the northeast – the largest existing proglacial lakes in North America. Other large proglacial lakes, now drained, existed formerly around the ice margin (Ives, 1962, p. 199, Andrews, 1965), and thus a detailed study of one which is only partially drained offers an unusual opportunity for comparative studies of the changing proglacial environment (*see* Price, 1965).

In 1950 the Arctic Institute Baffin Island Expedition had a major camp (A2) at the southwest corner of the most southeasterly of the impounded lakes to which they gave the name Generator Lake (Baird, *et al.*, 1950). Subsequently, Goldthwait (1951) wrote a major paper on the development of nearby end moraines and later Weertman (1961) considered them theoretically. The work reported here represents the first effort directed to studying the landforms and chronology of the lake basin itself. As the lake is proglacial, the characteristics of that part of the Barnes Ice Cap damming it must be considered as an integral part of the study.

THE GEOGRAPHICAL SETTING

Generator Lake (69°33′ N 71.°45′W) (Figure 1) is located at the southeastern margin of the Barnes Ice Cap at an elevation of approximately 400 metres. It lies in a southwestward sloping valley very close to the present drainage divide from which the rivers draining from the southern limit of the ice cap flow east, west and south. The source of the eastward flowing Clyde River is the overflow waters of Generator Lake. Earlier the lake basin had been the preglacial headwater of a southwestward flowing stream that formerly drained into Foxe Basin by way of Flyway Lake and Flint Lake. The subglacial topography of the ice dam was first shown by gravimetric profiles obtained by Littlewood (1952, p. 123), and data on the sub-lacustrine morphology obtained by the present writer confirm this postulated former drainage route.

The present lake has maximum dimensions of 13.5 km by 6.5 km and drains approximately 410 km^2 of ice-free land as well as a portion of the Barnes Ice Cap. The shoreline of the lake is extremely crenulate as partially emerged cross-valley moraines severely indent the eastern two-thirds of its shoreline (Figures 1 and 2). These moraines are similar to those studied in detail and described by Andrews (1963a, 1963b) and by Andrews and Smithson (1966). In the northeast part of the lake basin, drainage is locally impounded between individual cross-valley moraines to form elongate ponds, with the occasional narrow breach in the moraine maintaining the drainage system. The Clyde River is the only outlet of the present lake.

DEVELOPMENT OF GENERATOR LAKE

THE FORMER LAKE

In the subsequent discussion of the lake phases it is essential to keep in mind its proglacial origin, as its area and depth at any given time depend upon the location of the ice margin at that period and on the position of low points in the watershed. In general, the lake progressively increased in area and depth as the retaining ice-barrier retreated southwestward, down valley. This general trend was reversed whenever a lower col was uncovered to form a new overflow channel. In the case of the present outlet this process has resulted in a decrease in depth of some 75 metres. Continuing retreat of the present ice margin would lead to the draining of the lake and to the re-establishment of the former southwestern drainage route into Foxe Basin.

The initial phase in the development of the lake was the ponding of meltwater in the extreme northeastern area of the lake basin, adjacent to the watershed (Figure 1). Either concurrently or probably shortly after this initial ponding at least one other small higher level lake (*A* in Figure 1) was ponded against

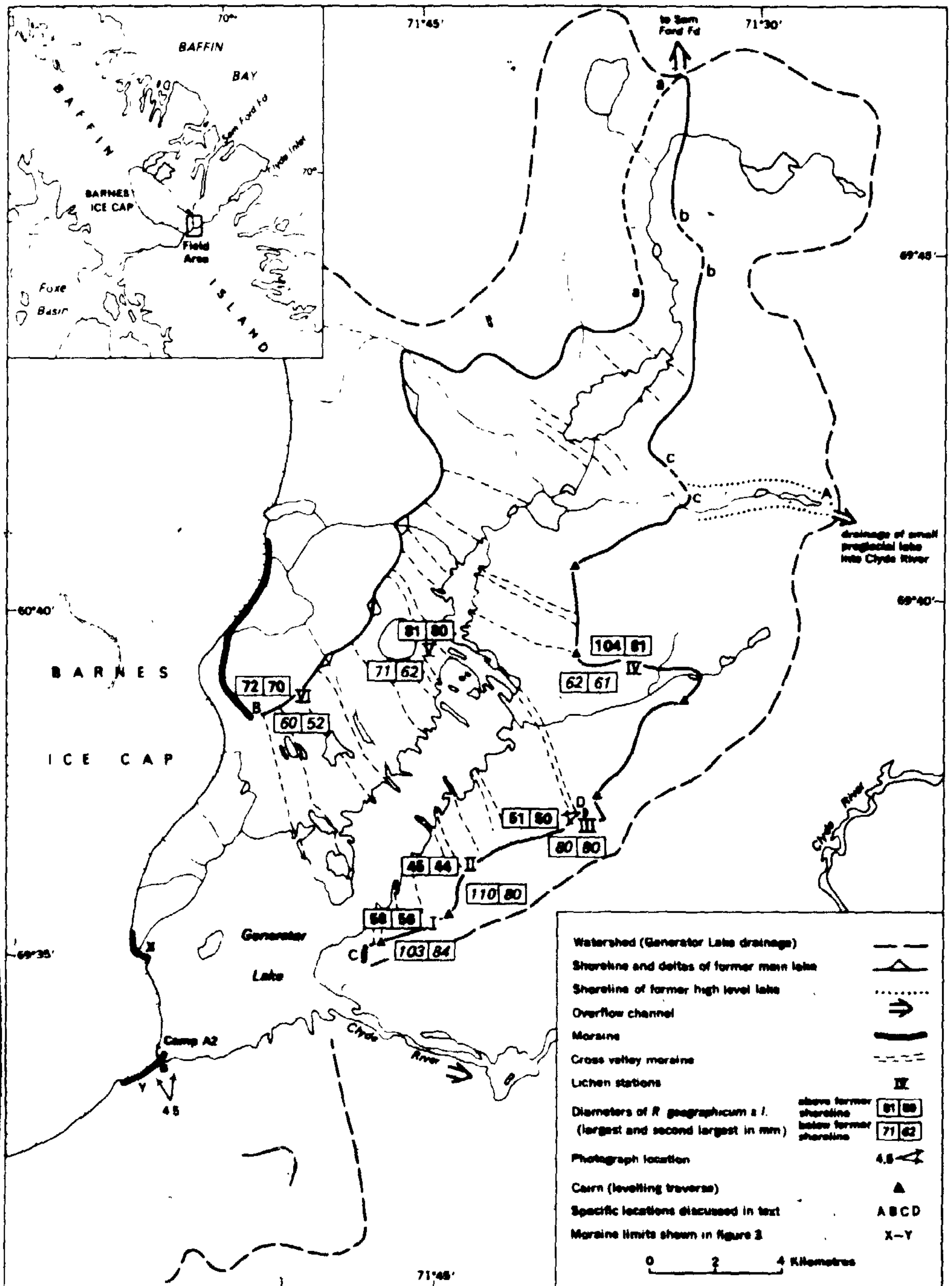

FIGURE 1. Generator Lake drainage basin showing the former and present extent of the lake.

FIGURE 2. Part of air photograph A15898-45 (21,500 ft. 6 in. lens)—(1) The island in the former lake, (2) ice-pushed ridge, (3) deltas of the former lake, (4) cross-valley moraine, (5) Barnes Ice Cap, (6) glacial lineations, (7) glacial drainage channels.

the ice margin in a tributary valley some 15 metres above the main lake level. Evidence for this small high level lake take the form of small marginal sand deposits and a small but well-defined overflow channel which carried its discharge into the Clyde valley. After an 11-km down valley retreat of the ice dam, the smaller lake must have drained into the main one which, at this stage, was increasing in width and depth.

The former Generator Lake, with an area of 120 km² at its greatest extent, was larger than the present one. Close to its northern shore was a nearly circular island which is now "dry" though it stands out remarkably well on air photographs (Figure 2) as does the prominent former main shoreline. The extent of this shoreline, shown on Figure 1, marks a level of the lake 75 metres above the present lake surface.

At the maximum extent of the former lake, the shoreline at *B* (Figure 1) reached to within one km of the present ice margin where it terminates abruptly. On the south side of the lake at *C* the shoreline becomes progressively less distinct and finally disappears some 5 km from the ice margin. This termination marks both the former extent of the lake and the position of the ice dam. From the present ice margin where retreat has been negligible a moraine extends down to the old northern shoreline and offers additional supporting evidence for delimiting the maximum extent of the former lake. On the south shore a much less distinct moraine is traceable above the shoreline though it is not possible to follow it any farther than indicated at *C*.

Above the shoreline other specific phases in the 26-km retreat from the watershed are not well marked, though to the northwest of the upper end of the lake some former ice marginal drainage channels are visible (Figure 2, no. 7); to the south a small ill-defined moraine (not shown) may indicate another phase. Below the former shoreline, evidence for phases of retreat is much more apparent. Although the numerous cross-valley moraines mark such phases, their exact relationship to the ice margin is, as yet, unproven.

During the evolution of the lake the shoreline was metachronous, though during any pause at its maximum extent its morphology may have been synchronously modified. These facts have significance when lichenometry is considered as a basis for chronology.

Drainage of the lake

The overflow water of the former lake spilled out at the head of the valley into the Sam Ford River system. The shoreline is traceable right up to the overflow channel though its morphological development is weak at some points (Figure 1 a a, b b, c c).

With the progressive retreat of the ice margin a new outlet via the Clyde River was opened and is still the drainage route used. Morphological evidence for this is supported by the occurrence of outwash along the north side of the Clyde River valley, where the first drainage breakthrough would have occurred. Much of the outwash stands several metres above the present river level, though it is not possible to say how much of this height difference is due to down-cutting of the former watershed and how much to deposition during the fall in level of the lake. However, lack of contrary evidence between the former and present shorelines, suggests that the fall in level was rapid.

No evidence was found for the draining and re-establishment of the former lake at the same level. The only other possible drainage route is the preglacial one via Flyway Lake, yet Løken and Andrews (1966, p. 350) conclude that the Ice Cap has persisted since the Wisconsin maximum. Hence it is assumed that, despite minor fluctuations of the ice margin, the lake only drained to its present level once.

The present lake and the adjacent ice margin

Initially, the drop in lake level resulting from the breakthrough into the Clyde River valley drastically decreased the lake volume, though subsequently, as the ice cliffs have receded another 5 km, this volume has been increasing. On air photographs flown at different dates, the island in the main body of the present lake shows evidence of differing degrees of submergence, but this is probably due to annual variations in lake level. Ward (1959, p. 437) reported a rise in level of about 0.6 metre in mid-July, 1950, and in 1966 a rise of 0.7 metre was noted between the first formation of a shore lead and July 14. In both 1965 and 1966 a fall in level of the same magnitude was observed by the third week in August despite the strong contrast between the warm weather of 1966 and the cool weather of 1965. The difference in mean temperature anomaly for August was 5°C (9°F) (*Monthly Record: Temperature Anomaly Maps*). This suggests that the annual variation of water level is not primarily controlled by temperature.

The estimated mean net retreat of the ice cliffs was between 5 and 7 metres per year from the time of the draining of the former lake to 1948 (*see* chronology section). Since the initial air photography in 1948, the rate of retreat of the ice cliffs is well documented by subsequent photography in 1960 and 1961 and a much more rapid rate of retreat is indicated (Figure 3a (1948) and 3b (1961)). Moreover, near-vertical photographs taken from a helicopter by a hand-held camera in August 1966

show further substantial retreat (Figure 3c and 3d). The retreat between 1948 and 1966 was approximately 400 metres, an average rate of almost 25 metres per year. Annual change, however, is erratic. For example, photographs (not shown) indicate a retreat of 90 metres at the northern end of the ice cliffs between June 30, 1960 and July 19, 1961.

The morphology of the ice cliffs has also changed in the past two decades. A photograph of the ice cliffs taken by Montgomery Ritchie of the 1950 Arctic Institute Expedition (Rand Corp., Figure 37) bears the caption "100 foot ice cliffs" probably a visual estimate only. Measurements made from a stereo pair of the 1948 air photographs give a maximum cliff height of 17.5 metres above lake (57 ft) (Mizerovsky, personal communication, 1966). In 1965 the highest cliff measured 17.2 metres (56 ft) on the helicopter altimeter, with an estimated error of ±5 feet. However, ground photographs (Figures 4 and 5) show the change in morphology and, at the extreme south end, the virtual absence of cliffs in 1965.

LANDFORMS OF THE FORMER LAKE

The shoreline

The shoreline has an observed maximum width of 60 metres, though generally it is less than 10 metres, and its morphological development is variable. In some places the form is etched into bedrock and is clear of depositional material; in other places a sandy beach has been left. The most characteristic form however, is that shown in Figure 6. Close to the original overflow channel where it is oldest, the shoreline is less pronounced than farther to the south and west. This difference in development

FIGURE 3

The ice cliffs damming Generator Lake: (a) in 1948 (part of air photo T219c-173), (b) in 1961 (part of air photo A17042-128), (c) in 1966 (oblique photo from 8,500 ft.) and (d) also in 1966, (oblique photo from 8,500 ft. taken from opposite direction to (c)). Note the retreat of the ice margin between the lateral moraines. Scale of (a) and (b) is 1:35,000, (b) having been enlarged.

TIGHT BINDING

FIGURE 4
The ice cliffs photographed from the vicinity of the Arctic Institute Expedition (1950) Camp A2. Photograph taken August, 1950 (courtesy Arctic Institute of North America Library). See also Baird, 1950, photograph p. 140.

FIGURE 5
The ice cliffs photographed from the vicinity of Camp A2, photographed August, 1965. Note that the top block has fallen from the cairn.

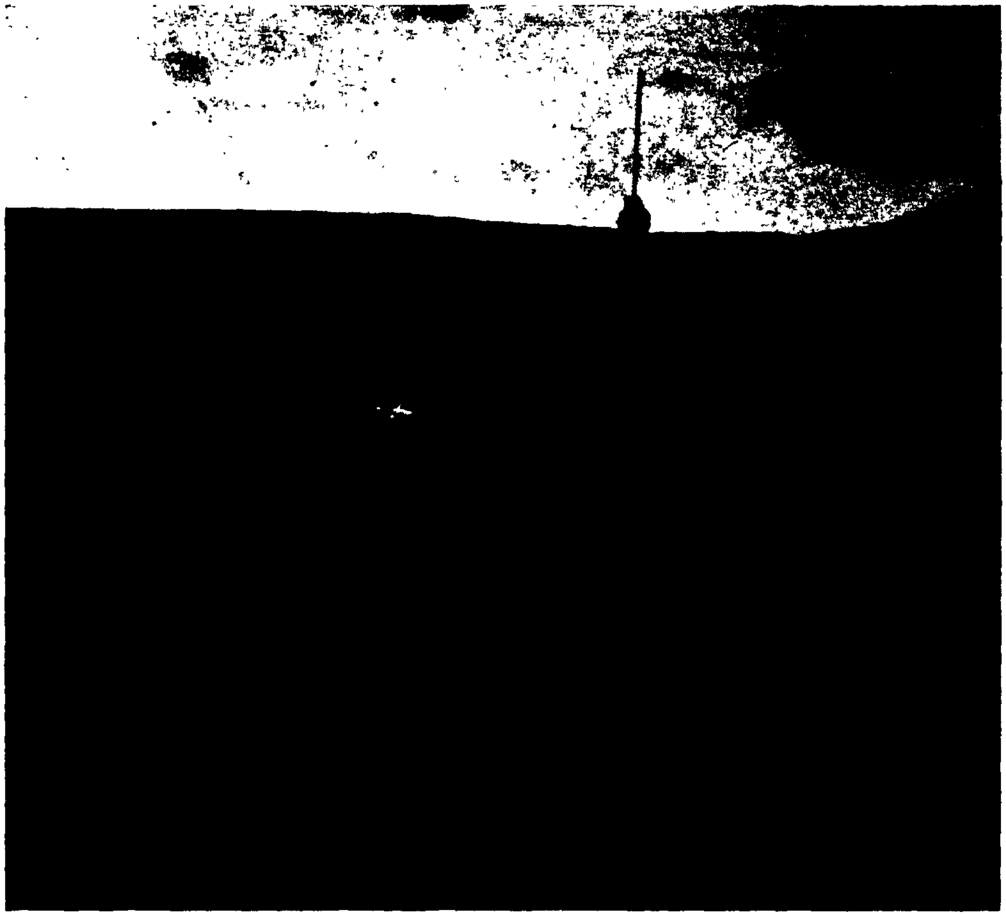

FIGURE 6. Former Generator Lake shoreline in vicinity of lichen station I.

is probably attributable to the disparity of fetch between the wide and narrow sections of the former lake.

*Ice-pushed ridges**

Several of these ridges occur just above the shoreline. They are composed of large boulders and reach heights of 2 to 3 metres (Figure 7), thus emphasizing the shoreline morphology.

Ice-pushed ridges occur on part of the eastern shoreline in the central section of the former lake and also on the east shoreline of the "island" (2 on Figure 2), locations that are gently shelving close to shore. As creation of the ridges depends both on the configuration of the shore and on the prevailing wind direction during break-up, no specific distribution pattern is apparent, for break-up, if and when it occurs, can be the result of a single storm with consequent variations in wind direction.

Deltas

Wherever large streams entered the former lake, sandy deltas were built (Figures 1 and 2). The largest and best developed are those located on the lake shore closest to the Ice Cap from which a great amount of the construction material must have come. Two of the larger deltas no longer have any drainage line associated with them and this indicates that the former stream was, at least partially, ice-marginal.

*Term as adopted by Nichols (1952).

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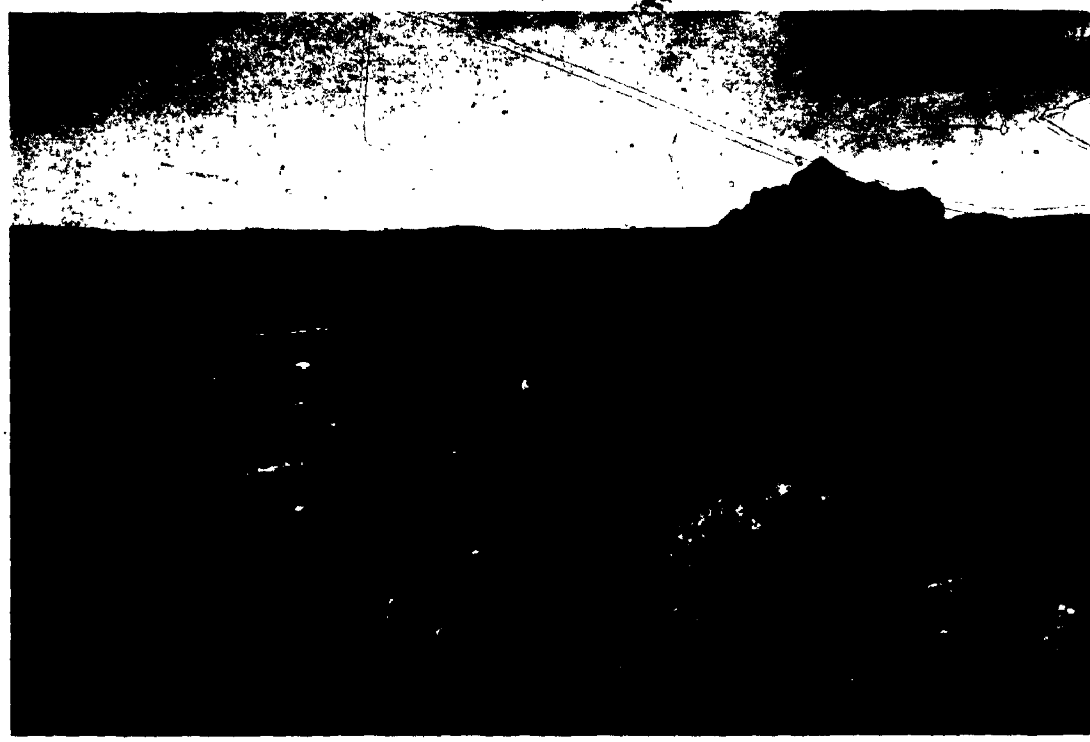

FIGURE 7. Ice-pushed ridge above the former shoreline (foreground).

Cross-valley moraines

The occurrence of these striking landforms is restricted to the lake basin below the level of the former shoreline. As observed by Andrews and Smithson (1966, p. 273) in their study of the Isortoq moraines this fact is a major diagnostic characteristic of this type of landform. Another distinct characteristic noted by Andrews (1963a, p. 63) and strongly substantiated by Smithson (1965, p. 72) is the asymmetry of the slope angles of the two sides of this type of landform. Paired slope measurements were made on sixteen moraines in the Generator Lake area in localities where they are distinct in form. Most showed steeper distal slopes, often 10 degrees steeper than their proximal counterparts (Figure 8). Each measurement was based on the mean of two readings, one up and one down the slope. The distal-slope angles ranged between 16 degrees and 32 degrees and the proximal-slope angles between 8 degrees and 27 degrees.

The vertical development of the cross-valley moraines tends to be greater toward the centre of the lake basin than toward the shore. A similar increase in amplitude and definition is apparent between the small moraines in the northeast near the head of the lake basin and the larger moraines in the southwest where the lake was wider and deeper.

Other landforms

A linear, sandy, flat-topped, dissected feature occurs from 2 to 4 metres below the old shoreline on the south side of the former lake (Figure 9). It is tentatively termed a proglacial lake terrace. To date, insufficient work has been done to offer an adequate genetic name for the feature, though it is felt that it may have sufficiently distinctive characteristics to warrant a specific term. Ives (1962, p. 198) has used the descriptive term "kame-delta" for apparently similar features. The valley-side location suggests that it could be either a beach deposit or a kame terrace. In fact Charlesworth (1957,

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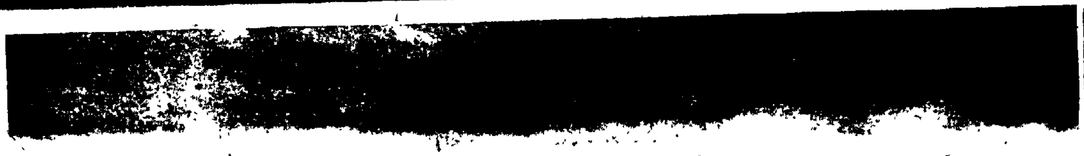


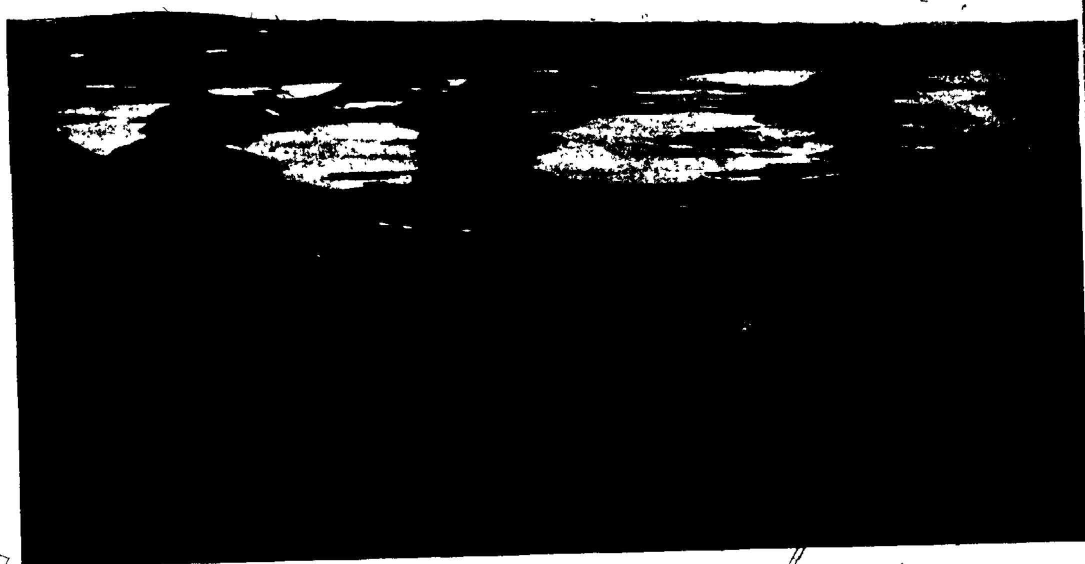


FIGURE 8. Emerged cross-valley moraines. Note their asymmetry, arcuate shape and best development in the central section. The ice cliffs are to the right.

FIGURE 9
Low level aerial oblique looking west-southwest along the old beach, showing the shoreline and dissected terrace on the south side of Generator Lake. Cross-valley moraines are present below the shoreline though the viewing angle distorts their relief. Note the marked contrast in tone above and below the shoreline (Photo: J. D. Ives).

p. 416) states: "Kame terraces may be easily confused with the strandlines of glacier lakes; the mistake has not infrequently been made." Preliminary work suggests that this feature is neither. Apart from the occasional cobble-size boulder on the surface, it is composed of bedded sands which dip obliquely away both from the shoreline and from the former ice margin.

Grain size analyses of six samples from the terrace, five samples from the abandoned beach and one sample from the present beach of Generator Lake are shown in Figure 10. The five samples from the abandoned beach are paired with terrace samples. Contrast in the degree of sorting, as determined by the Trask sorting coefficient and logarithmic coefficient, are shown in Table 1. The logarithmic coefficient, being an arithmetic measure, enables a mean to be calculated for comparative purposes (Krumbein and Pettijohn, 1938, p. 232). Friedman (1962, Table 5, p. 752) has reassessed the class limits of sorting proposed by Trask and would thereby classify the beach sediments as *moderately sorted* and the terrace sediments as *poorly sorted*. The poor degree of sorting implies a glaciofluvial origin. Deposition at the progressively retreating ice cliff where an ice marginal stream enters the lake is a probable explanation of its origin from the evidence currently at hand.

Stratigraphic relationship between two cross-valley moraines and the terrace

The stratigraphic relationship between these two features (at *D* on Figure 1) offers additional evidence for determining their respective origins. At a few localities along the south shore the moraines meet the terrace and at two points excavations were made to expose the junction for several metres. In both cases the sands of the terrace lay stratigraphically above the till of the moraines (Figure 11), the sub-surface boundary between pebble-free sand and stony till being very distinct. This marked stratigraphic relationship indicates the moraine must have formed first. Andrews and Smithson (1966, p. 289) favour two hypotheses for the formation of moraines – either an ice-frontal or in a sub-glacial position only a short distance behind the ice cliffs. In the event then, of moraine formation at the ice front in Generator Lake, contemporaneous lateral deposition of glaciofluvial material could be expected. This does not appear to have occurred and the evidence is therefore more in favour of the sub-glacial hypothesis.

Table 1

Sorting coefficients of grain size analyses

Morphological form	Laboratory number	Trask sorting coefficient	Logarithmic coefficient
"Proglacial lake terrace"	GBL 2001	1.95	0.290
*"Proglacial lake terrace"	GBL 2002	2.92	0.465
*"Proglacial lake terrace"	GBL 2007	1.78	0.250
*"Proglacial lake terrace"	GBL 2009	2.13	0.328
*"Proglacial lake terrace"	GBL 2011	1.87	0.272
*"Proglacial lake terrace"	GBL 2013	6.48	0.812
			Mean 0.403
*Abandoned beach (south shore)	GBL 2005	1.52	0.182
*Abandoned beach (south shore)	GBL 2008	1.58	0.199
*Abandoned beach (south shore)	GBL 2010	1.38	0.140
*Abandoned beach (south shore)	GBL 2012	1.81	0.256
*Abandoned beach (south shore)	GBL 2014	1.69	0.228
			Mean 0.201
Present beach (south shore)	GBL 2006	1.54	0.188
Kame delta (present lake)	GBL 2023	3.87	0.508
Kame delta (present lake)	GBL 2024	1.97	0.294
Kame delta (present lake)	GBL 2025	1.52	0.182
			Mean 0.355

Samples marked * are paired; *i.e.*, beach sample taken directly above terrace sample at approximately 100-metre intervals from west to east.

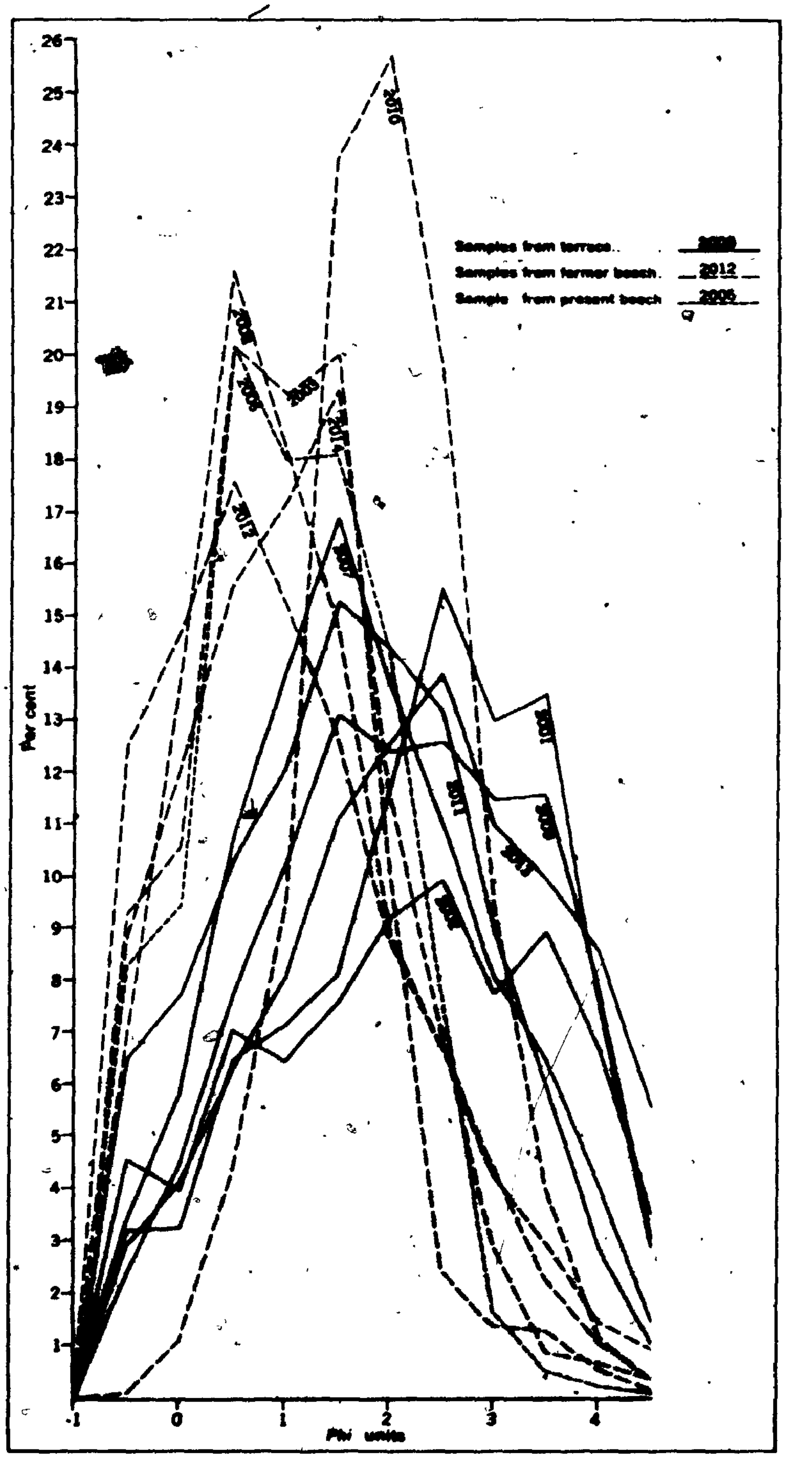


FIGURE 10
Grain sizes in *Phi* units showing contrast between beach and terrace samples. Particles finer than 44 microns have been omitted. The maximum omission occurs in No. 2013 where it amounts to 7.8 per cent of the sample.

FIGURE 11
Terrace sands overlying the till of the cross-valley moraine.

Landforms of the present lake

A similar series of landforms is associated with the present lake as with the former one. Particular investigation of submerged moraines by echo sounding will be continued and will be reported on separately.

A feature peculiar to the present lake is a "kame delta" that is forming between the proximal face of the lateral moraine and the ice at the south end of the ice cliffs. Variable sorting is apparent from three grain size analyses from this delta (Table 1). No investigation of possible sub-lacustrine prolongation of the delta into a terrace form has yet been undertaken.

Ice-pushed ridges occur around the edges of the present lake and Ward (1959, p. 437) observed and reported their mode of formation, though such vigorous genesis as he records is not an annual event.

Large sandy deltas are being built where streams enter the north side of the present lake.

THE AGE OF THE ABANDONED SHORELINE

Lichenometry as a dating technique

Beschel (1961) discussed the application of lichenometry to glaciology and physiography. Recently, Andrews and Webber (1964) assessed it as a geomorphological technique for the area northwest of the Barnes Ice Cap and found it a reliable relative dating method. More recently Jochimsen (1966) has questioned the validity of the technique and suggested that earlier Alpine correlations were fortuitous. On Baffin Island, Harrison (1966, p. 58) compared dating techniques in a small area and concluded that lichenometric discrepancies were probably microclimatic in origin. Nevertheless, lichenometry still appears to be the best available "absolute" dating method in the Arctic north of the tree line where sequential photography or material suitable for Carbon-14 dating is lacking. Although Reger and Péwé (*in* Péwé, *in press*) were able to use dendrochronology in conjunction with lichenometry in central Alaska, stems of arctic willow (*Salix Arctica*) collected by the writer close to the Barnes Ice Cap rarely showed more than 20 growth rings.

The writer was one of the persons involved in sampling for Andrews and Webber and is therefore familiar with the general problems of the technique which, it is emphasized, should remain simple in order to be a practical geomorphological tool. Andrews and Webber dealt with the problems of a non-lichenologist confusing sub-species in the field but they concluded that measuring only the largest thalli would favour sampling the fastest growing sub-species (Andrews and Webber, 1964, p. 85).

The assumptions made in lichenometry are:

– that identification of species is correct and that sub-species with the fastest growth rate would occur most frequently in the ten largest measurements at any one site. In practice *Rhizocarpon geographicum sensu lato* is easily identified by its green and yellow colouration. In the present study, diameter measurements of this lichen were made from scanning about one thousand thalli of all sizes at each station.

– that the environment was lichen-free on exposure to sub-aerial conditions.

– that micro-environmental conditions are similar in any one area sampled. This means in practice that snow patch sites and moisture retaining hollows in rocks are avoided.

– that the diameter growth rate is linear for the time span being considered.

– that the largest thalli are the oldest and the optimally growing ones.

As a result of the distribution pattern of measurements taken in the present study the author has some reservations about the last assumption. Most of the measurements taken were of *R. geographicum s. l.* though some were of *Alectoria* and of *Buellia*. Figure 1 shows the largest and second largest diameters of *R. geographicum* at each of six sites, both above and below the shoreline. No sampling area was of a predetermined size but each was at least 10 metres away from the shore zone and did not extend laterally more than 20 metres either side of the starting point. This minor variation from Beschel's technique was used by Burrows and Lucas (1967, p. 467) "because of the irregularity of the terrain." It was also necessary to avoid snow-patch sites.

Examination of the diameter measurements of lichens from above the shoreline (Figure 12) reveals that three are significantly larger than the rest (60 recorded and about 6,000 scanned). None occur with diameters between 85 mm and 102 mm inclusive and yet a much greater uniformity of size is apparent if only the second largest diameters are considered. The hiatus of 18 mm (Figure 12a, I, II and IV) is large and would represent approximately 250 years without further successful colonization, assuming the revised growth rate calculated by Andrews (*in* Løken and Andrews, 1966, p. 359). Although the 250 year time span is within the calculated error term for the growth rate (Andrews, March, 1967, personal communication) the gap in the recorded measurements is real, though its causes are not known.

An attempt was made to estimate the probability that the three large lichens are part of the same population as the second largest at each of the three sites. Data from the 198 lichen samples used by Andrews and Webber (1964) were examined. Ratios of the second largest to the largest diameter of *R. geographicum* were calculated and are shown in Figure 13. The small ratios are the result of the measurements taken in the youngest environments where maximum diameters were only two or three millimetres. Despite this, the ratio lies within 10 per cent of the maximum in nearly 50 per cent of the samples and within 20 per cent in almost 75 per cent of the cases. Therefore, the probability of the three large lichens being from the same population can be considered low, though no more than a qualitative statement seems appropriate.

What then are the alternative explanations for the largest lichens? There are at least three possibilities. The first is that the microenvironment was peculiarly conducive to growth, for example, the thalli grew in a nutrient-rich location. The second is that the thalli were not in fact single but were compound with a quasicircular outline. Senescence above the 85 mm diameter thalli is a third possibility. In the case of the three thalli under consideration, they all appeared to be in good condition with well-defined perimeters free from imminent coalescence and with no obvious decay at the centre. Assuming a thorough sampling of the area and, if one can, assuming a biological equivalence to the human population age pyramid, senescence is unlikely to be marked by so large a time gap near the peak. Beschel (1961, p. 1047) states that "the first thalli grow in a very scattered manner on a new rock surface with distributions of one plant to a few square meters" but even though such conditions could account for the few large ones they do not explain the gap to the next size range.

Although Sim (1961) took some measurements of *R. geographicum* from both above and below the Generator Lake shoreline it is impossible to assess these data accurately. Sim states that he measured the longest diameter where the thallus was elliptical in shape; this contrasts with the measurements of the shortest diameter in the present study.

The acceptability of a single large lichen diameter measurement much larger than the next size group appears to be a potential problem of the technique, although it was used by Harrison (1964, p. 68) in estimating the age of a moraine in the Bruce Mountains, Baffin Island. The resolution of the problem probably lies with the lichenologist but it is important to recognize its existence at least, especially when approaching the maximum size of a species, as the whole technique is so dependent on the largest (oldest) thallus. The use of ratios of samples of different species can serve as a check though the duration of the uniform growth period is a potentially limiting factor in this respect.

Insufficient data are available to calculate an independent growth rate for *R. geographicum* for this area though useful data on the time necessary for initial establishment of the lichen was obtained. Several single spores up to 1 mm diameter were found in 1966 on the smooth surface of drill cores left by the 1950 Arctic Institute Expedition. An attempt to determine recent growth of an *Alectoria* thallus was partially successful. During the 1950 Arctic Institute Expedition, a photograph was taken which showed quite incidentally, a large *Alectoria* lichen but only incompletely. Fortunately it was in a recognizable position and was found and measured in August, 1965. It

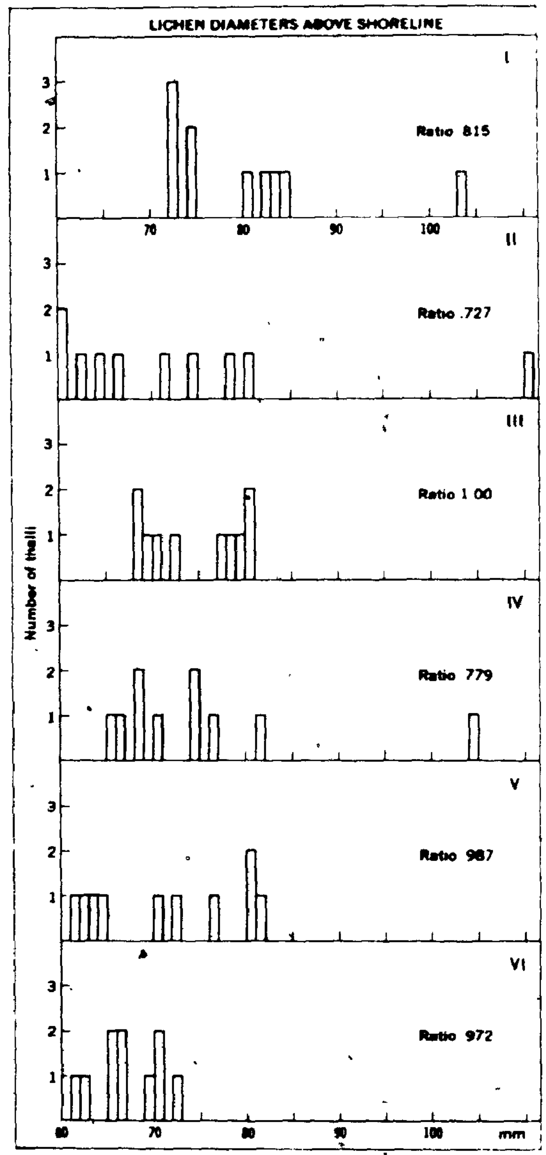
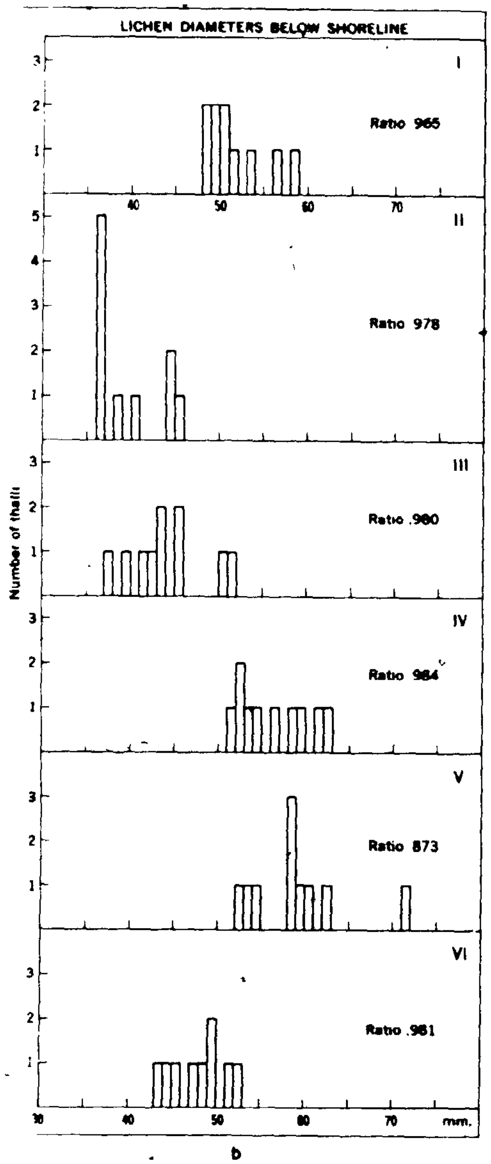


FIGURE 12. Ten largest lichen diameters of *Rhizocarpon geographicum sensu lato* from six paired stations located on Figure 1. Note ratios of second largest to largest.

has grown visibly during the 15 years, though it is not possible to be precise as it was not measured in 1950. The expected increase in diameter would be 6 mm, using the Andrews and Webber (1964) growth rate of 0.4 mm per year — a rate more soundly based than the pioneer estimate by Hale (*see* Ward, 1952). This order of magnitude appears correct and



obviously senescence has not yet set in at a diameter of 82 mm.

Provisional chronology

Allowing then for the reservations outlined above, a provisional chronology can be proposed for the lake from the findings of lichenometry. Thus far no material suitable for Carbon-14 dating has been found. Such a find would offer a check on the lichenometrical chronology, but only of an order of magnitude because of the inherent errors of each method.

By using the modified growth rate of 0.75 mm/year for *R. geographicum* as calculated for the northwest margin of the ice cap (Løken and Andrews, 1966, p. 359), an age and duration of the shoreline can be suggested.

As the present lichen measurements were confined to the western portion of the lake no specific allowance for the difference in time of sub-aerial exposure of the land above the shoreline has been included, as over the distance involved, the variation is probably within the error of the method.

If the large lichens are optimally growing then the lake existed approximately 1,470 years ago, though if the more uniform diameter measurements are used this decreases to 1,120 years. The difference in diameter of the lichens above and below the shoreline can be used to estimate the duration of the lake at its former level. The duration could be as long as 525 years (*R. geographicum* 71 mm) at a maximum, or 350 years if one ignores the largest lichens. When considering such time periods it should be remembered that the shoreline would be subject to water action for only a very small part of the year, perhaps only two months. Even these estimates give an order of magnitude and suggest stability for hundreds rather than tens of years.

Chronology and glacial rebound of the abandoned shoreline

It is believed that the area is still subject to glacial rebound, an assumption supported by the implications of the shape of uplift curves from both the east and west sides of Baffin Island (Løken, 1965; Andrews, 1966). Glacial Lake Lewis at the northwest margin of the Barnes Ice Cap, investigated by Ives (1962), drained approximately 400 years ago (revised date: Andrews *in* Løken and Andrews, 1966, p. 359). Ives (1962, p. 200) concluded that the shoreline was probably "so recent that no isostatic tilting has been able to affect it significantly." The age of the shoreline of Generator Lake was sufficiently greater than that of Lake Lewis to warrant levelling to detect any tilt. A line of levelling 22 km long was run along the more southerly shore of former Generator Lake but it failed to yield evidence for differential uplift. The end points (the extreme cairn

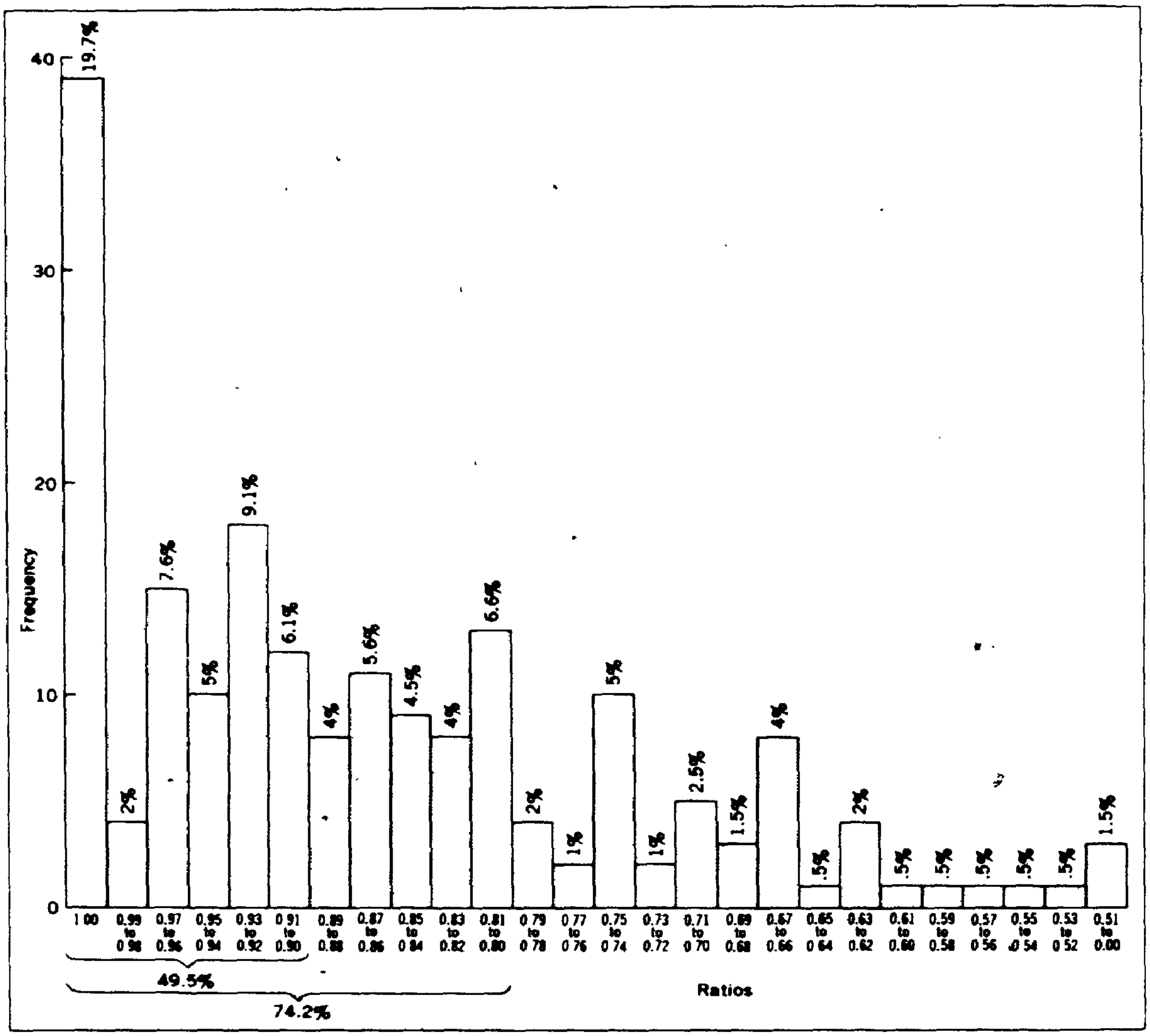


FIGURE 13, Ratios of second largest to largest *Rhizocarpon geographicum sensu lato* from 198 stations of Andrews and Webber's sampling (1964).

locations, Figure 1) had a difference in height of only 22.5 cm, well within the error of the method. The possibility of fortuitously following an isobase was avoided as the traverse has two components 65 degrees apart. Although it is concluded that the shoreline is essentially horizontal this does not preclude uplift but indicates that any which did occur was sufficiently uniform to be undetectable by levelling.

SUMMARY

During the 26-km retreat of the ice margin from the watershed, Generator Lake had two main phases in its evolution, tentatively estimated as lasting 1,500 years. The two bedrock controlled overflow channels were: first into the Sam Ford River and, currently, into the Clyde River.

Asymmetrical cross-valley moraines are the most striking landforms associated with the lake basin. Available stratigraphical evidence suggests that they have a sub-glacial origin, though further work is planned.

The present lake environment is similar to that of the former lake and a similar suite of landforms is present. This situation offers a valuable opportunity for comparative study of landform development in a proglacial environment.

The period 1948-1966 was one of very active retreat of the ice cliffs, averaging 25 metres per year, with consequent equivalent increase in the area of Generator Lake.

Caution should be exercised in using chronology based on lichenometry near the upper size limit of a species. The recognition of the upper limit is difficult but critical if maximum benefit of this useful technique is to be gained.

A tentative chronology is: former lake fully established 1,500 to 1,100 years ago and draining to present level between 950 and 750 years ago. Draining was a rapid process.

The abandoned shoreline is not measurably deformed by glacial rebound.

ACKNOWLEDGMENTS

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Origin, Morphology, and Chronology of Sublacustrine Moraines, Generator Lake, Baffin Island, Northwest Territories, Canada

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The morphology and chronology of exposed sublacustrine moraines is discussed. These are shown to have been formed in a sublacustrine environment. The moraines are characterized by their occurrence in large numbers within a former ice dammed lake basin and by their asymmetric cross section. Using moraine volume estimates combined with the time control, a till depositional rate is calculated.

A radiocarbon chronology for the Generator Lake area over the last 4500 years is presented.

Hydrological relationships are established which are shown to influence the formation of sublacustrine moraines. The relationships are physically connected and are not empirical. Current studies of the calving ice front at Generator Lake show that moraines must be forming under the tapered ice ramps which flow into the lake. Measurements of debris content within the ice combined with ice flow rate measurements show that the present till supply rates are consistent with those calculated from the exposed moraine field.

A model is set up which connects the lake hydrology with the existence or destruction of the ice ramps, as controlled by bending induced by buoyancy forces. The time of existence and subsequent behavior of the ramps is shown to influence the geometry and indirectly the spacing of the sublacustrine moraines.

Le rapport traite de la morphologie et de la chronologie des moraines à présent émergées dont on a démontré qu'elles ont d'abord été formées dans un milieu sous-lacustre. Ces moraines sont caractérisées par leur abondance dans ce qui était autrefois le bassin d'un lac et par leur profil transversal asymétrique. Un taux de déposition de ces moraines peut être calculé en combinant des estimations portant sur le volume des moraines et le facteur temps.

Le rapport présente une chronologie radiocarbon concernant la région du lac Generator pour la période sur les 4500 dernières années.

Il est connu que certaines relations hydrologiques peuvent influencer la formation de moraines sous-lacustres. Les relations sont physiquement liées et ne sont pas empiriques. Les études en cours au lac Generator concernant le vêlage de la falaise de glaces flottantes montrent que les moraines doivent se former sous le pentes glaciaires effillées s'écoulant dans le lac. Des mesures quantitatives portant sur les débris contenus à l'intérieur de la glace, combinées avec des mesures du tau d'écoulement de la glace montrent que les taux actuels de formation des terrains à moraines s'accordent avec ceux calculés pour ces terrains aujourd'hui découverts.

Dans ce rapport, un modèle que relie le caractère hydrologique du lac avec l'existence ou la destruction des pentes glaciaires, selon le fléchissement produit par les forces de flottabilité est établi. La durée d'existence et le comportement subséquent des pentes semblent influencer la gémométrie et indirectement l'espacement des moraines sous-lacustres.

1. Introduction

Generator Lake (69°35′ N; 71°52′ W; 403 m asl) is a proglacial lake abutting the southeastern margin of the Barnes Ice Cap (Fig. 1) in north-central Baffin Island. The ice is thought to be a Wisconsin relic (Goldthwait 1951) and lies on a dissected plateau surface about 500 m above sea level. Goldthwait (1951) first re-

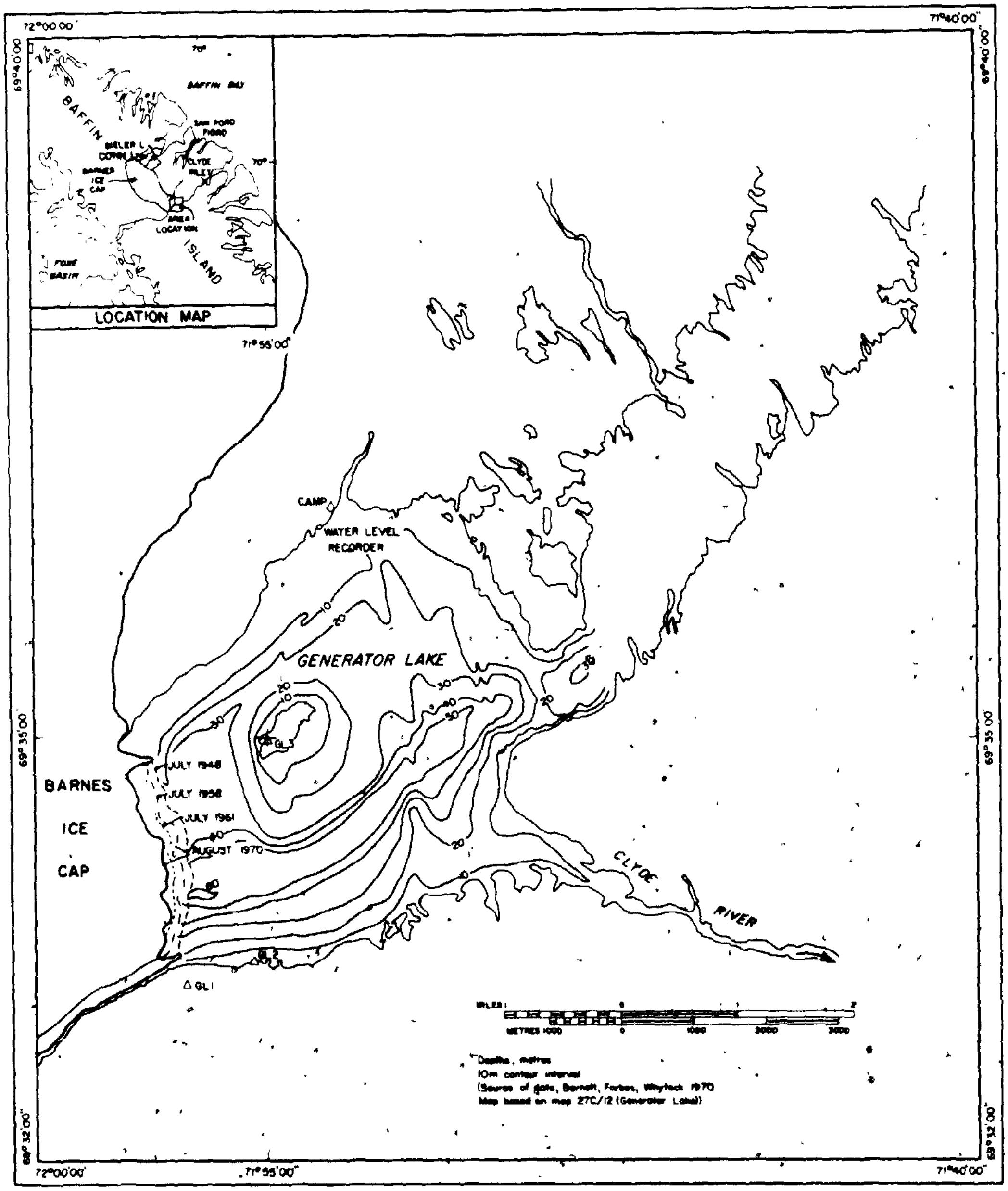

FIG 1. Map of the Generator Lake area.

ported "a succession of underwater sand and gravel end moraines" which he observed both above and continuing below the then current level of Generator Lake. Andrews (1963, Fig. 2) identified the same moraines from air photographs as being similar to features he termed cross-valley moraines. Barnett (1967) continued this terminology in a preliminary study of the development, landforms and chronology of Generator Lake, concluding that the moraines were restricted to the lake basin below the level of the highest shoreline (p. 178). The present study uses the term *sublacustrine* moraines as it indicates the environment of depo-

FIG. 2. View along the crest of a prominent moraine showing bouldery surface. Distal slope is to the left of the crest. (GSC photo 161626).

sition, a vital aspect of understanding the origin of these landforms.

The present paper reports work carried out independently but with some degree of overlap. The chronology and morphology aspects were primarily investigated by Barnett; the glaciological and morphometric aspects primarily by Holdsworth with complementary contributions on till fabrics (P. McLaren) hydrologic data, and meteorological data.

1.1 The Regional Setting

The mountainous eastern rim of Baffin Island was breached by glaciers which cut deep fiords extending westward of the height of land. Consequently the main drainage divide is now located approximately along the geographical axis of the island, 144 km of which is formed by the crest line of the Barnes Ice Cap. The ice has disrupted preglacial drainage lines and locally impounds proglacial lakes including the two largest existing in North America—Conn and Bieler Lakes (Fig. 1, *inset*). Other proglacial lakes existed when the Barnes Ice Cap was only slightly larger, particularly around the northwest margin (Ives and Andrews 1963). An earlier phase of Generator Lake outlined by Barnett (1967, Fig. 1, p. 171) also related to a slightly larger ice cap, when lake level was 76 ± 1 m above current levels and the outflow spilled across the main divide at the northeast end of the drainage basin (Fig. 1). Subsequent drainage has been *via* the Clyde River valley across a low point along the south side of the basin (Fig. 1).

It is now recognized that an interval of several centuries occurred during which the continued retreat of the ice cap margin caused Generator Lake to drain completely before it was reestablished by a recent surge of part of the South Dome, Barnes Ice Cap (Løken 1969), thus filling up the basin to the Clyde River outlet once again.

A provisional chronology of the lake based on lichenometry was given by Barnett (1967, pp. 182–185) who also outlined the probable shortcomings of this method. Section 3 gives a revised chronology based on radiocarbon dates, which should be judged collectively rather than individually.

2. Morphology of the Moraines

2.1 General

The dominant feature of the sublacustrine moraines, when considered singly, is their

FIG. 3. Aerial photograph (21 500 ft, 6 in. lens) of Generator Lake area showing locations of features referred to in the text. (*1*) Former island, (*2*) 'Embayment' feature in the moraine, (*3*) 'Headland' feature in the moraine, (*4*) Delta and beach 76 m above present lake level, (*5*) Delta at 55 m above present lake level, (*6*) Barnes Ice Cap (Photo: National Air Photo Library, Ottawa, A-15898-45).

asymmetry in cross section with steeper distal slopes ranging up to the angle of rest of the material (Fig. 2; Barnett 1967, p. 178). Collectively the striking features are their number and proximity to one another (Fig. 3). They occur in subparallel groups, are gently arcuate or sinuous and are characterized by greatest size in the central valley position (former deepest water) declining to nothing at, or just short of, the level of the highest shoreline. The longest continuous single ridge is 6 km and the greatest height measured is 35 m. Typically a moraine surface is dominated by boulders. At two locations it was possible to examine the internal composition of the moraines; no obvious structures were visible. Boulders remain

	SECTION WEST	CENTRAL	EAST
[illegible]	33	34	40
[illegible] (m) [illegible]	36	23	20
[illegible]	4	1	1
[illegible]	13	8	9
[illegible] (°) [illegible]	28	27	27
[illegible]	15	4	2
[illegible]	34	35	36
[illegible]	12.5	6	5
[illegible]	9	8	7
[illegible]	26	23	26
[illegible]	21.5	29	32
[illegible] > distal	2	7	4

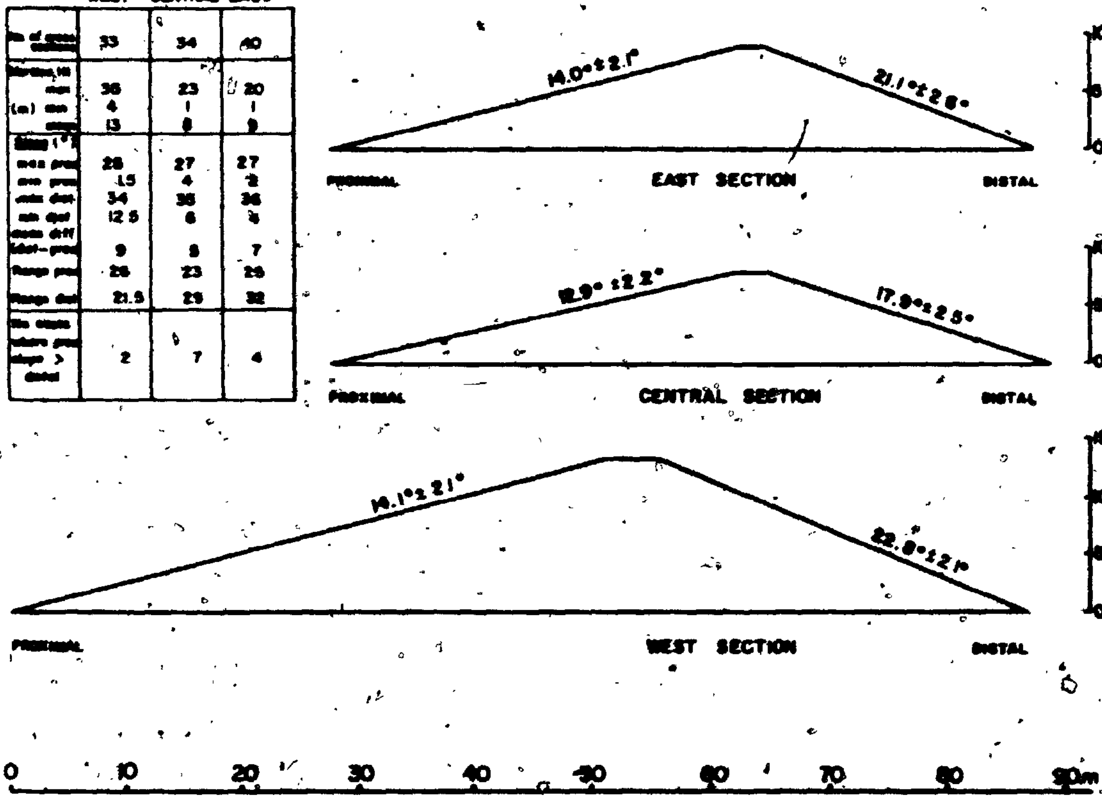

FIG. 4. Average moraine cross sections and slope data for east, central, and western sections (see Fig. 3).

common throughout although more coarse sand and gravel-size material was present at depth than at the surface.

2.2 *Cross Sections*

A total of 110 hand-levelled profiles show that the mean distal slope angle exceeds the proximal by about 7° (Fig. 4). The mean cross-section values are based on the *dominant*, measured slope angle of a particular moraine, if one or more breaks of slope was recognized. In only 13 cases were the proximal slope angles in excess of the distal. This characteristic asymmetry must reflect primary processes as it persists on moraines currently subject to subaerial and lacustrine processes. In the latter case morphological modification is limited to occasional ice-pushed ridges up to two m high. Areas subject to such ice action are characterized by boulders showing lichen-free faces of lighter tone than the generally darker and therefore more stable slopes above the shore zone. The typical proximal slope outlined is compatible with the slope of the underside of the ice ramps discussed in section 4. The steepest distal slope angles are compatible with angles resulting from a free fall of cobbles and boulders. Those distal slopes having angles less than the angle of rest, *may* be due to scraping by calving icebergs or subsequent subaerial flattening of the slopes. Alternatively there is some other mechanism operating, but it is not a dominant one.

2.3 *Plan View*

A general alignment of moraines perpendicular to the long axis of the lake basin (or valley) is apparent (Fig. 3), particularly in the east-central part of Fig. 3. Locally, variations occur, as at sites 2 and 3 on Fig. 3. At site 2 evidence of an 'embayment' occurs and at 3 a 'headland' to use a shoreline analogy. The embayment is close to a former island (1 on Fig. 3). At site 3 there is an indication of one moraine overriding the proximal slope of another. This represents a local advance of the ice margin which according to Table 2 may have lasted more than ten years, *i.e.* the mean time for the formation of one moraine.

2.4 *Moraine Morphology Adjacent to the Clyde River Outlet*

Special attention to this zone is appropriate in view of the chronology outlined in section 3, which indicates the ice margin may have remained in the vicinity for a period in excess of 1000 yrs. It is reasonable to expect fluctuation of the ice margin to have occurred in this position (section 7) and that the till deposited would have accumulated locally in greater amounts in this zone than further eastward in the lake basin. Evidence of readvance is present particularly on the south shore where some moraine segments have been pushed at approximately 45° to the axis of the valley, and as many as three short sections run together. Other overridden segments are present on the north side of the present lake. Nevertheless it is clear that these features indicate short distance readvances, as there is no evidence of one moraine completely overriding the next most easterly. This would seem to indicate pulses of faster moving ice within the ice front with perhaps only a slight readvance across the whole ice front.

Evidence of till in this zone occurring in larger amounts than farther east is present but not overwhelming. Examination of Profile 7 (Barnett *et al.* 1970, Fig. 2) shows that the general lake bottom profile has a substantial hump in the zone of the present lake constriction (Fig. 10). However, a correction of ~10 m should be applied to this part of the profile, reducing the apparent bottom relief because this part of the profile line is upslope from the deepest water at the lake constriction. An apparent excess of material still remains and this could be explained by (*a*) an accumulation of till, (*b*) a bedrock rise or ~~(*c*) a combination of~~ the two. If it is till then this would offer strong support to the proposed chronology (lake phase II, section 3.2). Above present lake level where moraine ridges occur close together there is a thicker accumulation of till between them rather than just till veneer over bedrock. A further strong indicator of ice margin behavior is the presence of a subaerial moraine above the 76 ± 1 m shoreline and shown as 'ice marginal moraine' on Fig. 5 (note that this was ice marginal over 1000 yrs ago, section 3.2). This must indicate a halt or slight advance of the land-based margin just at the critical time when drainage down the Clyde River was imminent or partially underway. It is the only such moraine occurring between the present ice cap marginal ice-cored moraine and the 76 ± 1 m shoreline. This qualitative evidence supports the chronology, which, when taken in conjunction with the magnitude of the five deltas discussed in section 3.2, provides strong indication for a significant time interval for the ice margin in this vicinity.

2.5 *Till Fabrics*

Pebble orientations were measured at 13 sites, to study the structure of elongated pebbles within the moraines. All pebbles had a measured minimum $a:b$ axis ratio of 2:1, and all measurements were taken below local frost table as far into the moraine as possible to minimize the possibility of measuring pebbles reoriented by freeze–thaw action. This potential problem was probably not serious as the coarseness of the material is such that little moisture is retained, particularly close to the moraine crests, which tend to be kept snow-free by wind.

Fabrics (Figs. 6, 7, and 8) indicate preferential orientation of the *a* axis for both azimuth and dip, with generally stronger orientation on the proximal side (Fig. 7). Pebbles are preferentially oriented at right angles to the long axis of the moraines and must be approximately parallel with the former direction of ice flow. Preferred dip angles are clustered at less than 35°. In the case of the paired (distal and proximal) fabrics at locations 6, 7, and 8 (Fig. 5) when grouped by 20° classes, the strongest modal class occurs on the proximal side at location 7 (Fig. 8) where 23 of the 50 pebbles dip up glacier at between 1° and 20°. Similarly the modal class for the proximal slope at locations 6 and 8 is in an up glacier attitude between 21° and 40°, although at location 8 an equal number (12) are between 1° and 20°.

Examination of the modal 20° classes for the

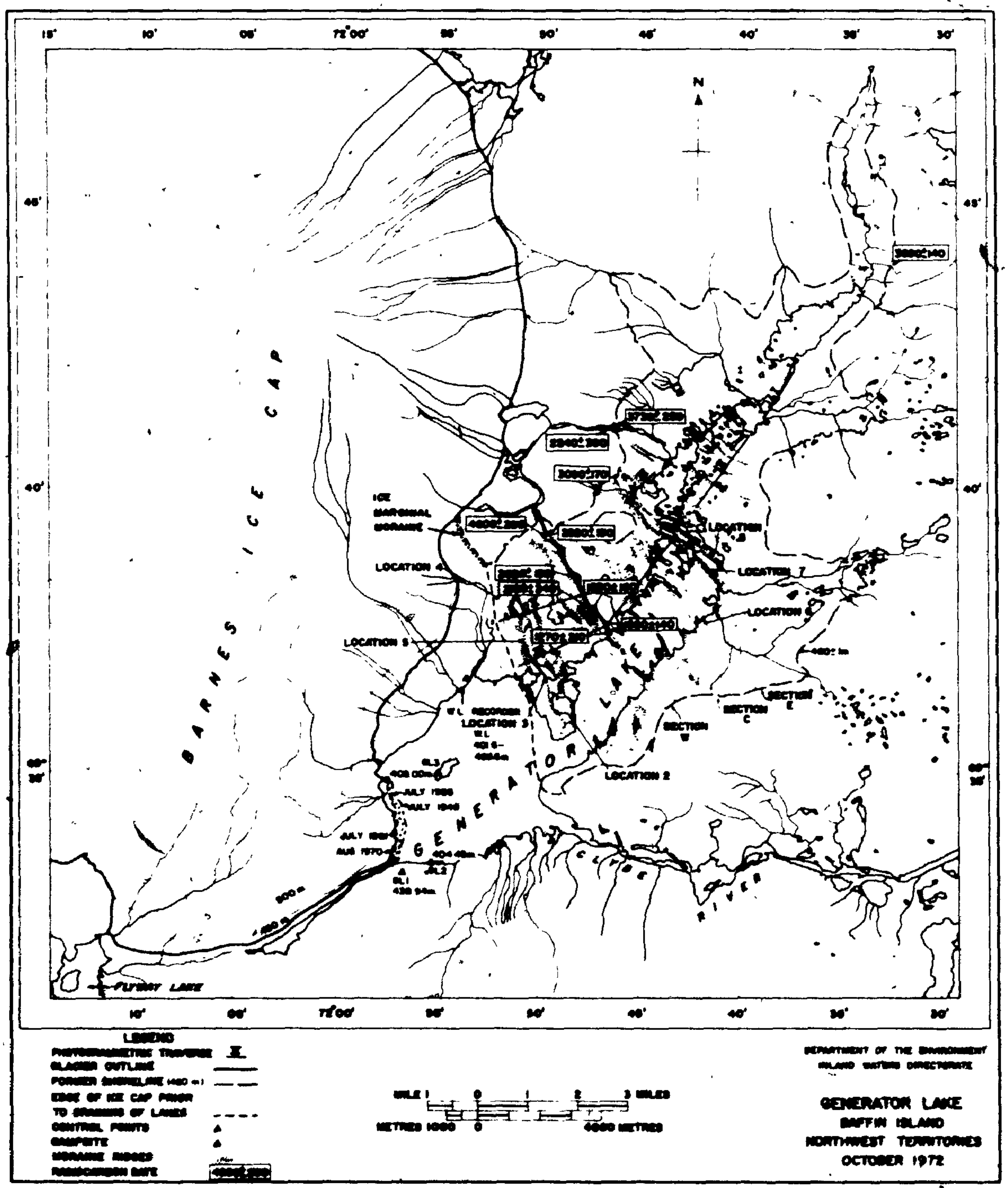

FIG. 5. Map of the Generator Lake area showing locations of traverses, sections, pits, radiocarbon samples, and control points referred to in text.

distal fabrics reveals a down glacier dip between 1° and 20° for locations 6 and 8 and between 21° and 40° at location 7. Secondary modes are quite strong at 6 and 8 with up glacier dips between 1° and 20°. A small tertiary mode is visible at 7 with an up glacier dip between 1° and 20°.

This contrast suggests differences in depo-

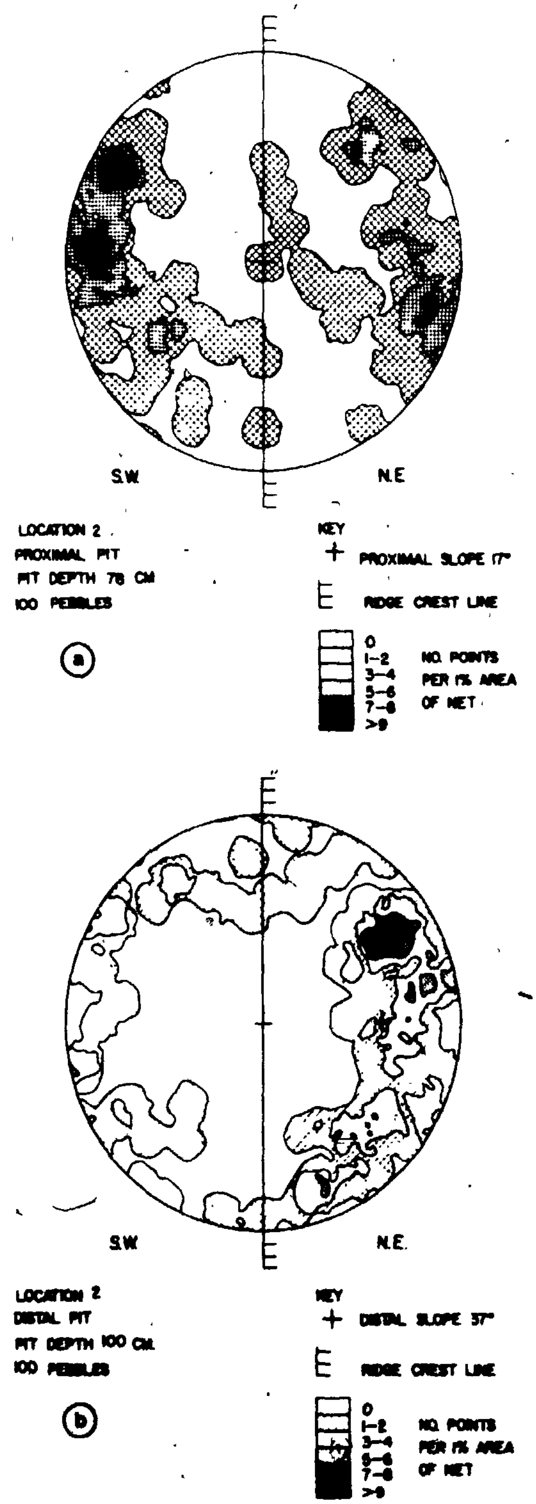

FIG. 6. Fabric diagrams for location 2, (*a*) proximal pit, (*b*) distal pit.

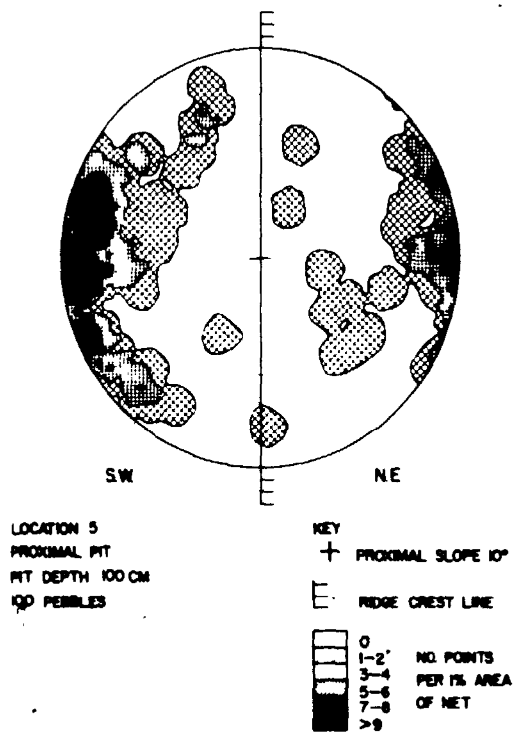

FIG. 7. Fabric diagram for location 5.

sitional processes on the two sides of the moraine. The dominant down glacier dip on the distal side taken in conjunction with the strong secondary mode in the up glacier attitude is difficult to explain. It may be noted that Andrews and Smithson (1966) infrequently recorded such a dip distribution pattern (their Fig. 6, pattern 6). It should be emphasized, however, that they analyzed 103 fabrics and so no further comment is made on the basis of our smaller sample. An important observation made during excavation of pits was the assessment that proximal slope material was more highly compacted than distal slope material. Data obtained by J. T. Andrews (personal communication, 1973) in the Isortoq valley moraines supports this observation. This result is consistent with the model of section 6.

3. Chronology

3.1 Radiocarbon Samples

The map of the retreat of the Wisconsin and Recent Ice in North America (Prest 1969) shows a closed 6000 yr isochrone, surrounding the Barnes Ice Cap; the isochrone encom-

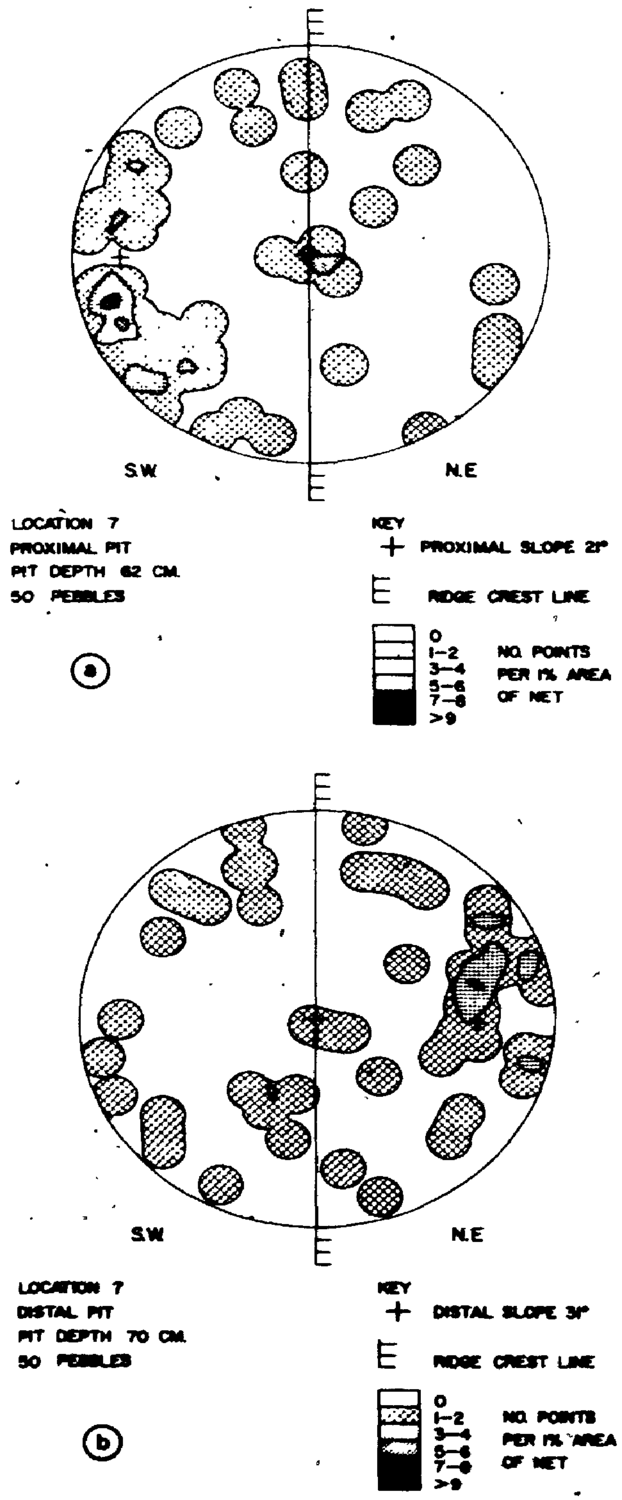

FIG. 8. Fabric diagram for location 7, (*a*) proximal pit, (*b*) distal pit.

passes the area of Generator Lake both past and present. Eleven samples of plant remains from the area have been dated by the Geological Survey of Canada, Radiocarbon Laboratory (GSC); all yielded dates which are significantly younger than 6000 yrs, and are listed in Table 1. Sample locations are shown on Fig. 5. For such a remote area this represents a high density of dates necessitating careful evaluation. All samples consisted of detrital vegetation deposited in deltas above current lake level. The lake basin is very sparsely vegetated with scattered lichens, mosses, and few vascular plants. Organic matter in the sandy foreset beds of deltas is therefore not abundant and in some cases only a few grams were located and in others none at all. All samples were carefully collected from sites free from surface vegetation and associated roots and from sites which showed no visible sign of slumping. Only one sample (GSC 1087) was located southeast of the lake. Seven samples (GSC 1168, 1177, 1239, 1304, 1315, 1325, 1622) were from one flight of five deltas along the same stream. The stream is only 7 km long from ice margin to the present lake shoreline.

3.2 *Lake Phases*

There were three distinct phases in the development of the lake and the associated landforms: First, the evolution of the former lake (76 ± 1 m above the present one) gradually increasing in size by a net retreat (26 km) of the ice margin to the southwest and then west; the initiation of this phase is thought to date at about 4500–4000 B.P. This estimate is not based on closely bracketing dates, but on radiocarbon dates from Sam Ford Fiord and Clyde Inlet (Andrews and Drapier 1967). Second, a five interval adjustment of lake level from 76 ± 1 m to present level lasting between 1000 and 1500 yrs. This implies that the ice margin was in the vicinity of the Clyde River outlet for this length of time. Third, a continuing *net* westward retreat (6.5 km) of the ice margin, a phase which has probably been complicated by complete drainage and refilling of the lake basin. The refilling followed a surge of the Barnes Ice Cap which occurred before 1948 and probably early this century (Holdsworth 1973*b*). It is not possible at present to determine exactly when the drain-

TABLE 1. Radiocarbon data (from Geological Survey of Canada, Ottawa)

GSC sample no.	^{14}C Date B.P. (yr)	$^{13}C/^{12}C$ Corrected date B.P. (yr)	Range: 95% Confidence limits (yr)	Lake level above present lake level (m)
GSC 1087	3650 ± 140	3690 ± 140	3830–3550	76
GSC 1244	3690 ± 250	3730 ± 250	3980–3480	76
GSC 1621	2240 ± 390	N/A	2630–1850	76
GSC 1276	3080 ± 170	3090 ± 170	3260–2920	76
GSC 1168	4600 ± 290	N/A	4890–4310	76
GSC 1304	2480 ± 150	2520 ± 150	2670–2370	76
GSC 1315	2600 ± 150	2620 ± 150	2770–2470	55
GSC 1622	2180 ± 240	N/A	2420–1940	23
GSC 1325	1480 ± 160	1530 ± 160	1690–1370	23
GSC 1177	1560 ± 140	1660 ± 140	1800–1520	12
GSC 1239	1240 ± 210	1270 ± 210	1480–1060	3

NOTES. 1. All samples were small fragments of plant detritus, dominantly mosses.
2. Corrections for GSC 1168, 1621, and 1622 unavailable due to laboratory problems.

age took place as evidence would probably lie outside the field area. Lake bottom evidence for the draining can be advanced on the basis that repeated drops of a Phleger gravity corer at several sites yielded no more than a few centimeters of sediment and several bent cutting rims. Such sediment depths on bedrock are not compatible with a thousand yrs of deposition in the proglacial environment. This evidence would seem to indicate that sediment had been flushed out of the lake on draining and/or that it had been removed by wind action after being exposed subaerially. This complete phase has already occupied more than 1000 yrs and may approach 1500 yrs in length. Unfortunately no direct chronological control is available from the end of phase II until 1948, the year of the first air photography. Good control is available since that time.

The upper shoreline (76 m above present lake level) yielded six datable samples from four different deltas. Sample GSC 1276 gave a date of 3090 ± 170 B.P. (Fig. 5). The stream which created this delta was ice marginal but soon changed course to a channel further west. Therefore the organic material within the delta was deposited before any of the deltas to the west were formed, unless one invokes a marginal fluctuation for which there is no geomorphic evidence. The flight of five deltas immediately west of it (on the northeast edge of section W in Fig. 5) were created by the same stream after its course changed as the ice margin withdrew. Of the three dates available northeast of GSC 1276 (3090 ± 170 B.P.), two are older and of the seven to the southwest, six are younger (Table 1) as is compatible with the already proposed geomorphic evolution of the lake and associated landforms (Barnett 1967). The two dates which are not obviously compatible with this sequence are GSC 1168 (4600 ± 290) and GSC 1621 (2240 ± 390), for which we do not suspect human errors such as mislabelling, sample switching or laboratory errors to exist. However, both samples were minute (4.5 g and 1.2 g, respectively), an undesirable characteristic for accurate dating.

Contamination of the samples with material of a different age remains a possibility. *Betula* pollen has been identified by R. J. Mott (GSC) both in deltaic samples and lake bottom samples. No exotic macrofossils have been identified and no living specimens of *Betula* have been found in the area. The nearest known occurrence of fossil *Betula* pollen is in the Isortoq plant-bearing beds 200 km to the west-northwest (Terasmae et al. 1966). The concentration of pollen grains, though small, exceed expected airborne values. However, one of the two dates is older and the other younger than the 'preferred' chronology.

Four deltas at 55, 23, 12, and 3 m above present lake level occur between the uppermost shoreline and the present active delta. All deltas have yielded datable material (Table 1) ranging from 2620 ± 150 (GSC 1315) at the top to 1270 ± 210 (GSC 1239) at the bottom. This second period in the chronology documented from the flight of deltas reflects a

TABLE 2. Chronology, moraine formation, and retreat rates

Lake phase	Time Interval (yr)	Number of moraines	Axial retreat distance (km)	Retreat rate (myr^{-1})	Moraines per 100 yr	Average number of years per single moraine	Remarks
	4500[1]–3850 B.P.	—	7.2	10.7	—	—	Moraines nonexistent
	4000[1]			42.4			to very small
	3830–3480	75±10	9.2	26.3	21±3	5±0.6	small to medium
	3480–3090	17±2	2.6	6.7	4±1	23±2	medium to very large
	3090–2520	40±2	7.8	13.7	7±1	14±1	medium to large
I	3090–2820	24±2	3.7	13.7	9±1	11±1	medium
	2820[2]–2520	16±1	4.1	13.7	5±1	19±1	medium to very large
	3830–2520	132±15	19.6	15.0	10±1	10±1	full range
II	2520–1270 B.P. (5 phase adjustment to 3 m level; may have lasted until 1060)						deposition confined to area of lake constriction (Fig. 11)
III	1060[3](B.P.) } 1948 (A.D.)	3(?)	6.0	5.7	—	—	nonexistent to very small
	1948–1971(A.D.)	3±2	0.4	17.4	13±8	8±3	very small

[1]Initiation assumed; not definitely dated but 4000 B.P. is considered more reasonable.
[2]Interpolated date.
[3]Mixed chronology (see section 3.2). Phase includes lake draining and subsequent surge.

basically unstable environment. The summer ice marginal outflow from Generator Lake drained down the Clyde valley thereby lowering the lake level. Temporarily, stable water levels occurred at 55, 23, 12, and 3 m above the present shoreline. It is possible that lake level dropped to a level approaching the present one on several occasions during this phase. If it did, the relative time spent at the low level was considerably shorter than the period at each of the delta levels because of instability of the ice margin (section 7; Appendix B), and the size of the deltas. As a result, lake level would probably have followed a cyclic pattern of a rapid drop followed by a gradual build-up to a new stable level marked by the elevated deltas. These are morphologically distinct indicating a stable water level and therefore requiring a slight readvance of the ice margin on each of four occasions thereby partially sealing off Clyde River each time. The volume of sand in the delta 12 m above present lake level suggests that water level remained longest at that level, if we assume similar stream discharge and load throughout the adjustment period. Elapsed time for the total oscillation phase suggests approximately two centuries per level.

There is no chronological control currently available for the evolution of the present lake between the 3 m adjustment phase and the 1948 aerial photography, a time span of 1270 ± 210 radiocarbon years. An unimposing moraine, both above and below the lake surface some 3 km east of the ice cliffs (Barnett *et al.* 1970, Fig. 4), offers some morphological evidence for a pause or readvance of unknown duration in the retreat of the ice margin during this period. After the formation of this moraine the ice margin probably retreated sufficiently to drain the lake completely, thus establishing fluvial drainage to Foxe Basin. Later a surge of the Barnes Ice Cap advanced ice across the valley and Generator Lake reestablished itself to the level of the Clyde River outlet which is the current one.

It should be noted that the lake chronology is based on radiocarbon years which may differ from calendar years (on which the glaciological calculations are based) by up to some hundreds of years (Suess 1970), but the assumptions involved in both studies are such that any attempt at reconciliation between them is inappropriate at this level of investigation. The possible divergence of the two time scales does not, in principle, invalidate interpolation between dates from the different scales but simply amounts to a small additional uncertainty concerning the derived retreat rates (Table 2).

Reliance has been placed on the radiocarbon dates without specific reference to the ten-

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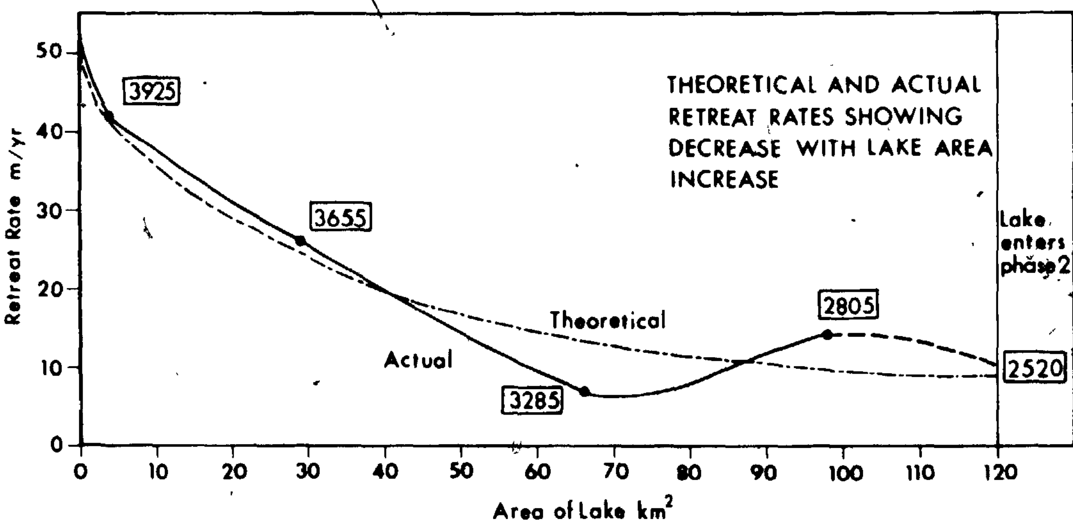

FIG. 9. Plot of ice retreat rates for phase 1 of former lake *versus* lake area. Full line calculated from ^{14}C dates (in blocks). Dashed line, theoretical relationship (see section 4).

tative chronology based on lichenometry already discussed by Barnett (1967). The two chronologies do not coincide but the *relative* chronology derived from lichenometry is not at variance with that proposed here. The significance of the radiocarbon dates for growth rates of lichens forms the basis for a further note in preparation.

3.3 Retreat Rates

The average net retreat rate for the present lake basin west of Clyde River is about 6 m yr^{-1} and post-1948 rates for the north half of the calving ice front have exceeded this figure by a factor of four. This recent trend is associated with the intersection of about one half of the calving ice front by the heavily crevassed zone which is visible on the 1948 air photographs and which remains fixed relative to bedrock. The 6 m yr^{-1} apparent retreat rate is lower than the complete basin average of 8 m yr^{-1} but when allowance is made for the surge it can be seen that actual retreat rates must have been much higher.

The mean retreat rate for the 76 m water level basin was 19 m yr^{-1} based on a distance of 26.8 km along the axis of the basin and lake inception at 4000 B. P. and initial marginal drainage at 2600 B. P. The mean decreases to 13.4 m yr^{-1} if inception was as early as 4500 B. P. and drainage as late as 2500 B. P. A graphical plot (Fig. 9) shows a general decline in retreat rate with increasing lake area, indicating more stable conditions tend to prevail as the lake increases in area. The theoretical aspects of this are discussed in Section 6.

Table 2 summarizes net retreat rates based on the radiocarbon dates. Some judgement has been used in deciding on reasonable probabilities within the statistical error terms of each date. From this outline the number of moraines formed per unit time may be calculated and compared with the apparent speed of glaciological processes. However, the number of moraines is not as significant as the volume of till deposited per unit time.

3.4 Till Volume and Time

Table 3 shows the amount of till apparently deposited annually from an ice column one m wide, and at least 100 m thick (Fig. 10) calculated for three different sections (Fig. 5) in the former lake basin. The data are dependent on the following factors: (*1*) The calculated average retreat rates apply over a strip 500 m wide, (*2*) mean cross sectional areas of moraines are based on hand-levelled profiles (minimum 33 profiles) at 200 m intervals across the lake basin (parts of section C at 100 m intervals), (*3*) a planar base is assumed between the end points of moraine profiles, and (*4*) we assume no significant amount of till between individual moraines. This latter approximation is reasonable as bedrock fre-

TABLE 3. Till release at three cross sections in the former lake

	Volume (m^3) per m width	No. of profiles measured	Retreat rate myr^{-1} (see Table 2)	Time (yr) for 500 m retreat	Till released m^3 m^{-1} (width) yr^{-1}	Equiv. uniform depth till (m)
West section:	930.5	33	13.7	36.5	25.5	1.861
Central section:	310.1	35	13.7	36.5	8.5	0.620
East section:	514.4	42	6.7	74.6	6.9	1.029

NOTE: Location of sections shown in Fig. 5.

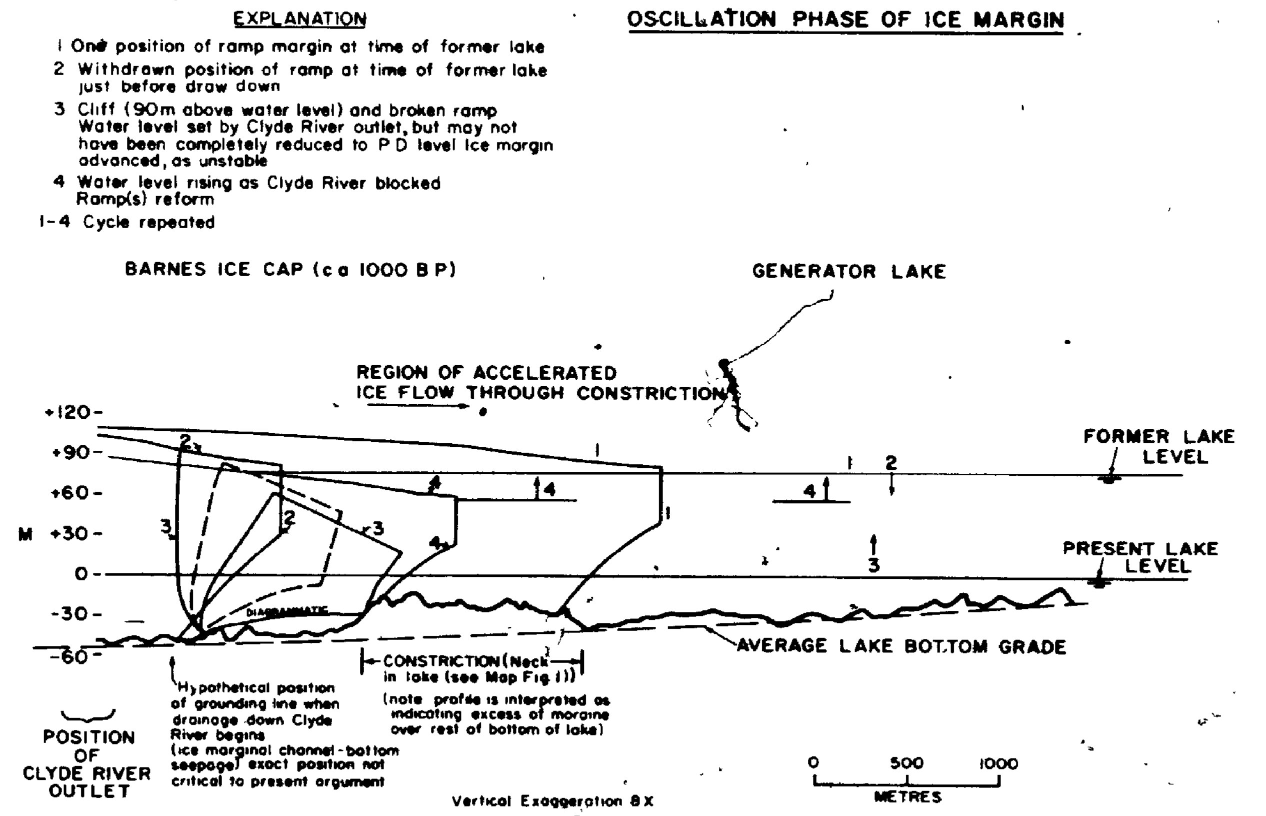

FIG. 10. Diagram illustrating phase 2 of the lake history. Ice cap profile is a reconstruction based on present profiles.

quently crops out between moraines. All three profiles are partially submerged in the central sections. Sublacustrine morphology was estimated from cross valley profiles between ridges, lake sounding where available, and the longitudinal valley profile. The values derived are minimum estimates as no allowance for subaerial erosion could be made.

The section of smallest moraines and the narrowest lake width (central section C) yielded the smallest depositional rate. The western (W) section gave the highest depositional rate and has the most persistent well developed moraines, and the biggest single cross section measured. The eastern (E) section is the longest and shows the greatest variation in form between no measurable till at one close-to-shoreline profile and a considerable till volume in one cross section nearing the largest measured in any section. The central and eastern sections yield similar till release rates, but the western section yields almost three times as much till per unit width.

If the available till is spread uniformly

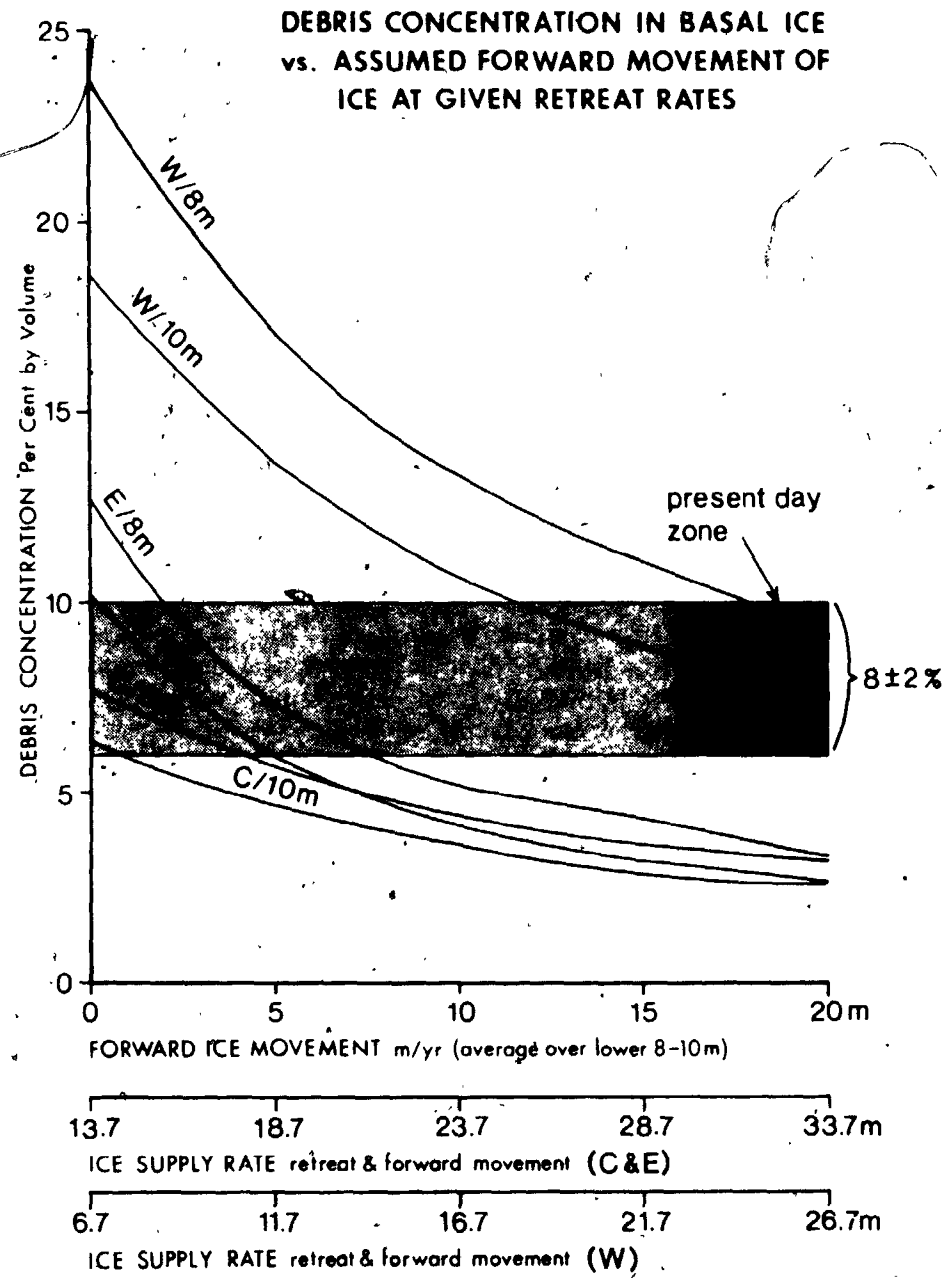

FIG. 11. Plot of debris concentration (% by volume) in the basal 8 or 10 m of ice *versus* forward ice movement averaged over the lower 8–10 m of ice. For the sections C, E and W, the ice supply rate is the sum of the average retreat rates (Table 2) and the forward ice flow rate.

within the 500 m wide areas of the sections examined the resulting thickness is quite variable; from 0.62 m in the central section to 1.86 m in the west, with an average 1.18 m (Table 3). The object is to obtain the order of magnitude of till deposition rates for comparison with rates derived from glaciological considerations (section 5.7). Figure 11 shows debris concentration within the basal ice for assumed limiting basal debris layer thicknesses of 8 m

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and 10 m, plotted against ice flow rates from 'stagnant' (a hypothetical limiting value) to 20 m yr^{-1} forward movement. The data on which this figure is based comes from measurements of till content in sections E, C, and W. Superimposed on these curves are the measured limiting values for debris concentration (section 5.5). The shaded zone intersects all curves; but in the case of the east and central areas the intersection is at ice movement rates generally less than 5 m yr^{-1} and in the west at rates in the 15–20 m yr^{-1} range. Such rates are compatible with *maximum* present day values (heavy shaded zone, Fig. 11).

As the rates of both till release and ice movement (based on the present day measured values) for the western section are generally higher than for the other two sections, it is desirable to examine the results in more detail. One critical factor is the retreat rate which influences many of the other values and which is based on an assessment of the error terms of radiocarbon dates. Therefore if we assume the *maximum* till release rate of 16 m^3 per m width per yr (section 5.7), calculated from present day glaciological data, a retreat rate of 8.6 m yr^{-1} is indicated. This rate, when applied between the locations of radiocarbon sample GSC 1304 and GSC 1276 requires a 900 yr interval, which lies within the error terms of each value (±115; Table 1). The agreement on till release rate values obtained by different methods is then well within an order of magnitude.

4. The Former Lake and Moraine System

4.1 Lake Hydrology

Hydrological conditions that existed in the former lake were probably very similar to the conditions in the present lake, which are now described.

Generator Lake is ice covered for most of the year. During the winter the lake ice, which is about 1.7 ± 0.1 m thick, is covered by 0.4–1 m snow. Measurements of water level, relative to a shoreline bench mark, in several holes drilled in the ice in the spring (May 1970, 1971) revealed water level to be the same within the limits of the measurement (±1 mm). This is interpreted to mean that the lake level reaches a minimum value during the winter and that this level is essentially the same each year. Beginning usually in the second part of June the surface snowpack and the lake ice begin to melt rapidly; especially at the edges of the lake where a lead or *moat* develops. By this time, runoff from land-based snowmelt and the ice cap produces an influx of water via streams flowing through the end moraines, over the ice cliffs and by way of two or three moulins. As a result, there is a rapid rise in lake level over a period of about 10 to 20 days.

The lake level reaches a peak in July, when the outflow via Clyde River is increasing after the channel becomes ice free. From this time on, depending on temperature and insolation the lake level may continuously drop or fluctuate (Fig. 12) in response to periodic higher runoff.

Runoff from the ice cap is strongly influenced by local climatic conditions (Anonymous 1967). Appendix A gives a summary of the meteorological measurements made from 1968 to 1971, together with observations of the condition of the lake ice. The data tend to show a broad relationship between the daily meteorological conditions and the shape of the water level curve.

By early August, major cracks develop in the ice cover which is then about 1 m ± 0.1 m thick. Depending on the degree of crack development and windiness the ice may either continuously shift position with minor amounts being lost down the river, or else, the whole ice cover will disintegrate and flow down the river (as it did by late August in 1966, 1969, and 1971). The only ice that may remain are small icebergs which become grounded. Formation of new lake ice begins in September.

Figure 12 shows that the maximum summer water level rise is about 1.8 ± 0.3 m covering the years 1968–71.

4.2 Area–Rise Relationship for the Present and Former Lake

We next consider the input water volume from spring snow melt, which results in a rapid rise in lake level. The amount of glacial ice melt involved at this time is minimal. Strictly, as the former lake grew in area (corresponding to a retreating ice margin), the number of streams flowing into the lake increased. It is

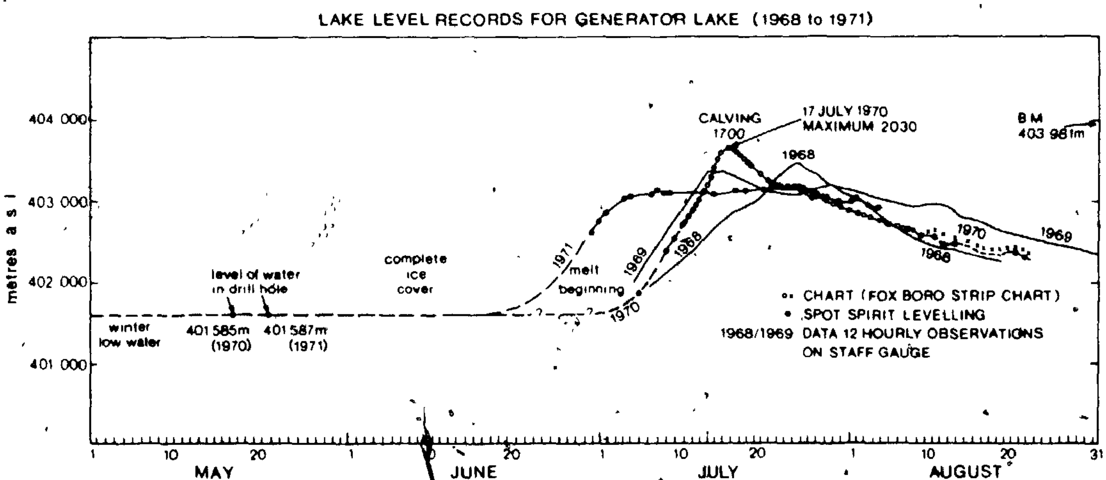

FIG. 12. Lake level records for Generator Lake, 1968–71.

judged from the topographic maps that after the lake area reached about 15–20 km², the drainage *volume* into the lake probably did not increase significantly, other things being equal, because, the change was, essentially, in the drainage *pattern*. In the later calculations this same assumption is held up to the present. We have, however, ignored the small increased area of southerly hillside snow, the melt of which has been available to the present lake, but not the former lake.

We have assumed that the reduction in area of the ice cap with time has not significantly altered the amount of snow available for the spring melt. With time, more snow lay on deglaciated ground, rather than the ice cap.

A certain estimated percentage ($<50\%$) of the runoff volume contributes each summer to raising the lake level, from the winter low level to the maximum (July) level. We now consider, under the limitations just outlined, that on a long term mean basis, this water volume was essentially constant (equal to an *average* value) for $15 \text{ km}^2 < \bar{A}$ (lake area) $< 125 \text{ km}^2$ which includes values for both former and present lakes. Because we will consider in section 7 the behavior of the future lake, it should be mentioned here that there is a physical restriction on the value of the upper limit of $\bar{A}$, for the *present* lake, of about $47 \pm 1 \text{ km}^2$, before drainage into Flyway Lake is likely to occur. If z_w is the maximum summer water level rise and $\bar{A}$ is the average area of the lake then

$$[4a] \qquad z_w \cdot \bar{A} = C$$

where C is a constant; an average value by hypothesis. An estimate of the magnitude of C may be obtained using the values of z_w and $\bar{A}$ for the present lake, from which $C = 1.8 (\pm 0.3) \text{ m} \times 40 (\pm 1) \text{ km}^2 = 72 (\pm 14) \cdot 10^6 \text{ m}^3$. Using this result and Eq. [4*a*] a hyperbola is plotted (Fig. 13) from which it is possible to predict, in principle, a value of z_w if $\bar{A}$ is known. C is *not* the total volume of water flowing into the lake each summer, but probably less than half of the total. It is necessary to consider the lake hydrology in these average terms because it simplifies the arguments that follow. Even if some of the assumptions made in this section are not strictly true this does not invalidate the general principles which are developed in the following sections.

4.3 Moraine Morphometry

Since it is the aim of sections 4, 5, and 6 to demonstrate a basic connection between the lake hydrology, the marginal ice deformation and the mode of formation of the moraines, a morphometric analysis based on photogrammetry of the exposed moraines in relation to the former lake is presented next. This work was done by the Special Projects Section, Topographic Mapping Division, Department of Energy Mines and Resources. A Wild A 7 Stereoplotter was used.

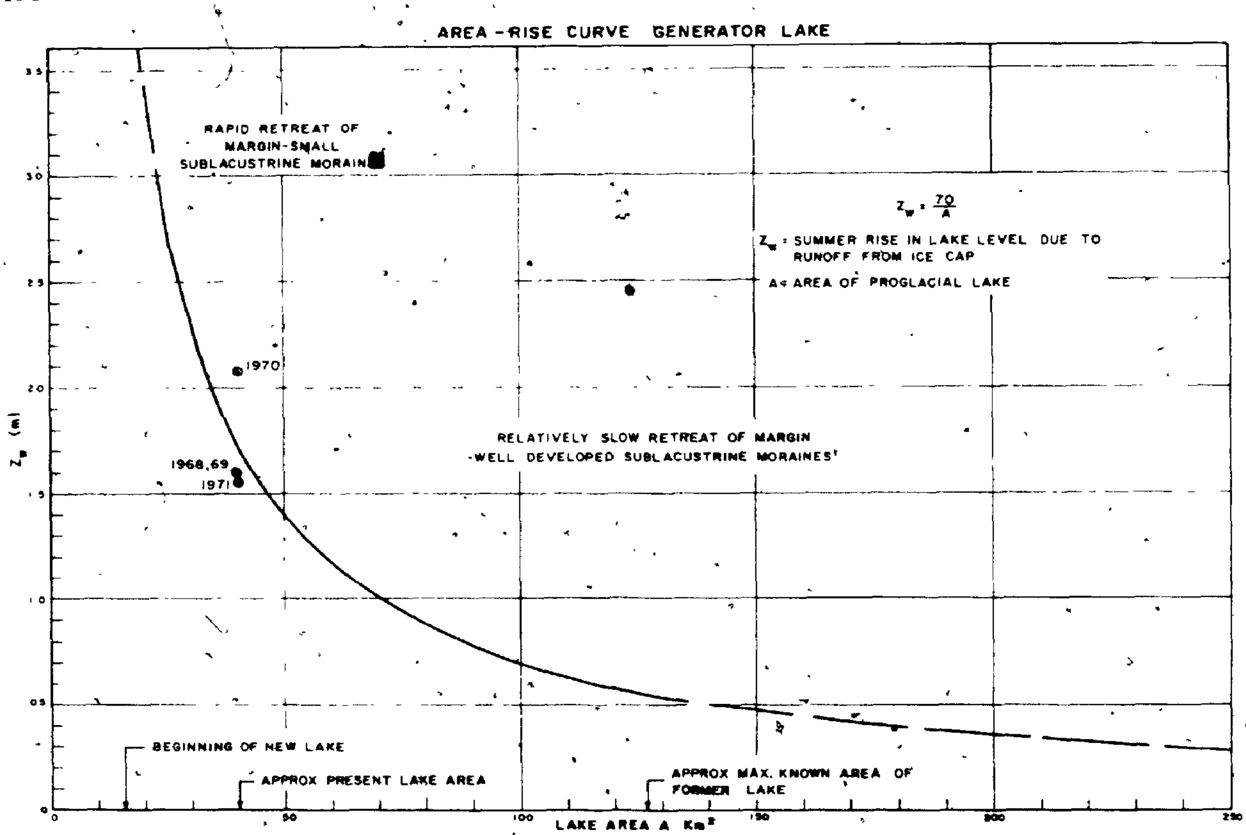

FIG. 13. Idealized relationship between lake area and summer water level rise.

4.3.1 Height/breadth ratio for the moraines of the former lake

The moraines exhibit a characteristic cross-sectionally asymmetric form (Fig. 2 and section 3). Figure 14 shows a plot of 128 values of height, h_m, and breadth, b_m, along 4 traverse lines (Fig. 5, lines I–IV). The estimated error in h_m is 0.2 m and in b_m ± 2 m. A least squares line adjusted slightly to pass through the origin yields

[4*b*] $$h_m/b_m = 0.13\ (\pm 0.01)$$

which expresses a characteristic ratio for the moraines of the former lake basin. We shall use this quantity to calculate (independently from section 3.4), the average amount of material in the moraine system.

Consider the model shown in Fig. 14 where two consecutive moraines of similar geometry have heights of h_m and h_{m+1} and base widths b_m and b_{m+1} and are contained within a length $L_{m,m+1}$.

The amount of material contained within this space (per unit width) is approximately

[4*c*] $$\tfrac{1}{2}b_m h_m + \tfrac{1}{2}b_{m+1} h_{m+1}$$

provided we neglect the amount of till between moraines and assume that they rest on bedrock. According to our field observations this is a reasonable simplification of the problem. If this is not true, an *underestimate* for the volume of till will result.

On the basis of the characteristic moraine cross section expressed by Eq. [4*b*], expression [4*c*] can now be rewritten as the average volume of moraine per unit width, per unit length as

$$\alpha = 3.85\left(\frac{h_m^2 + h_{m+1}^2}{L_{m,m+1}}\right) m^3 \times m^{-1}\ (\text{width})\ m^{-1}\ (\text{length})$$

An average value of $(h_m^2 + h_{m+1}^2)\ /\ L_{m,m+1} = 0.30\ (\pm 0.03)$, using 60 pairs of moraines, was

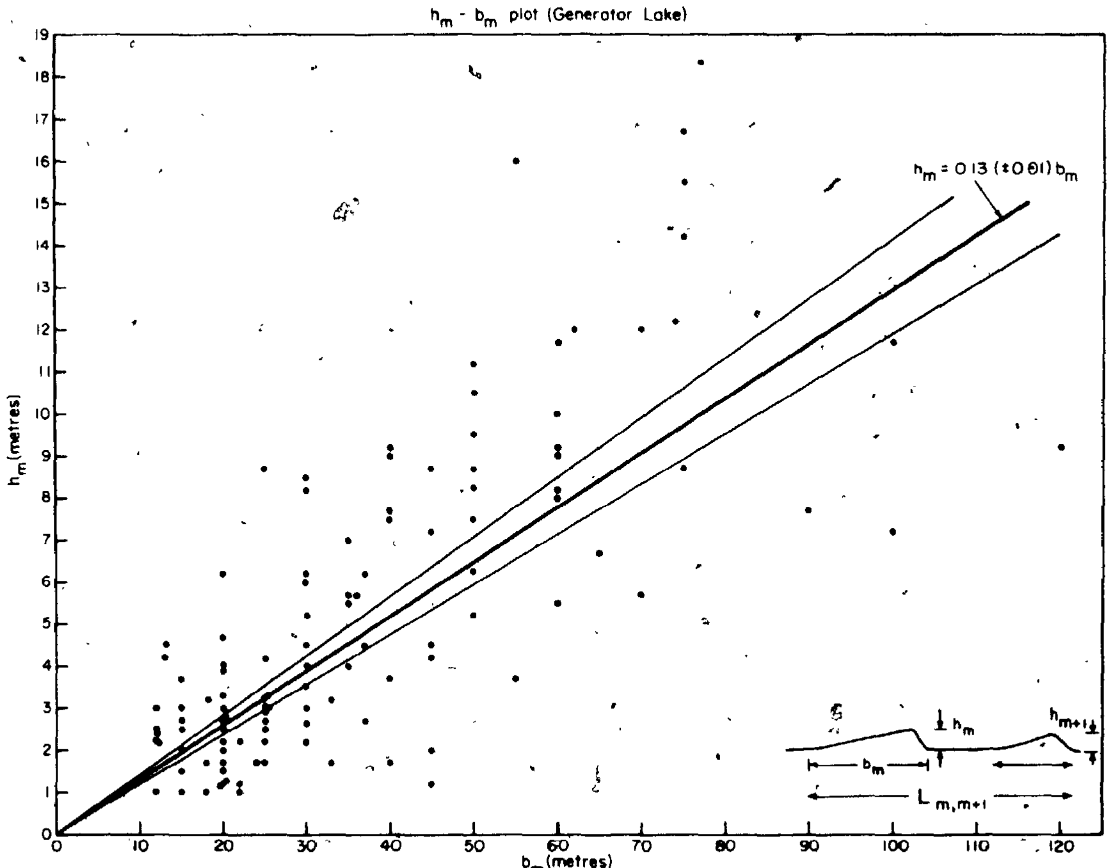

FIG. 14. Plot of moraine height *versus* base length for 128 moraine cross sections from traverses I–IV (Fig. 5).

obtained; hence

$$\alpha = 3.85 \times 0.30\ (\pm 0.03)$$

$$= 1.2\ (\pm 0.2)\ \mathrm{m^3\,m^{-1}\,m^{-1}}.$$

Thus, a minimum estimate of the average amount of till contained within the moraine system of the former lake basin is *about* 1 m^3 per m width, per m length along the *central* axis of the basin (c.f. section 3.4). If the average retreat rate of the *former* ice margin was ~15 m per yr (section 3.3), this gives a value of ~15 m^3 per m (width) per yr for the *average* rate of till deposition along the central axis of the former basin. Section 5.7 gives the present day rate of till deposition.

Towards the shores of the lake the deposits are probably associated with a cliffed margin, and hence controlled by different depositional conditions. Trough to crest heights of individual moraines were observed to decrease quite rapidly toward the margin of the former lake.

4.3.2 Size of the moraines in relation to distance from the former outlet

Figure 15 shows a plot of moraine height against distance from the former lake outlet. The traverses are shown on Fig. 5, as I–IV. A general increase in moraine height with increasing distance is clearly shown. On the basis of this result a plot was made of $\bar{h}_m$, the mean moraine height over intervals of 1500 m, versus $\bar{A}$, the mean area of the former lake corresponding to the midpoint of the group of moraines (usually 7–16). A regression line through the seven points (Fig. 16) has equation

$$\bar{h}_m = 0.074\,\bar{A} \qquad [4d]$$

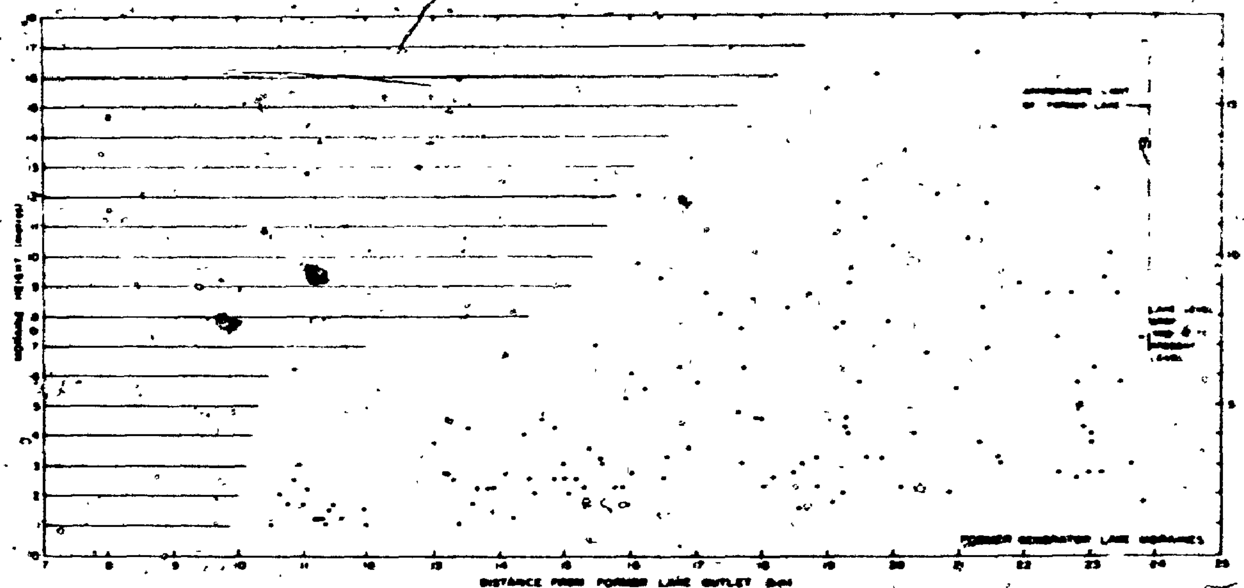

FIG. 15. Plot of moraine height *versus* distance from former lake outlet along traverses I–IV (Fig. 5).

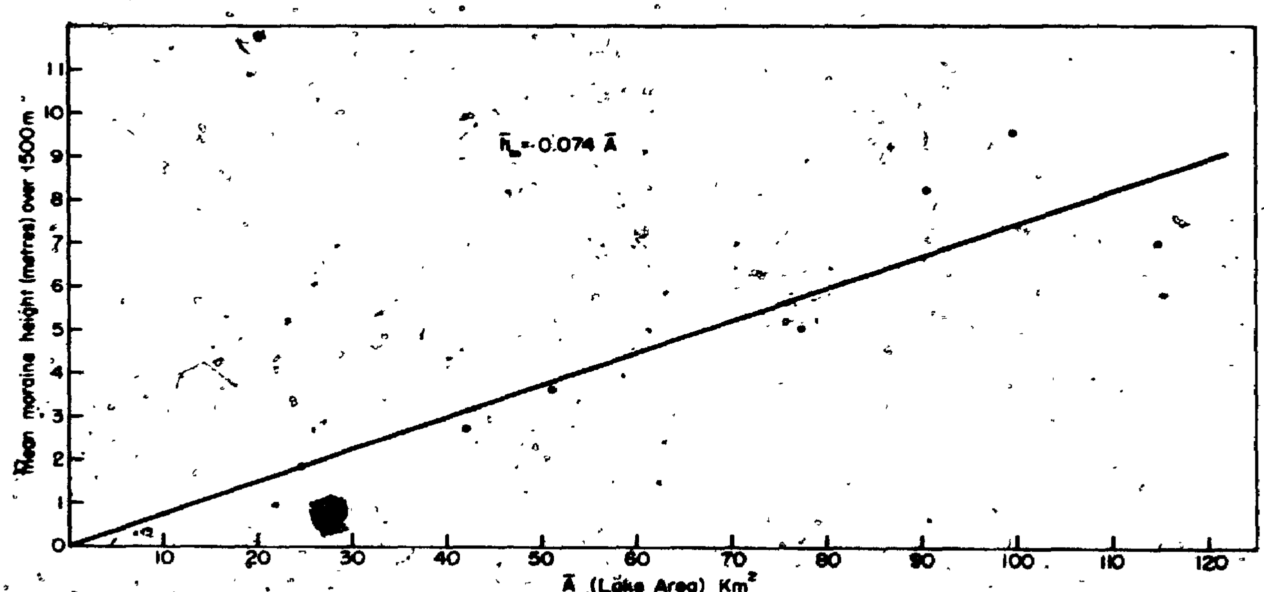

FIG. 16. Plot of mean moraine height (averaged over 1500 m) *versus* average former lake area measured through the middle of the moraine group (7–16 moraines per group, 7 data points).

with a correlation coefficient of 0.90. An insignificant adjustment was made so that the line passed through the origin as it should. In this instance, an indication of a linear relationship between the two parameters is important, physically.

4.3.3 Moraine size – water level rise relation

The height, h_m, of an individual moraine is taken as an index of the size and, hence, volume of the moraine (per unit width). The mean moraine height, $\bar{h}_m$, can be related to z_w by using the data of Fig. 13 and 16. The result

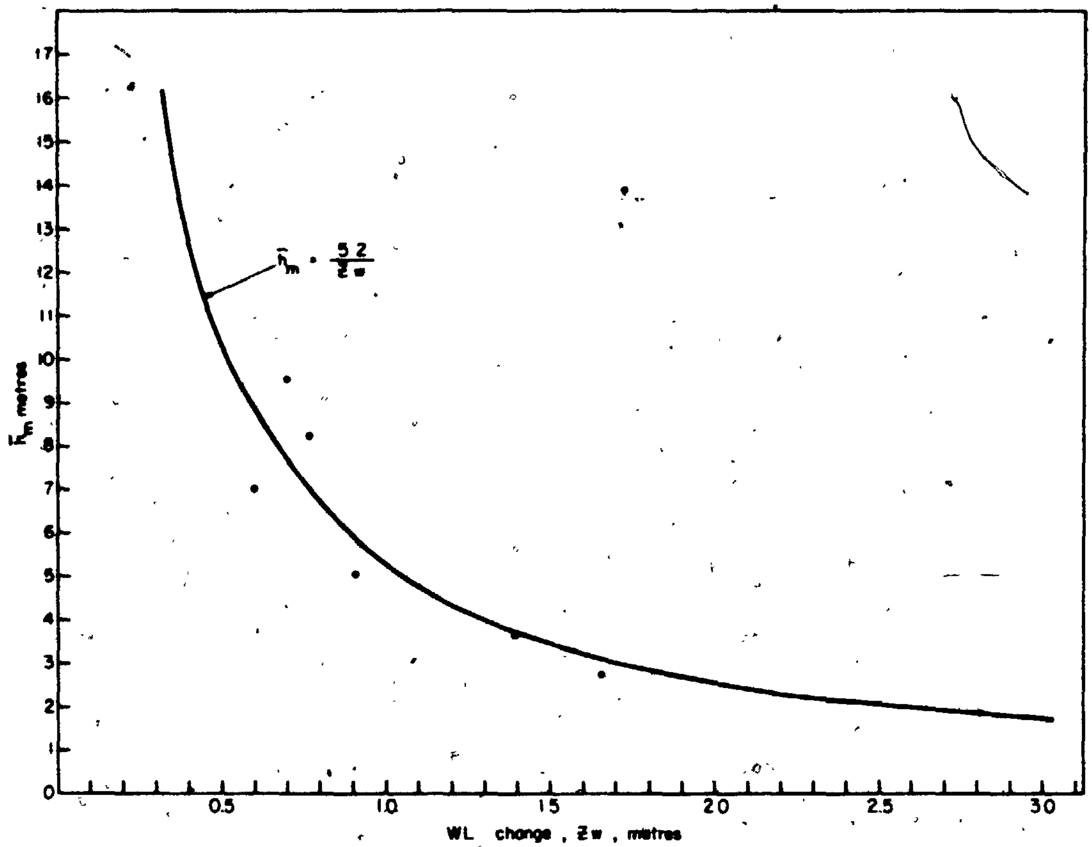

FIG. 17. Plot of mean moraine height *versus* summer water-level change. Based on Figs. 13 and 16; 7 data points.

is shown in Fig. 17, and it suggests it is now worth seeking a *physical* connection between the fluctuations in lake level and the size of the moraines associated with the ice front existing at that time. Such a connection is demonstrated in section 6.

5. Ice Ramp and General Margin Morphology

5.1 Ramp Morphology

The shape of the ice ramp containing pole T1 (Fig. 18) was determined in 1970 from equilibrium considerations after a major fracture had occurred, separating in several pieces, about 0.4–0.5 × 10^6 m^3 of ice, from the ice cap. A process of successive graphic iterations was used to reconstruct the underwater form of the largest block. Subsequent movement of this block, initially grounded and confined by lake ice, confirmed the cross section was markedly tapered. Fracture of the rear edge (Fig. 18) indicated the existence of a curved initial fracture surface which was suggested by the graphic analysis and which is also expected from theoretical considerations (Holdsworth 1973*a*). The cross section of the three main ramps is shown on the right in Fig. 18.

5.2 Hydrography

Water depth soundings close to the ice front were made with an Elac (Castor type) electromagnetic echo-graph enabling continuous profiling to be done. At distances of 10–15 m from the ice front, the vertical return echo became poorly defined and spread over a wide range. Similar results were obtained with a Kelvin-Hughes MS 26B echo sounder earlier. Some angled profiling was done where the transducer was pointed at various angles towards the submerged ice front, ensuring in this case a reflecting surface of ice. The signal was the same as obtained from the vertical profiling at 10 m from the ice front. Partial recon-

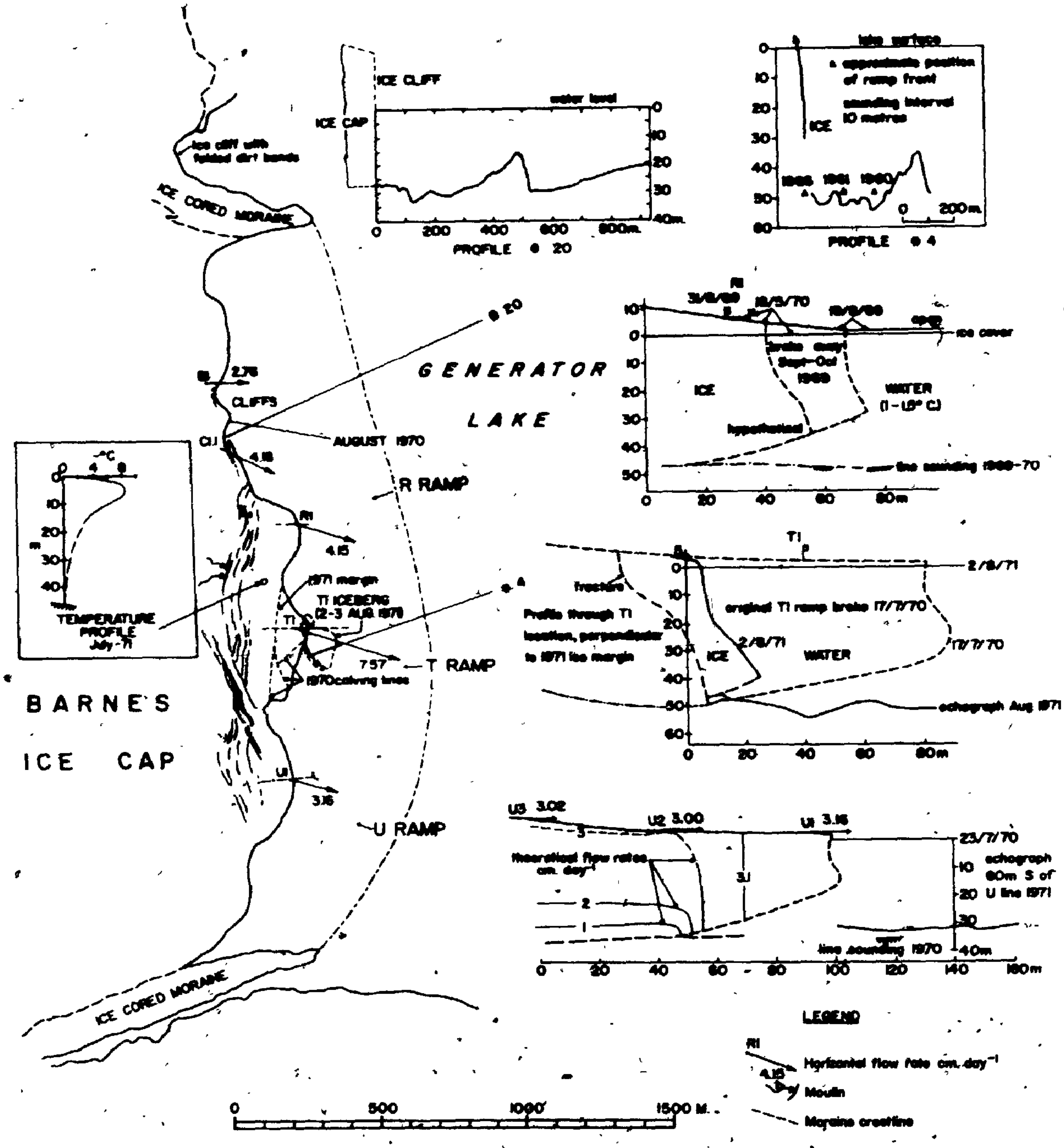

FIG. 18. Map of ice front at Generator Lake showing the ice-ramp shapes and submerged moraine profiles.

structions indicate that a *ram* (underwater projection) is common. A ram is the natural result of an initial fracture due to bending and once formed, it has a tendency to be maintained.

5.3 Seismic Soundings

A Century 444 6 channel oscillograph connected to 6 (30 Hz) geophones through a 12 channel amplifier served to record reflections from the ice–rock interface. Charges varied from booster caps to ½ lb charges of Forcite. The speed of the seismic wave in the surface ice was found to be 3750 ± 50 m s^{-1} (at an average temperature of −3.5 °C).

Because of anisotropy of the ice and the presence of crevasses, many returning pulses were often difficult to interpret. Several short profiles were obtained over a distance of about 150 m between the grounding line and the major belt of crevasses. Ice depths varied between about 50 and 80 m.

5.4 Temperature Profiles

A horizontal tunnel was dug in the ice 8 m below the surface of the glacier in an abandoned, partly filled moulin. A SIPRE corehole, 6 m deep, was drilled into the base of the tunnel and several angled, 1 in. diameter, 2 m long holes were drilled from the end of the 5 m long tunnel. Ice temperatures were measured in all these holes using a string of thermistors. The resulting temperature profile is shown in Fig. 18 (inset). The remaining part of the profile is dashed and is assumed to reach 0° (strictly the pressure melting point) at the bottom. This result is based on the following observations: (*1*) A geothermal gradient of 1°/22 m (Ward 1952). Even if the gradient was as low as 1°/44 m (world-wide average), it would change the curve only slightly. (*2*) At least three active moulins were seen above the crevassed area and water, which was flowing into them, must have been transmitted along the base of the glacier. This has been confirmed by dye experiments (Barnett 1966, p. 27).

Assuming the base to be at or near the melting temperature of ice, the average temperature of a vertical column of ice is found to be −3.5 ± 0.5 °C. In addition to probably having a perennial water film at its base, this ramp ice must have once passed through the crevasse zone, thus exposing at least the upper 20 to 25 m to warmer than normal temperatures during the summer. Thus, the residual 'cold wave' is probably warmer than it would be for the uncrevassed ice. Lake water temperatures indicate vertical mixing and subsurface ice melt. From the change in freeboard and hence thickness of a large floating section of ice ramp, it would appear that subsurface ice melting is continuous and of order 3–4 m yr^{-1} (~1 cm day^{-1}), whereas upper (measured) surface ice ablation is about 1 to 1.5 m yr^{-1}. (*3*) The surface ice flow rates are much greater than is expected from the flow law and known ice thickness and temperature. Basal sliding would explain these observations.

5.5 Debris Content in Ice

On the north shore of the lake (Fig. 18), an ice cliff containing a large flow fold (Fig. 19) defined by dirt bands provided an opportunity for sampling the dirt content in what once must have been basal or near basal ice. This fold is tentatively associated with a recent surge of part of the south dome of the ice cap. Three dirty ice samples provided a means of estimating the debris content of the ice. Larger sized particles, not occurring in the samples but occurring abundantly in the ice cliff, were accounted for by constructing a grading curve based on the adjacent moraine particle distribution and with this as a basis, computing the total weight (and hence volume) of a 'theoretically complete' dirty ice sample.

An average value of 8 ± 2% by volume (based on a till density of 2.2 g cm^{-3}) was determined. By considering the geometry of the fold, a prefold thickness of dirty ice would have been about 8 m. The amount of debris above that height was negligible. The rate of supply of debris may now be calculated knowing the ice flow rates.

5.6 Ice Flow Rate Determinations

A surface survey in 1969–71 showed that

FIG. 19. Flow fold in ice cliff (approximate height above lake level is 15 m). For location see Fig. 18. (GSC photo 161627).

the T1 ramp contained the fastest flowing ice of the whole margin at Generator Lake. Before the fracture, T1 had a speed of 27.63 m yr^{-1}, horizontally. From the strain rates, a surface speed of about 27.37 m yr^{-1} was determined for a point just behind the hinge line of the ramp where the surface flow changes from slight compression to slight extension. On the basis of a simple flow model (Nye 1957), assuming laminar flow, the flow rate, u, parallel to the surface can be expressed:

$$u = u_s - [2/(n+1)]\,[\tau_b/(BH)]^n\, z^{n+1}$$

where u_s is the surface flow rate, τ_b is the basal shear stress, H is the ice thickness perpendicular to the surface, z is the depth below the surface, and B and n are constants in the flow law; n has been taken as 1.7 (Hooke 1973) and B has been taken as 1.0 bar $yr^{1/n}$ which is close to the value for temperate ice. This value may be too low but it sets an upper limit to the velocity *difference* between the surface and the bottom (about 12 m yr^{-1}), implying that the basal sliding rate is about 15 m yr^{-1}, or more than half the measured surface speed. Under these conditions the average flow rate in the lower 8 m of ice is about 17 m yr^{-1}. A 20% variation from this value might exist.

5.7 Rate of Till Supply

Based on the data of sections 5.5 and 5.6 the rate of supply of basal till from an approximately steady state lower surface is:

$$8\ (\mathrm{m}) \times 17\ (\pm 3)\ (\mathrm{m\ yr^{-1}}) \times 8\%\ (\pm 2\%)$$

$$\text{per m width} = 11 \pm 5\ \mathrm{m^3\ m^{-1}\ (width)\ yr^{-1}}$$

Assuming the 'typical' moraine form of section 4.3, the time taken to build a 3-m high moraine (× 35 m base length) would be about 5 yrs. The question then arises: What value of till release rate should be appropriate for the former lake basin? Considering the probable differences in ice thickness and surface slope of the ice ramps between the former lake and the present lake, the ± error in our value is likely to cover differences that might have existed between the two. This assumption is strengthened by the calculation given in section 3.4.

6. Mechanics of Ramp Bending and Relationship to Lake-Level Fluctuations

6.1 Bending Theory

Holdsworth (1973) shows that at distance x from the hinge a bending moment, $M(x)$, occurs along the ice ramp, due to the *upward* thrust exerted by the buoyancy force acting on the ice ramp for the case of *rising* water level where the *rate* of rise water is much greater than the rate of vertical response of the ice with respect to the same datum. The bending moment produces a bending stress, σ_{xx}, which, irrespective of whether elastic or plastic theory is used, can be written to a good approximation in the form

$$\sigma_{xx} = [K \cdot M(x)]/H^2 \qquad [6a]$$

for the upper or lower surfaces. K is a physical-geometric constant dependent primarily on whether elastic or plastic theory is used and on the position of the neutral axis of bending. Whichever theory is used, K is not the same above and below the neutral axis if this is not at the midslab position. H is the thickness of the ice ramp at distance x.

For a given geometry and appropriate constitutive law, and for the case of upbending of the ramp, the tensile stress deviator on the *lower* surface, at the hinge ($x = 0$) can be expressed as

$$\sigma_{0x} \propto M(0), H^{-2}. \qquad [6b]$$

But

$$M(0) = \int_0^L \rho_w g\ \Delta z\ x\ dx$$

(neglecting the Reeh bending effect at the end). Thus, $M(0) \propto \Delta z, L^2$ where L is the length of the ramp measured from the hinge line, and Δz is the difference between the equilibrium water level (with respect to the ramp) and the actual water level. This can be expressed as

$$\Delta z = z_w - z_i \qquad [6c]$$

where z_i is the rise of the ice in response to drowning.

Since by observation, $z_w \gg z_i$ and also

$$\frac{dz_w}{dt} \gg \frac{z}{dt}$$

then we can write $\Delta z \approx z_w$

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TABLE 4. Ice front retreat rates: former lake

Lake level above present level	Lake area (km^2) at mid-range	Age range yr B.P.	Average retreat rate $m yr^{-1}$	Remarks
76 ± 1 m	3.8	4000–3830	42	See Table 2
76 ± 1	29	3830–3480	26	
76 ± 1	66	3480–3090	7	Shortest distance
76 ± 1	98	3090–2520	14	

Therefore

$$[6d] \qquad \sigma_{0x} \propto z_w, L^2/H^2$$

which implies that as z_w decreases (by Eq. [4*a*]—corresponding to an increase in lake area—on a time scale much greater than a year), the ramps are able to grow longer (*or* become thinner) for the same value of σ_{0x} (critical) necessary to cause a calving. In particular, according to Eq. [6*a*] for $x = 0$, *i.e.* at the hinge, if σ_{0x} (critical) and K remain the same and M (0) is reduced by, say 50% (*e.g.*, by a reduction in z_w), then the new ice thickness at the hinge could be about $(1 - 1/\sqrt{2}) \times 100\% \approx 30\%$ less than before. As a result of reduced lake level fluctuations, ramps would survive longer and the *net* rate of retreat of the ice margin (averaged over times much greater than the time of existence of an individual ramp) would progressively *decrease* as z_w decreased (corresponding to an increase in A (Fig. 13)).

6.2 *Ramp–Moraine Model*

The ramp–moraine model is based on the *assumption* of a retreating margin, and hence an average negative net mass balance of the glacier. Any deviations from this state are clearly covered by the bending theory in this section, but obviously complicate interpretations based on the geometric arguments of section 4 dealing with the moraines. However, over long periods of time, in the order of 10^2–4×10^3 yrs, with which this model study is concerned, a mean retreat rate emerges which masks any fluctuations. The data of Table 4 indicates the magnitude of *retreat rates* corresponding to average lake *areas* existing at *that* time. The data tend to support the hypothesis that retreat rates are lower, the *larger* the area of the lake.

If a wedge-shaped moraine (Figs. 2 and 4) is forming underneath a ramp (Fig. 18), the size to which the moraine may develop depends on the time, t_r, of existence of the ramp, assuming that the rate of supply of till from within the ice is essentially constant over the same period of time. If the moraine height h_m is used as an index of size, then

$$[6e] \qquad h_m \propto t_r$$

under a steady state condition, over t_r. But Fig. 16 indicates that $h_m \propto A$, therefore, by Fig. 17

$$[6f] \qquad h_m \propto z_w^{-1} \propto t_r.$$

Thus, the *size* of moraine is proportional to the time of existence of the ramp or inversely proportional to the maximum summer lake level rise (provided $dz_w/dt \gg dz_i/dt$).

Because there are other mechanisms of calving and because a steady-state condition is not realistic over long periods of time, the limiting case $t_r \to \infty$ as $z_w \to 0$ is not physically meaningful. Figure 13 will tentatively be applied to both the former and the present lakes in accordance with the remarks of section 4.2. As the area A of the present lake is just exceeding 40 km^2, z_w is still decreasing quite rapidly with increasing A, thus t_r is tending to increase quite rapidly.

According to Eq. [4*d*] the expected average height of the moraines forming now should be about 3 m, *i.e.* they should be easily detectable with continuous depth profiling equipment. The sublacustrine moraines detected near the present ice front (Fig. 18) are from 1.5 to 3 m high in many places with the exception of the large outer moraine which is up to 15 m high. This moraine may have been associated with (*1*) a period of positive net budget of the ice cap associated with an advance of the ice front, (*2*) a sufficiently long steady state condition, implying that the ice front was stationary for about 200 yrs, or (*3*) a surge. This latter explanation is strongly supported by

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glaciological evidence (Holdsworth 1973*b*). Furthermore the prominent ice-cored moraine peninsulas north and south of the ice-front (Fig. 18) are continuous with the submerged moraine and, therefore, the latter cannot be, by definition, the type of sublacustrine moraine with which we are primarily dealing. Also, the 15 m high moraine is remarkably uniform in height above the lake bed at each site sounded. This observation is at variance with the measurements made on the exposed sublacustrine moraines (section 2.4), but is consistent with a surge involving block ice movement and bulldozing of sediment, irrespective of lake depth.

In the present lake, recognizable moraines occur within 400 m of the ice front whereas in the moraine model h_m must be averaged over distances of the order of 1500 m; therefore, it is premature to apply the results of Fig. 16 and Eq. [4*d*]. Probably one of the reasons why the present set of sublacustrine moraines has appeared is that (*1*) the lake depth has only recently become deep enough along a sufficiently wide front of ice margin and (*2*) summer water level fluctuations are now only just becoming sufficiently low, for well developed ramps to form.

7. Mechanism of Moraine Formation

A model has been presented which permits the buildup of moraine in a wedge-shaped space beneath a tapered section of ice (a *ramp*) projecting into the lake. Because it is unconfined, a ramp has a tendency to spread and thus could maintain a space between the lower surface of the ice and the lake bottom for at least several years, provided, lake level fluctuations remain within certain limits (section 6). During this time, till, released from the ice by melting, gradually fills up the space, which may have been increasing with time, if the local net-ice balance is negative.

Since 1948 a number of discrete ramps (3 or 4) have been characteristic of the ice margin, but it should be noted that this relationship with the present lake may be only several decades old. With different sublacustrine/ice topography and deeper water it seems possible that one larger ramp may have occurred. Such a condition could then explain long continuous sections of moraine occurring in the former lake basin. Sand rich 'caps' or kames occurring usually on the crests of the exposed ridges in infrequent numbers may be explained as moulin washings. There are three sources of this material (*1*) surface, of aeolian origin, (*2*) englacial, and (*3*) basal ice. Because of the necessity of transmitting water through cold ice, the existence of crevasses must be invoked. The present ice front and crevasse system offer a basal environment for sand—transported via the moulins—to collect. Finer fractions are carried in suspension into the lake.

It has been shown that the volume (per unit width) represented by a cross section of moraine can be explained sufficiently on the basis of sample volumes of debris contained within the ice, ice flow rates, and time control. The final shape of a particular moraine may be largely determined by subsequent movement of the grounding line or by rotation of the ramp after it has calved off. It is maintained that neither of these undoubtedly common processes is likely to obliterate completely the original form of the moraine. In support of this contention: If all the moraines of the exposed moraine 'system' as well as the presently submerged ones are considered, the predominance of the markedly asymmetrical cross section type is conspicuous and this would mean that the results of the fundamental formation process are not erased. The till fabric analyses support our model. The ramp–moraine model is expected to fail when lake water depths become less than about 30–40 m. Under these conditions, mechanical breakdown of the ice edge is increased and retreat of the margin is faster than for the ramps.

However, sublacustrine moraines have been formed in water less than 30 m deep in the 76 m basin. They are typically smaller in magnitude and their asymmetry of form is less marked. Undercutting of the ice cliff by melting still takes place in shallow water creating a space for moraine formation. For example a large cliffed section of the ice margin sank into the lake on July 4, 1968 sending shock waves completely across the lake. This indicated a substantial undercut. A similar slumping of the cliff north of the R1 ramp occurred in 1969–70. The effect of such collapse on any moraine present could be disruptive if the block grounds. Although the process of till release will remain similar under these circumstances, the glaciological environment is less stable than for the ramp form, thus typically yielding smaller, less well formed moraines. With even more rapid retreat of the margin the lake bottom will lack significant moraines (or till relief) as is typified by most of the present lake basin west of the Clyde River outlet. This in turn is typical of the proposed relationship of small lake area at the beginning of a new phase and rapid retreat rates of the margin.

In section 3 it was found that the apparent time of retreat of the margin from the time of emptying of the former lake to the present position (1000–1200 yrs) seemed anomalously long. This anomaly is explained by a more complicated history of the ice cap than was hitherto realized. There is moraine evidence (Løken and Andrews 1966) to show that the ice margin withdrew considerably farther than its present position allowing most of the then Generator Lake to be drained down Flyway Lake. A surge then took place, putting the ice margin at a position marked by the 15 m high submerged moraine (Fig. 18).

Figure 10 shows that at the time of first emptying of the former lake the ice must have been at least 110 m thick at the haunch of the ramps which would have broken under downbending due to drastic water level drops (the total possible drop was 76 ± 1 m). At most this would have produced an ice cliff rising about 80 m above water level. Such a cliff is unstable (Appendix B) and the ice margin would have rapidly thinned and readvanced, partially closing off the Clyde River outlet. Gradually, as the lake filled up again, ramps would have formed once more until a more stable condition had been reached. A subsequent net retreat of this margin, associated now with slightly thinner ice, would have again resulted in partial emptying of the former lake as the Clyde River outlet was further exposed. This cycle could have been repeated several times, but four is suggested by the deltaic evidence relating to the lake phase (section 3.2).

The subsequent reduction of the ice thickness near the margin so that it would be consistent with a water depth of 30–40 m could have occupied hundreds of years, after which time the Clyde River outlet would have become established. Because of the oscillations of the ice margin through the constricted part of the lake, an anomalously large amount of moraine should have been dropped in this region. The depth sounding profiles (see Fig. 2, p. 16 and Fig. 4, p. 20 of Barnett *et al.* 1970) suggest this could be the case. Also, there would be evidence of local readvance of the lobe of ice and these may be seen on the aerial photographs as sections of arcuate moraines to the east of the constriction in the lake.

After the withdrawal of the ice margin west of the Clyde River outlet, lake depth was predominantly shallower than about 30–40 m, so we expect ramps were only weakly developed and cliffs rising 20–30 m above lake level, predominated. Under this regime no prominent sublacustrine moraines would be expected unless there were periods of marked standstill of the margin. It is no surprise, therefore, that the bottom topography of the lake, west of the constriction (near Clyde outlet), is deficient in moraine ridges. According to Table 4, the apparent average retreat rate from the Clyde River outlet to the present margin is anomalously low if the radiocarbon dates are correct and assuming that the ice front has been only retreating. If the surge theory is accepted we must reevaluate the average net retreat rate of the margin over the last 1000 yrs. It is premature to try to calculate this value exactly.

The presurge front is estimated to have been 4–4.5 km (along a flow line) behind the present margin and in a position which no longer retained the lake. Because the lake would have essentially drained even before the ice reached the 'presurge' position, the margin must have been retreating at rates appropriate to a land-based margin for a considerable time before the surge. The possibility of limited ponding in the area of the lake basin is not excluded. Thus the history of the ice front in the last 1000–1200 yrs involves a relatively rapid retreat until emptying of the lake (estimated at 900–800 B. P.) followed by slower retreat on land until near the turn of this century (Holdsworth 1973*b*) when a surge occurred. This assumes only one surge, as the evidence permits.

In the model set up for the lake–ice front–moraine system, it is possible, in principle, to predict the average height of moraines forming now and later, simply by knowing the area of

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the lake and using Fig. 16. This is with the provision that ramps exist and/or lake depth exceeds approximately 40 m. On the basis of (*1*) the present ice surface elevations, (*2*) ice thicknesses obtained from Littlewood (1952) for the south lobe of the ice cap, and (*3*) taking the ratio between the recent average retreat rate of the land based margin to the recent average retreat of the adjacent lake-based margin as being up to about 1:8, it is possible to construct the future configuration of the south lobe margin just before preglacial drainage is reestablished into Flyway Lake (Fig. 5). The above ratio was determined by using distances measured west from the junction of the last exposed sublacustrine moraine, the former lake shoreline and the extension of the original land based marginal moraine dated at about 1000 yrs B. P. According to the above calculation, just before drainage, Generator Lake will then be about 47 km^2 in area. The average size of the sublacustrine moraines that will be last to form should be no more than 3.5 to 4 m high. Although it is not essential to the main point, the assumption has been made that the ice cap margin will continue to retreat (at a decreasing rate according to the model) until the lake drains. Local variations will occur, for example in zones corresponding to a bedrock convexity where crevasses form and retreat rates locally speed up. The ice front happens to be now in such a phase.

It is concluded that having accepted our interpretation of the recent history of the ice margin, the behavior of the present lake–ice front–moraine system supports the model presented in sections 4 and 6, bearing in mind the recent departures from the idealized model conditions (such as extensive areas of shallow water unfavorable for ramp formation and the zone of crevasses near the present front). Another factor which causes departures from the idealized model are mass balance changes in the ice-cap producing fluctuations in the ice margin position, rather than a continuous retreat.

No attempt is made here to compare the present sublacustrine moraines with several other groups of lake associated multiple moraine ridges recorded in the literature. There are already similarities recognized to exist between the present study area and other areas containing such ridges; the task of comparison should be treated in a separate paper.

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Appendix A
Meteorological and Related Data Collected at Generator Lake, 1968–71

1968 No systematic observation program was followed except for the lake level readings at 12 hr intervals. The lead or moat was first observed on July 6. The summer period was cool and cloudy and undistinguished except for very strong winds which occurred on July 11 following a blizzard on July 10. The lake ice did not disintegrate and a substantial amount remained in late August.

1969 Systematic meteorological data were collected—including maximum and minimum temperatures, total wind, precipitation, and pressure—with data from July 5–August 31. The first observation of water around the lake was on July 3. August was particularly pleasant with uncharacteristically high amounts of sunshine which gave a higher mean temperature for August than July. The lake was clear of ice on August 24.

Summary

	July (date)	August (date)
Temperature, mean	44.4 °F (6.9 °C)	45.5 °F (7.5 °C)
Temperature, max.	67.8 °F (28) (19.9 °C)	64.9 °F ~~(3) (18.3 °C)~~
Temperature, min.	28.8 ° F (7) (−1.8 °C)	29.7 °F (23) (−1.3 °C)
Precipitation	0.1 in.	0.595 in.
Maximum fall	0.04 in. (17) (rain)	0.46 in. (29) rain
Wind run:		
12 hr mean	131.33 km	141.35 km
12 hr max.	320.3 km (10)	340.8 km (16)
12 hr min.	12.1 km (7)	38.0 km (5)
Prevailing wind	NE (19% of observations, 2 per day).	NE (24% of observations, 2 per day).

1970 Approximately 8-hourly observations were made at the lake camp from July 9–August 23. The first observation of a marginal moat was July 1. Lake ice thickness in May was 1.6 ± 0.1 m (snow cover 0.4 ± 0.1 m) decreasing to 1 m by August 7 and 0.7 m by August 20. Lake ice remained until August and it is assumed that no substantial break-up occurred. This was confirmed by observations in 1971 when second year ice could be distinguished from first year ice.

The following is a summary of the temperatures of July and August. Location of station: (see Fig. 6) Elevation 30 m above lake level.

Temperature	July (date)	Remarks	August (date)	Remarks
Mean	40 °F (4.4 °C)	part month only	37 °F (2.8 °C)	part month only
Maximum	54 °F (28) (12.2 °C)		52 °F (2) (11.1 °C)	

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Minimum	32 °F (0 °C) (20)	"	28.4 °F (−2. °C) (23)	"
Precipitation:	drizzle or rain–5 days		rain or snow–8 days	
Wind run				
12 hr mean	162 km			
12 hr max.	445 km			
12 hr min.	53			

1971 Observations were made from July 1 to August 5. The marginal moat formed June 17–18. Lake-ice thickness was 1.7 ± 0.1 m in May. Breakup of lake ice occurred in early August and the lake was clear of most ice by August 17. T1 iceberg grounded.

	July (date)
Temperature	
Mean	43.6 °F (6.4 °C)
Maximum	57.2 °F (14 °C) (26)
Minimum	31.6 °F (−0.2 °C) (15)
Precipitation	rain–12 days
Wind run: 12 hr mean	175 km
12 hr. max.	385 km
12 hr min.	32 km

Appendix B
Stability of a Free Ice Cliff

To see the main elements of the ice cliff deformation and the establishment of a 'controlling height,' a simplified model. is used. Consider the behavior of a perfectly plastic slab of uniform thickness h resting on a flat base. There is a free vertical edge. The material has a yield stress of k (bars) in pure shear. The horizontal stress deviator at the base of the cliff (excluding atmospheric pressure) will be approximately $\bar{\rho}_i gh$, where $\bar{\rho}_i$ is the average density of the material and g is the acceleration of gravity.

Assuming that $h > 2k/\bar{\rho}_i g$, the material will tend to spread until $\bar{\rho}_i gh = 2k$. For the simple model of Nye (1951) $k = 1$ bar; therefore $h \approx 23$ m is the height of the ice cliff when the deformation will cease.

In a more realistic, but still simplified model, k may be taken as 0.76 bar (Holdsworth 1973*b*), whence $h \approx 17$ m. In the actual case the creep is then significantly reduced compared with the case when $h \gg 17$ m.

CANADA

GENERATOR LAKE, BAFFIN ISLAND, N.W.T. and TASIUJAQ COVE, EKALUGAD FIORD, BAFFIN ISLAND, N.W.T. 1968

No. 1

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CODC Reference: 22-68-777

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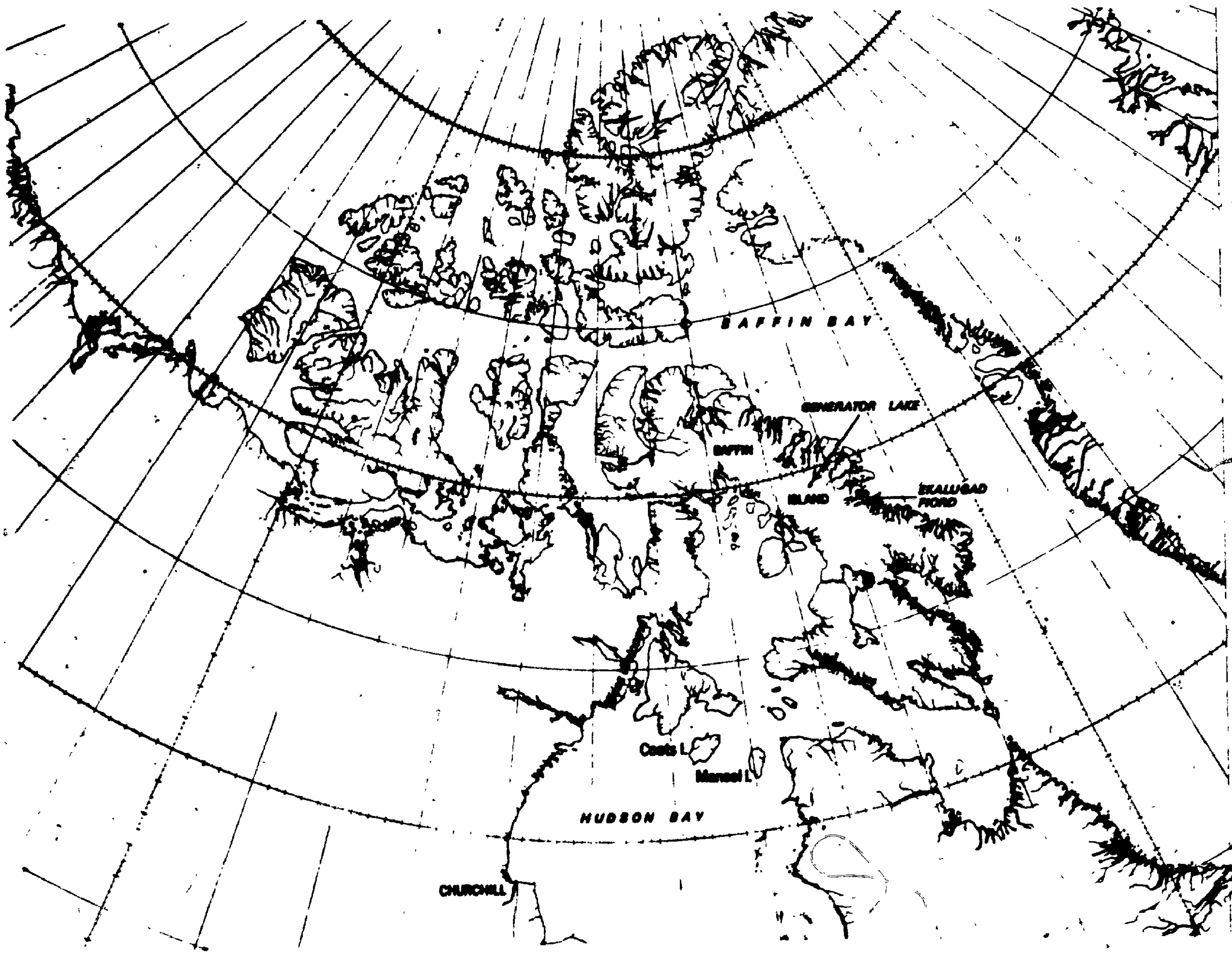

A reproduction of oceanographic plotting sheet number CHS 750 showing the location of Generator Lake and Ekalugad Fiord.

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COLOUR

GEOLOGICAL SURVEY OF CANADA

Generator Lake, Baffin Island

Period of Survey: May 30 - August 16, 1968

Observers: D.M. Barnett
D.L. Forbes
J.K. Whytock

Division of Quaternary Research and Geomorphology

Plate 1 Aerial photograph (A-17042-129) of part of Barnes Ice Cap (left) and part of Generator Lake (from RCAF photo taken July 19, 1961 from 30,000 feet with a 6 inch lens).

GENERATOR LAKE

The data presented here were collected from Generator Lake (frontispiece) as part of a proglacial geomorphology project (680040) of the Quaternary Research and Geomorphology Division of the Geological Survey of Canada; the data are also being used in the preparation of a thesis to be presented to the University of Western Ontario. The development, landforms, and chronology of Generator Lake have been described by Barnett (1967) and further progress reported in Barnett (1969).

Most investigations of the lake environment were made through several feet of lake ice. During the summer of 1966 five bathymetric profiles (Fig. 1) were obtained using a Kelvin-Hughes M.S.26B echo-sounder. Profiles 2 and 7 are shown in Figure 2. Four additional profiles were run in 1968, together with the collection of bathythermograph data, water samples by Knudsen bottle, bottom temperatures by reversing thermometers and sediment sampling by Dietz-Lafond grab at selected stations. A total of 26 bathythermograph traces were collected, 21 water samples and 20 bottom sediment samples. The station positions are shown in Fig. 3 and an interpretation (by DMB) of the bathymetric data in Fig. 4. Additional line sounding data, collected in August 1969 were used to clarify some details in areas less than 20 m. deep.

Generator Lake is located on the interior upland of north-central Baffin Island at an elevation close to 400 metres. It occupies the upper valley of a consequent stream which prior to the last glaciation drained westward into Foxe Basin but which is currently dammed by the Barnes Ice Cap, causing overflow waters to spill eastward to Baffin Bay by way of the Clyde River. The lake's maximum dimensions at present are 13.5 km by 6.5 km; it drains over 400 km^2 of ice-free land and part of the Barnes Ice Cap. The greatest depth sounded was 60.5 metres at a point close to the ice cliffs on profile 3 of Fig. 1.

A preliminary analysis of the bedrock geology of the area has been presented by Jackson (1969); granitoid migmatites and gnesses predominate with some iron formation immediately south of the lake.

Climatically the area is sufficiently cold to maintain lake ice cover of almost two metres and it is only rarely that the lake surface becomes ice-free, as it did on August 23, 1966, and again on August 26, 1969.

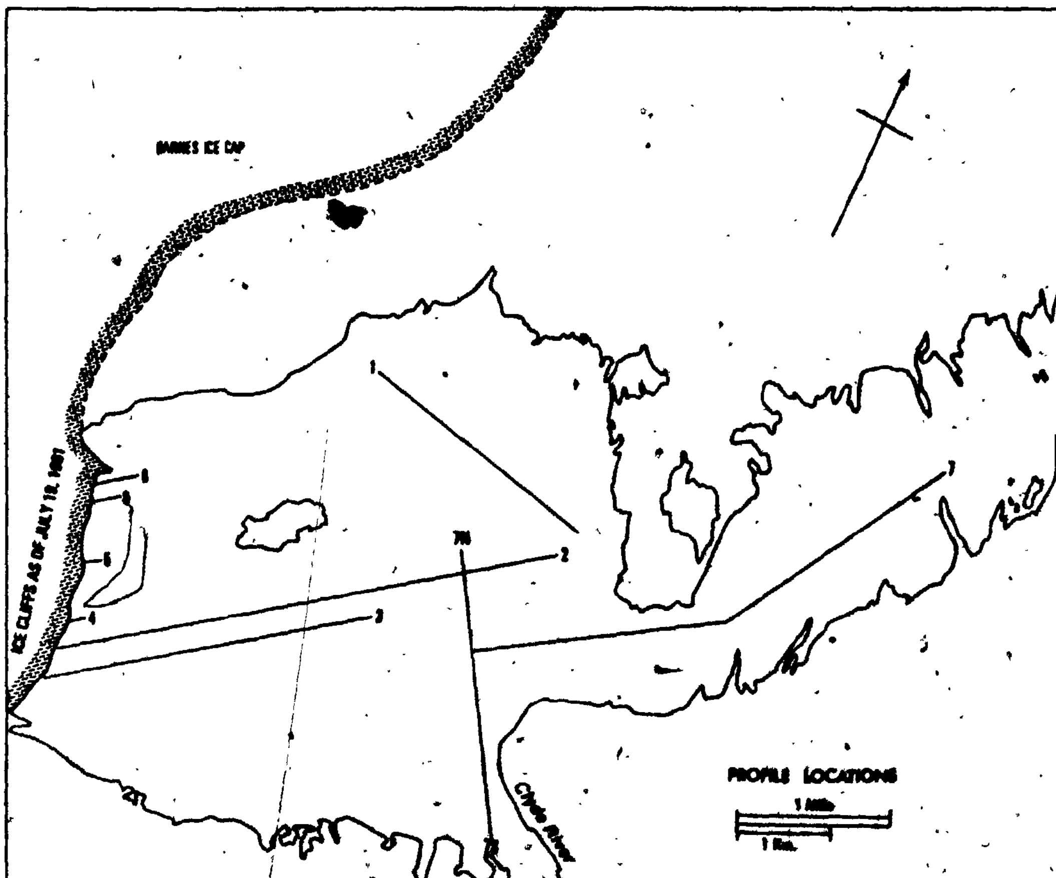


Figure 1 Location of profiles in Generator Lake along which echo sounder observations were made.

The Barnes Ice Cap, which is thought to be a remnant of the last Wisconsin ice sheet has been the subject of glaciological studies for several years (Baird, 1952; Ward, 1952; Sagar, 1966; Løken and Andrews, 1966; Løken and Sagar, 1968) and a related study by Barry and Fogarasi (1968) presented climatological findings designed to develop models for conditions favouring glacierization. As current theory suggests at least the land-based margin of the Barnes Ice Cap is frozen to its base, the thermal distribution of Generator Lake waters is of interest to other investigations of this proglacial environment.

D.M. Barnett

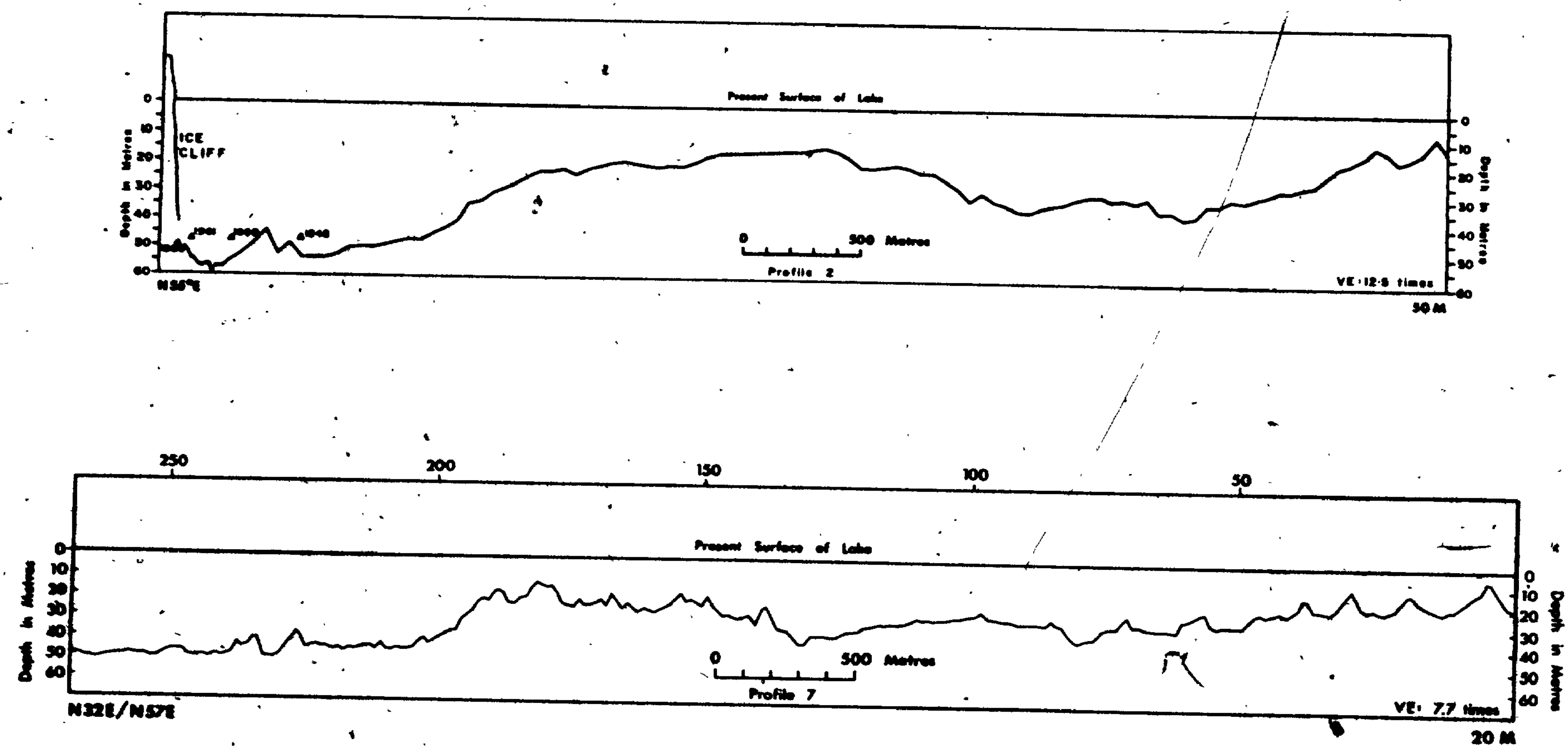


Figure 2 Profiles 2 and 7 showing bottom morphology.

Plate 2 Gasoline powered drill with 4 inch bit. A five hole pattern was drilled and then the centre chiselled out. Note candling of upper few centimeters of ice.

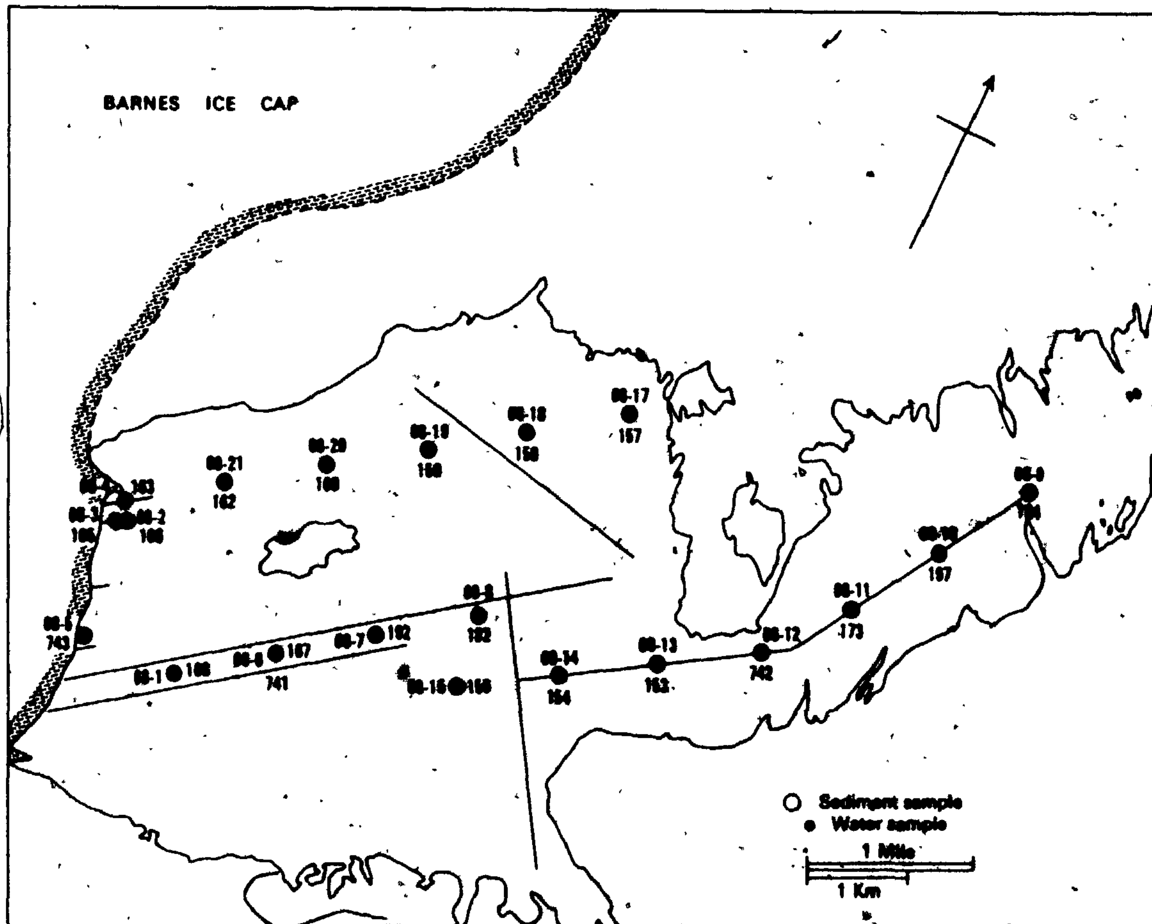


Figure 3 Locations in Generator Lake at which sediment samples and water samples were obtained.

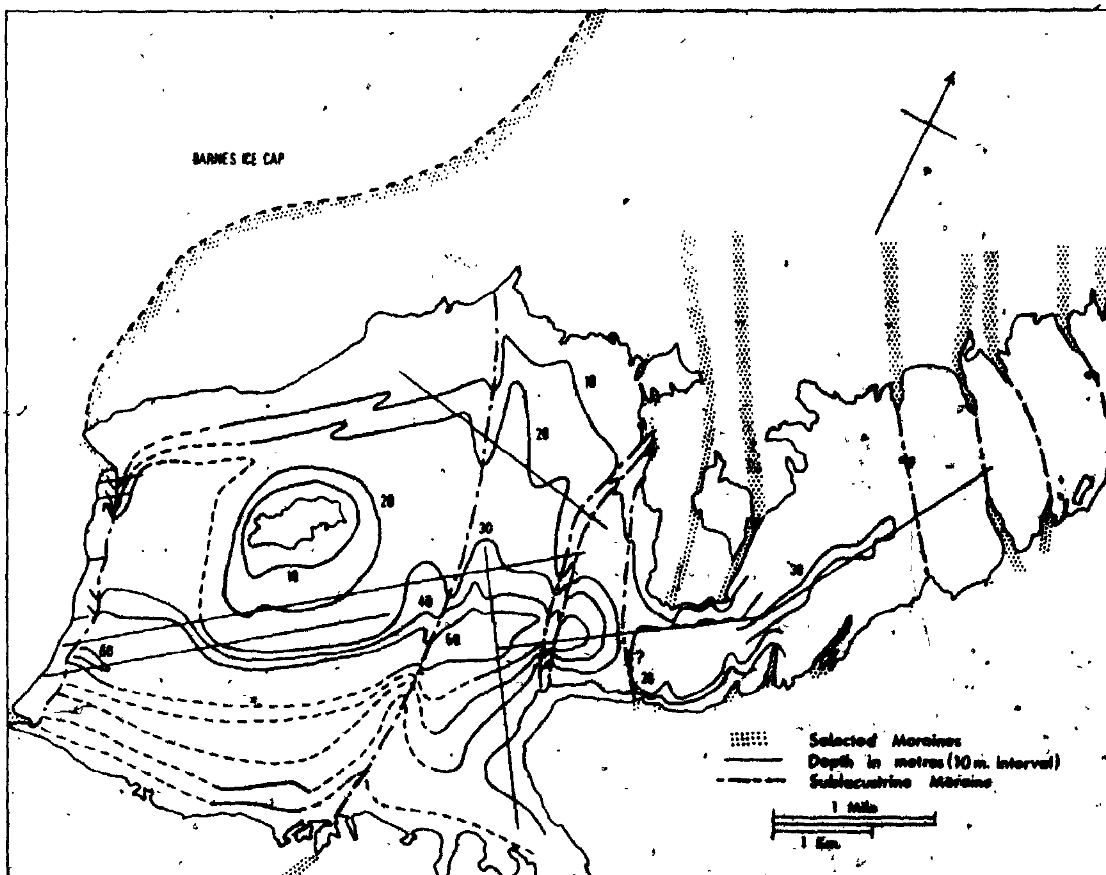


Figure 4 Bathymetry in metres of Generator Lake as interpreted (by DMB) from the echo sounder data, and related geomorphological investigations.

Plate 3 Preparing to lower the Dietz-Lafond grab sampler. The power winch was used for retrieval at slow speed.

DATE (1968)	WATER SAMPLE (No)	DEPTH (m)	LAB. TEMP (°C)	COND (m/mho)	HARD	Ca	Mg	Na	K	Cl	NO_3	SiO_3
July 23	153	37	24.5	16.9	3.0	0.75	0.26	1.3	0.3	1.5	0.14	1.8
July 23	154	44.5	24.5	13.4	2.0	0.50	0.20	1.0	0.3	1.0	0.18	1.8
July 23	156	33	24.5	15.3	3.5	1.0	0.24	1.0	0.3	0.8	0.14	1.8
July 27	157	16.5	24.5	16.9	2.3	0.50	0.26	1.3	0.3	0.8	0.20	1.8
July 27	158	23.5	24.9	14.7	2.1	0.50	0.22	1.1	0.3	0.8	0.14	1.6
July 27	159	24	24.9	14.9	2.1	0.50	0.22	1.2	0.4	0.7	0.10	1.7
July 27	160	19	24.8	16.8	2.3	0.50	0.26	1.1	0.3	0.7	0.13	1.6
July 27	162	27.5	24.8	16.9	2.2	0.55	0.20	1.1	0.3	0.8	0.11	1.6
July 27	163	16	24.9	16.5	2.1	0.50	0.22	0.9	0.3	0.6	0.13	1.5
July 27	165	20	24.8	14.8	2.3	0.55	0.22	1.0	0.3	0.7	0.14	1.7
July 27	166	30	24.9	14.6	2.1	0.50	0.22	1.0	0.3	0.8	0.17	1.4
July 27	167	52	24.9	15.1	2.2	0.50	0.24	1.0	0.3	1.3	0.29	2.0
July 23	168	47.5	24.9	13.7	2.3	0.52	0.24	1.0	0.3	0.8	0.27	1.7
July 23	173	22	24.9	30.8	2.1	0.50	0.22	2.7	0.3	5.1	0.10	0.9
June 29	182	19	24.8	15.9	2.3	0.60	0.20	1.2	0.3	1.5	0.19	1.8
July 23	184	22.5	24.9	14.7	3.1	0.60	0.40	2.8	0.5	5.0	0.18	1.7
June 28	192	21	24.8	15.9	2.0	0.50	0.20	1.5	0.4	1.6	0.14	2.5
July 23	197	29	25.0	18.9	2.7	0.45	0.40	2.5	0.4	4.1	0.27	2.1
June 22	741	52	25.0	15.8	2.0	0.50	0.20	1.4	0.3	1.0	0.24	1.4
July 23	742	19	25.0	14.4	2.2	0.50	0.24	1.2	0.3	0.8	0.14	1.6
July 18	743	40.5	25.0	14.4	2.2	0.50	0.24	1.2	0.3	0.9	0.15	1.5

TABLE 2

Bottom Temperatures derived from Reversing Thermometers in Generator Lake

Date of BT cast (1968)	BT slide #	Location Sediment Sample # (Fig. 2)	Temperature at bottom by Reversing Thermometer (°C)
June 12	1	1	1.7
June 12	2	2	1.4
June 13	3	3	1.3
June 16	4	4	1.1
June 18	5	5	1.4
June 22	6	6	1.5
June 28	7	7	1.3
June 29	8	8	1.3
July 23	9	9	1.3
July 23	10	10	1.5
July 23	11	10	—
July 23	12	11	1.3
July 23	13	12	1.2
July 23	14	13	1.5
July 23	15	14	1.8
July 23	16	15	1.4
July 27	17	17	1.4
July 27	18	18	1.3
July 27	19	19	1.4
July 27	20	20	1.3
July 27	21	21	1.4
July 27	22	4	—
July 27	23	3	—
July 27	24	2	—
July 27	25	6	—
July 27	26	1	—

Notes: Temperatures in column 4 are rounded to the nearest 1/10 of one degree. Column 3: GSC sediment sample prefix BDA 68- has been omitted for brevity.

THE DATA

At each of a total of 20 stations the following data were collected: a bathythermograph trace, a water sample from immediately above the bottom, reversing thermometer readings at the same depth (Table 2) and, on a second cast, a bottom sediment sample by grab sampler. Six bathythermograph traces were repeat casts, as was one water sample. The water samples were analysed through the cooperation of the Water Quality Laboratory of the Inland Waters Branch and showed that the older northeast part of the lake basin (east end of profile 7) contained higher amounts of some dissolved salts than the western portion of the lake (Table 1). The bathythermograph traces are shown in section III of this report.

Plate 4 Bathythermograph and Knudson bottle beside sampling site. Two reversing thermometers are attached to the bottle.

GEOLOGICAL SURVEY OF CANADA

Tasiujaq Cove, Ekalugad Fiord, Baffin Island

Observers:

R. John Knight
Department of Geology
Queen's University, Kingston, Ontario

Michael Church
Department of Geography
University of British Columbia, Vancouver, B.C.

Division of Quaternary Research and Geomorphology

Plate 5 Aerial view of Tasiujaq Cove.

TASIUJAQ COVE

During the summers of 1967 and 1968, a field party from the Division of Quaternary Research and Geomorphology, Geological Survey of Canada, engaged in studies of recent alluvial sedimentation at the head of Ekalugad Fiord (frontispiece) on the Home Bay coast of Baffin Island (GSC project 680042). In conjunction with this survey, investigations were made of the bathymetry and bottom sediments in Tasiujaq Cove (Fig. 5).

Tasiujaq Cove is the shallowest of three basins in the Sarvalik arm of Ekalugad Fiord (Fig. 6). While the sill at the outer end of the arm is bedrock, the next may be morainic, like that separating Tasiujaq Cove. This moraine was deposited underwater about 6100 years ago (Andrews, 1969) by the last Pleistocene ice tongue to occupy the valley. By 5700 years B.P. the ice had retreated to the head of Ekalugad valley (6.8 km upvalley) and alluvial deposition began into the fiordhead from this position. The moraine probably emerged about 4500 years ago, so that the circulation between the cove and the Sarvalik Fiord has been restricted through most of its history.

The delta being deposited at the distal end of the alluvial outwash plain continues to encroach slowly into the cove; it is now about 2.25 km long and 1.85 km wide at its widest. The deepest point, determined from soundings, is 70 m and the mean depth is 39.7 m; the surface area is 3.61 km^2. The channel connecting the cove with the fiord is only 3 m deep and is floored by large boulders that are lag deposits from the eroded moraine.

Freshwater enters at the head between late June and early September from three rivers with maximum runoff during the snowmelt period in early July and during summer rainstorms. The average daily influx of water during the 1967 season was 4.24×10^6 m^3 with a maximum daily influx of 16.79×10^6 m^3 on July 14. In the low runoff season of 1968 only about 1/4 as much runoff was observed.

In 1967 soundings of the delta foreslope were made by lead line, with positions fixed by simultaneous theodolite observations from shore. In 1968 echo sounding was carried out. Positions were fixed periodically along the boat track from sights on a series of "range" markers on shore. The bathymetric map (Fig. 5) was subsequently drawn by interpolating contours amongst the track profiles plotted on the map.

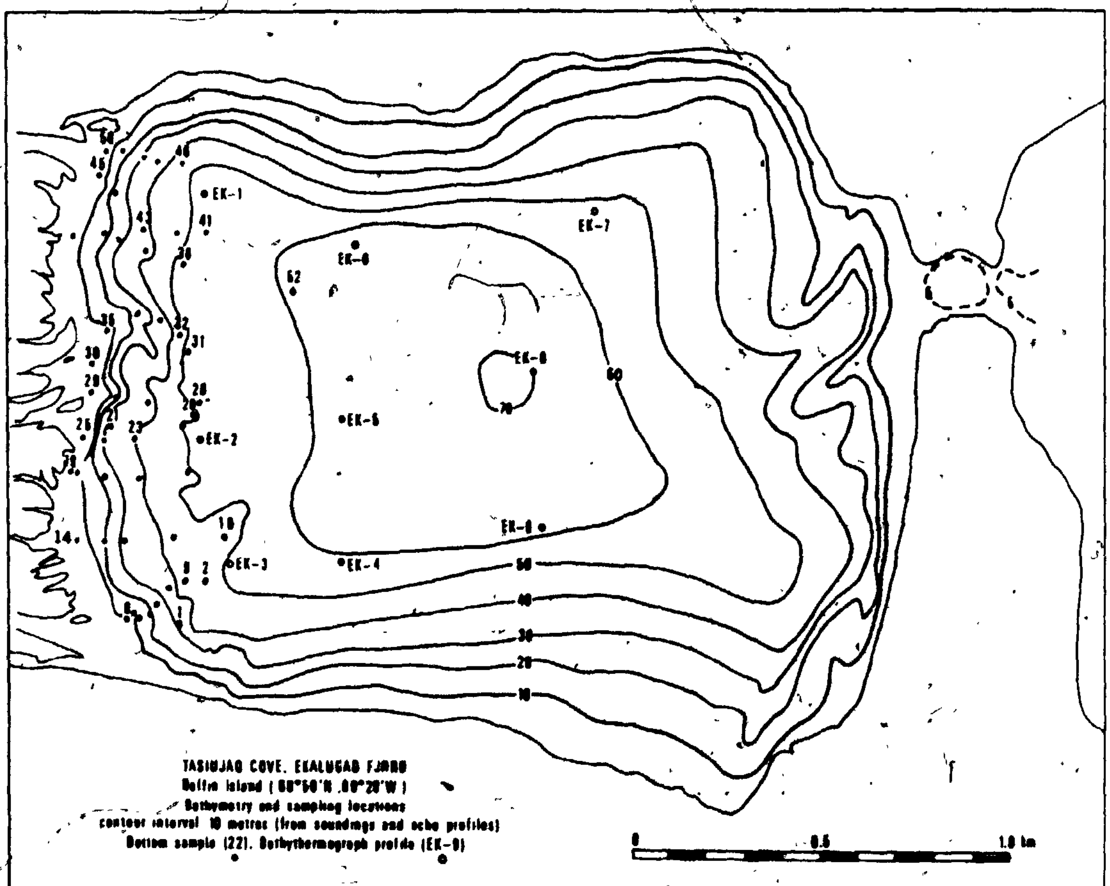


Figure 5 Location of stations within Tasiujaq Cove indicating the bathymetry and sampling locations.

A tide gauge with a pressure transducing sensor was set up in 1967 near the northwest corner of the cove, and records maintained between July 22 and August 15. The tide appeared to be basically diurnal, but with a double high tide often occurring. High tide characteristically occurred early in the morning. For the period of record the average daily tidal range was 0.84 m: the maximum was 1.03 m and the minumum, 0.58. Without considering freshwater discharge from the rivers, the average tide would require a daily movement into and out of the cove of 3.05×10^6 m^3 of water. In view of the data presented above, it appears that the stream discharge may affect the water levels during periods of significant runoff. Water level data are available from the observers.

The cove is ice covered for about 9 to 10 months of the year, opening only between late July and the end of September. The channel at the moraine may or may not freeze over; in the winter of 1966-67 the tidal exchange apparently maintained open water there.

Temperature profiles were obtained using a bathythermograph at the nine stations shown on Fig. 5. Two reversing thermometers were used to determine the water temperature at water sampling positions. The temperatures reported in Table 3 represent the mean of the two thermometers, after corrections. The mean discrepancy between the two thermometers was 0.03°C and the maximum discrepancy was 0.06°C. Surface water temperatures were read with an ordinary thermometer divided in 1/10ths °C.

All the results reported in Table 3 were obtained on August 20, 1968.

Water samples were taken in each BT profile, from two to five samples being taken in each. Surface samples were dipped from the water and the others were obtained using a Knudsen bottle.

Analyses were undertaken at the Industrial Waters Laboratory, Water Quality Division of Inland Waters Branch. Complete analysis was carried out on 13 of the samples and the specific conductance determined for the remainder with a multiple range conductivity meter.

All data are given in parts per million. Salinity is obtained by taking 10^{-3} x the sum of major constituents as given in the table. "Total hardness" is the total of hardness producing ions expressed as $CaCO_3$ equivalents ppm. Here, Ca^{++} and Mg^{++} are considered.

TABLE 3

DATA FROM CHEMICAL ANALYSIS OF WATER SAMPLES TAKEN AT TASIUJAQ COVE, EKALUGAD FIORD, BAFFIN ISLAND, N.W.T.

St No	Depth (M)	(°C)	Specific Conductance	Total Hardness	Ca	Mg	Na	K	Cl	NO_3	Sio_2	Sum of Major Constituents
1	sfc.	5.2	1,280	111.0	8.0	22.0	182	10.0	332	0.06	0.70	554.76
	13.7	-0.63	45,600	5,830	340	1,210	9,400	480	16,600			28,021
	27.7	-1.37		no data								
	41.5	-1.40	48,200									
	54.8	-1.40	47,700									
2	sfc.	5.2	982	90.3	6.5	18.0	145	7.8	265	0.05	0.60	442.95
	13.7	-0.84	48,400	6,060	350	1,260.	10,000	510	18,000			30,120
	27.7	-1.33	48,700	6,061	350	1,260	10,000	510	18,150			30,271
	41.5	-1.46	48,700	6,060	350	1,260	10,000	510	18,150			30,270
	56.9	-1.39	48,200	6,020	350	1,250	10,000	510	18,000			30,110
3	sfc.	5.2	790	57.8	5.0	11.0	92	5.2	170	0.00	1.20	284.40
	13.7	-0.77	48,600	6,090	360	1,260	10,000	510	18,000			30,130
	27.7	-1.37	48,200									
	41.5	-1.46	48,200									
	53.9	-1.44	48,200									
4	sfc.	5.2	1,000									
	58.4	-1.42	48,200	6,090	360	1,260	10,000	510	18,150			30,280
5	sfc.	5.2	690	51.9	5.0	12.0	97.5	5.4	186	0.05	1.10	307.05
	61.6	-1.44	48,700									
6	sfc.	5.2	1,110									
	64.7	-1.41	48,200									
7	sfc.	5.5	1,120									
	53.8	-1.43	50,000									
8	sfc.	5.3	1,505	65.8	8.2	11.0	222.0	10.5	418	0.05	2.30	672.05
	18.2	-1.11	48,200	6,060	350	1,260	10,000	510	18,050			30,170
	36.8	-1.38	48,300									
	55.4	-1.45	49,000									
	69.4	-1.36	48,200									
9	sfc.	5.2	1,440									
	63.1	-1.41	48,200									

The samples were stored between August 20, 1968 and February 17, 1969, when they were analyzed. They were retained in sealed, glass sample bottles during this period. Three freshwater samples analyses of water from the inflowing rivers are appended for comparative purposes.

The bottom samples were obtained using a Dietz-Laford grab sampler. Mechanical analysis was carried out in the Sedimentology Laboratory of the Division of Quaternary Research and Geomorphology, Geological Survey of Canada. Dry sieves were used to 0.44 mm, after which pipette analyses were carried out as necessary. Table 4 shows the results.

Little oceanographic work has been done in the Baffin Island fiords. Elsewhere fiords have been studied in considerable detail. The interest of the present investigations lies in the possibility that Tasiujaq Cove could, in the future, become isolated from the sea, under the aegis of continued post-glacial uplift, and form a lake.

R.J. Knight
M. Church

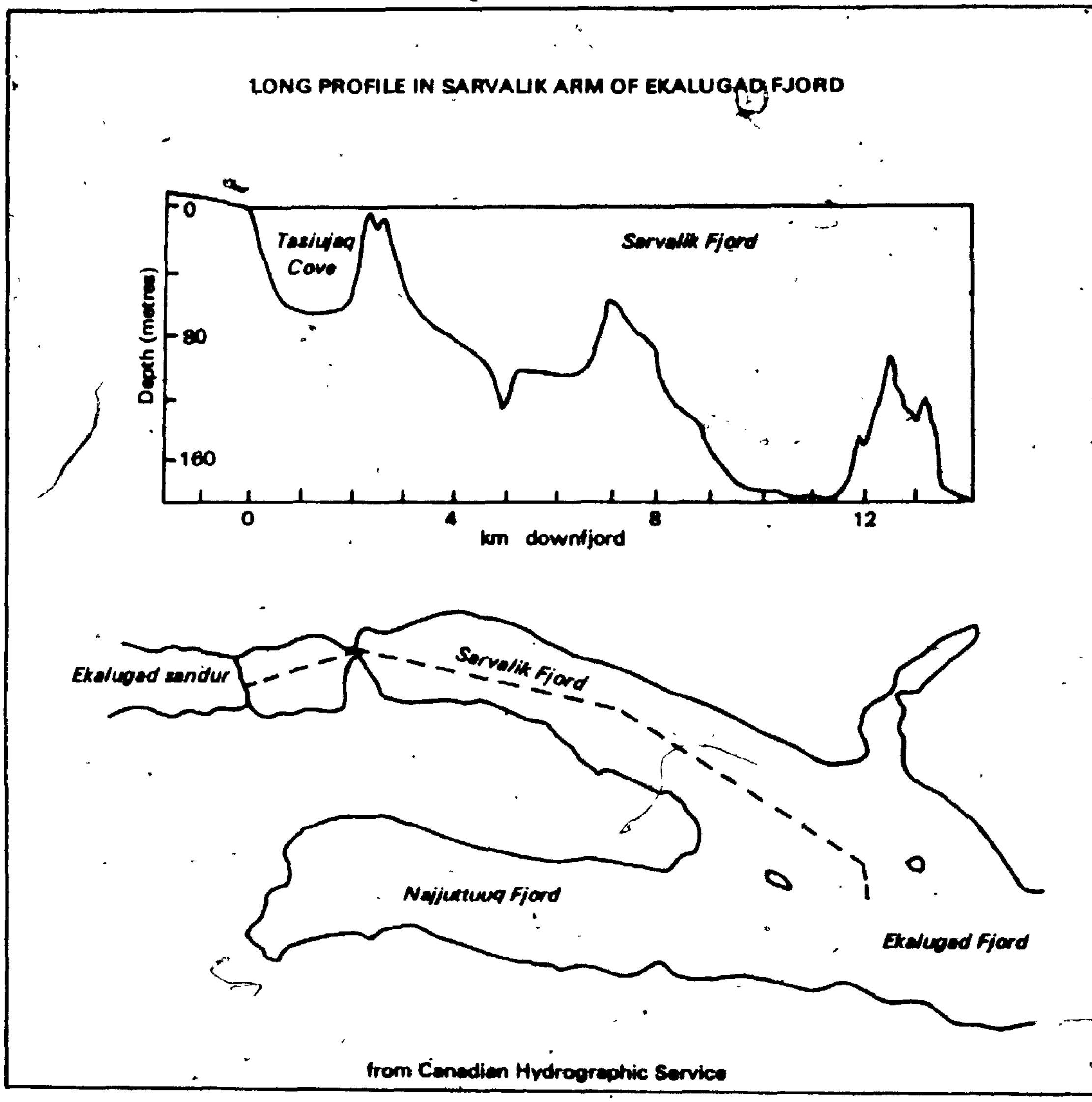


Figure 6 Place names and depth profile in the seaward approach to Tasiujaq Cove.

TABLE 4

Grain Size Data of Bottom Samples

Sample No.	Water Depth (metres)	Percentage of Total			Percentage of Fines		
		32-8 mm	8-2 mm	<2 mm	Sand	Silt	Clay
01	32.56			100.0	62.05	31.66	6.29
02	44.37			100.0	70.31	24.54	5.15
03	37.86			100.0	71.79	22.12	6.09
04	26.20			100.0	62.98	30.92	6.10
05	19.85	66.9	4.4	28.4	91.23	6.18	2.59
06	15.87	70.2	11.5	18.1	93.51	6.49	
07	7.02	17.4	34.9	47.5	98.88	1.12	
08	1.51	59.7	14.7	25.3	99.58	0.42	
09	46.91			100.0	63.43	30.61	5.96
10	46.44			100.0	58.49	34.73	6.78
11	42.92			100.0	40.15	49.91	9.94
12	31.78			100.0	45.39	44.39	10.22
13	20.34	2.0	5.4	92.3	55.06	36.22	8.72
14	7.15	38.3	5.0	56.4	67.28	27.02	5.70
15	49.26			100.0	21.06	66.03	12.91
16	37.45			100.0	44.95	46.32	8.73
17	26.37			100.0	49.28	43.25	7.47
18	12.64			100.0	38.49	50.85	10.66
19	4.24		0.3	99.6	66.85	26.91	6.24
20	49.73			100.0	48.55	43.22	8.23
21	34.63			100.0	58.84	35.44	5.72
23	40.13			100.0	64.36	30.55	5.09
24	33.67		0.3	99.6	87.24	9.59	3.17
25	3.54	9.8	19.9	70.1	99.99	0.01	
26	53.83			100.0	35.53	56.67	7.80
27	36.30			100.0	69.67	25.52	4.81
28	11.21			100.0	85.40	11.39	3.21
29	1.19		0.5	99.4	97.17	1.21	1.62
30	1.61	0.9	18.1	80.8	99.80	0.20	
31	53.04			100.0	46.41	47.02	6.57
32	40.08			100.0	57.80	36.69	5.51
33	31.69			100.0	50.90	42.61	6.49
34	18.73			100.0	52.65	40.26	7.09
35	1.58	0.5	5.8	93.5	94.93	3.24	1.83
36	54.53			100.0	50.42	42.95	6.63
37	33.19			100.0	58.76	35.91	5.33
38	25.56			100.0	58.12	35.96	5.92
39	16.41			100.0	47.93	45.33	6.73
40	1.58		0.2	99.7	91.99	5.93	2.08
41	53.56			100.0	49.85	44.20	5.95
42	45.94		0.1	99.8	58.23	34.17	7.60
43	36.80		0.1	99.8	89.53	9.18	1.29
44	26.29		14.4	85.5	84.34	13.75	1.91
45	5.57		0.1	99.8	61.26	29.60	9.14
46	48.24			100.0	67.24	28.98	3.78
47	33.72			100.0	62.89	32.39	4.72
48	32.95			100.0	87.83	10.30	1.87
49	16.96	6.8	2.6	90.2	81.17	16.52	2.31
50	5.53	9.0	0.1	90.8	84.78	13.39	1.83
52	61.74		0.2	99.7	28.73	60.06	11.21

Plate 6 Boat and equipment used in the collection of the field data at Tasiujaq Cove.

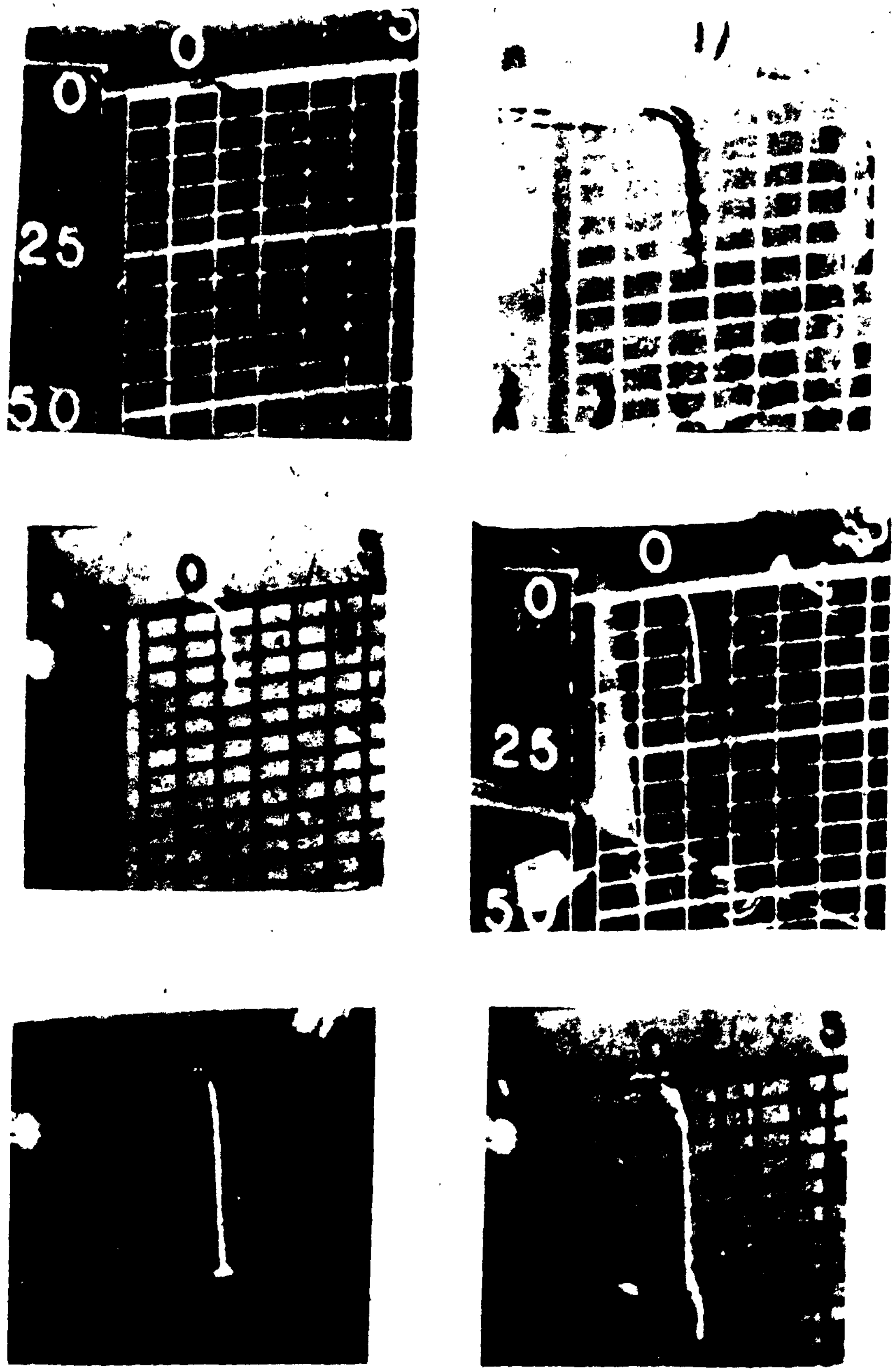
Plate 7 View from North River estuary looking into Tasiujaq Cove.

ACKNOWLEDGEMENTS

Agencies of the Department of Energy, Mines and Resources which made equipment available included the Canada Centre for Inland Waters, the Polar Continental Shelf Project and the Canadian Hydrographic Service; staff of the Marine Sciences Branch assisted and encouraged the preparation of the data record. Wayne Brydges of Atlantic Helicopters kindly volunteered both time and technical skill in assisting with part of the sampling programme.

GENERATOR LAKE

The number on each refers to the station number of Table 1 and the locations of Figure 3.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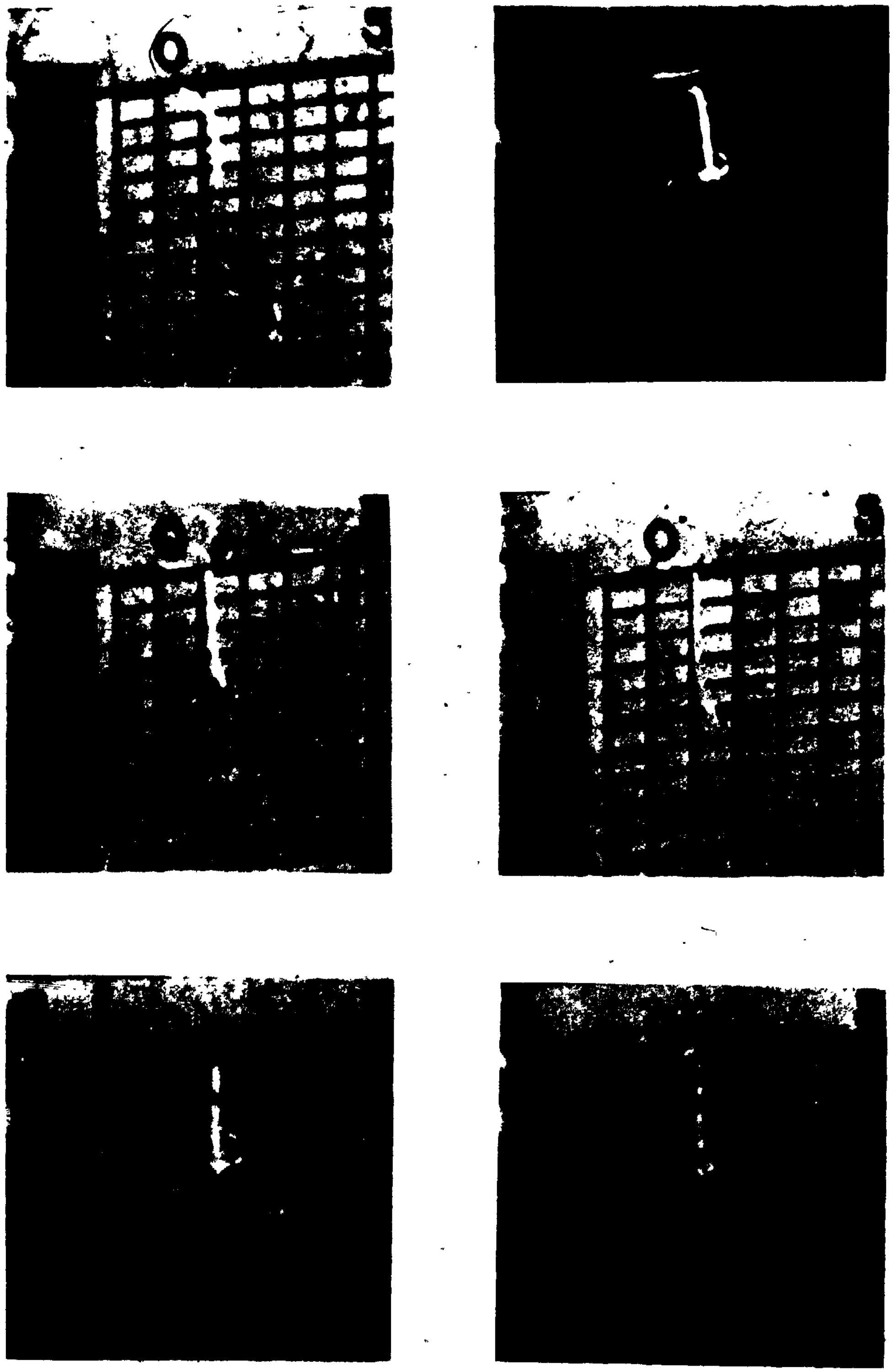
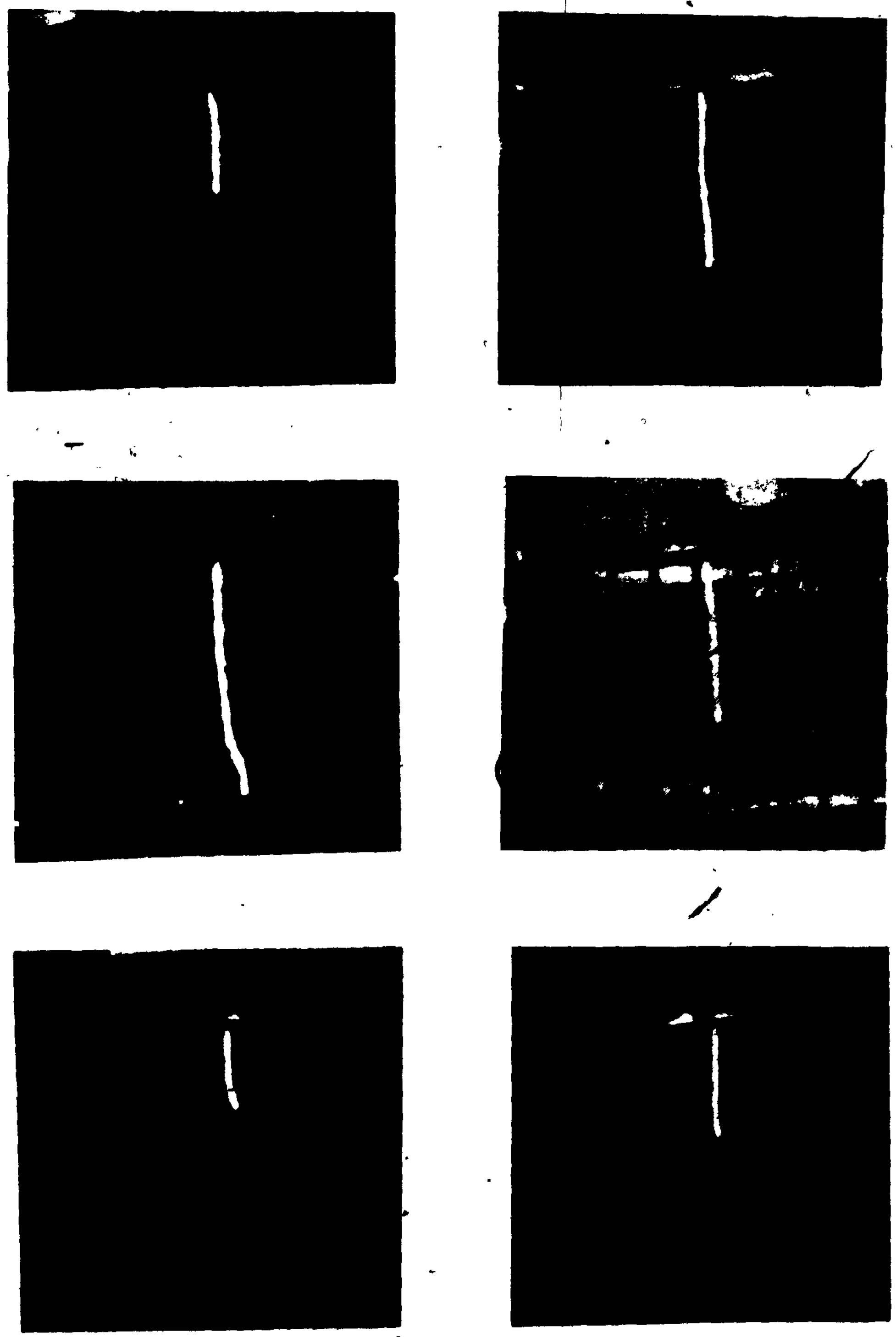
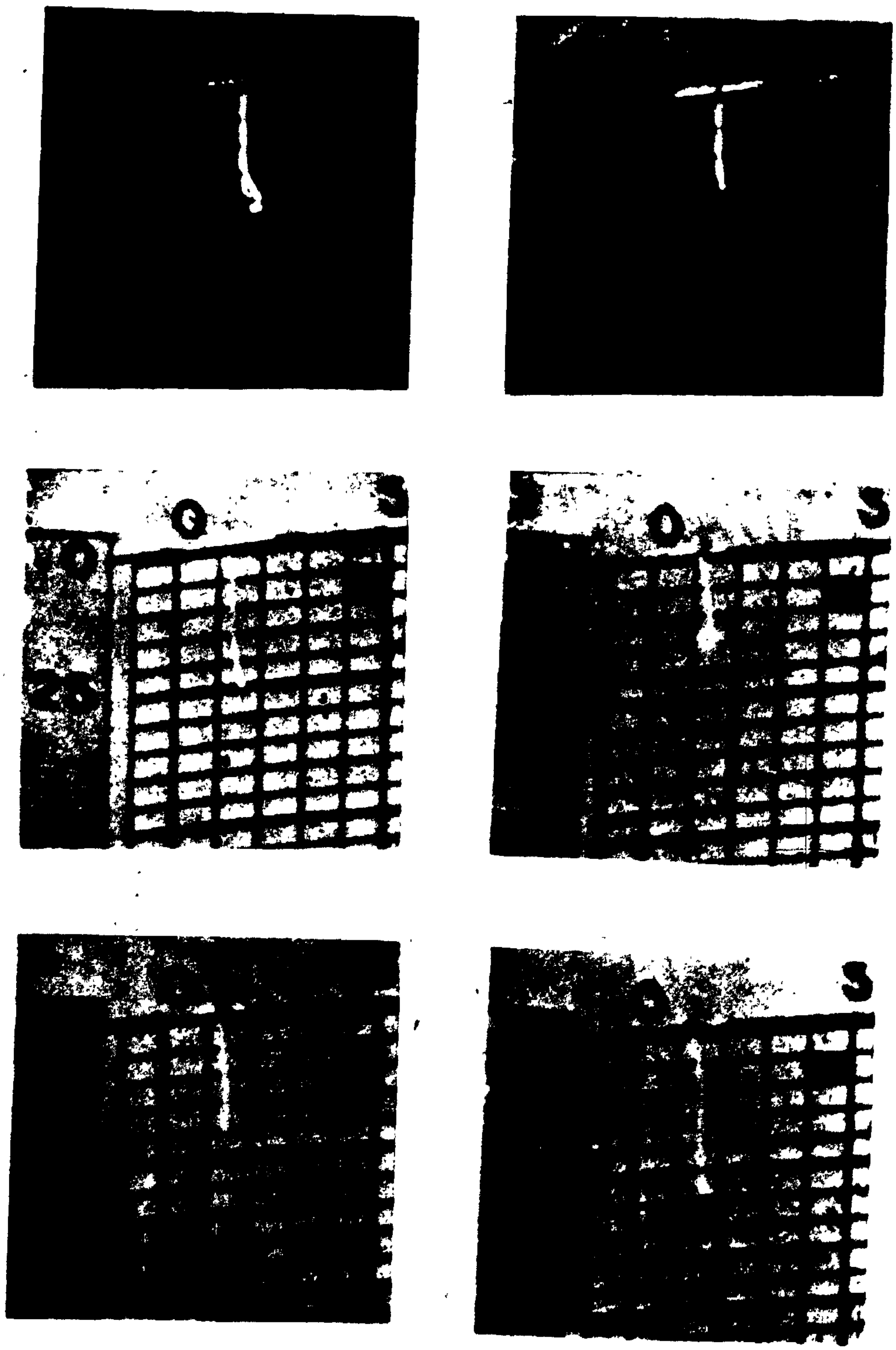
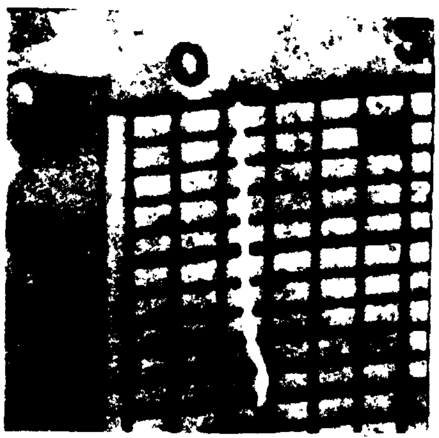
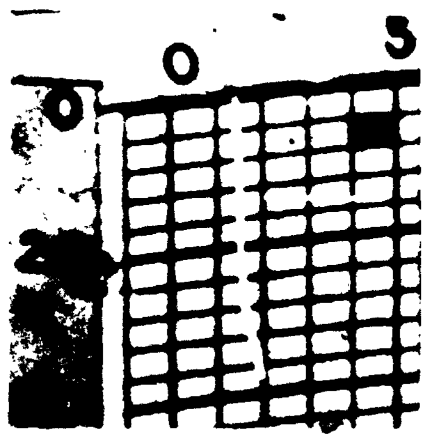




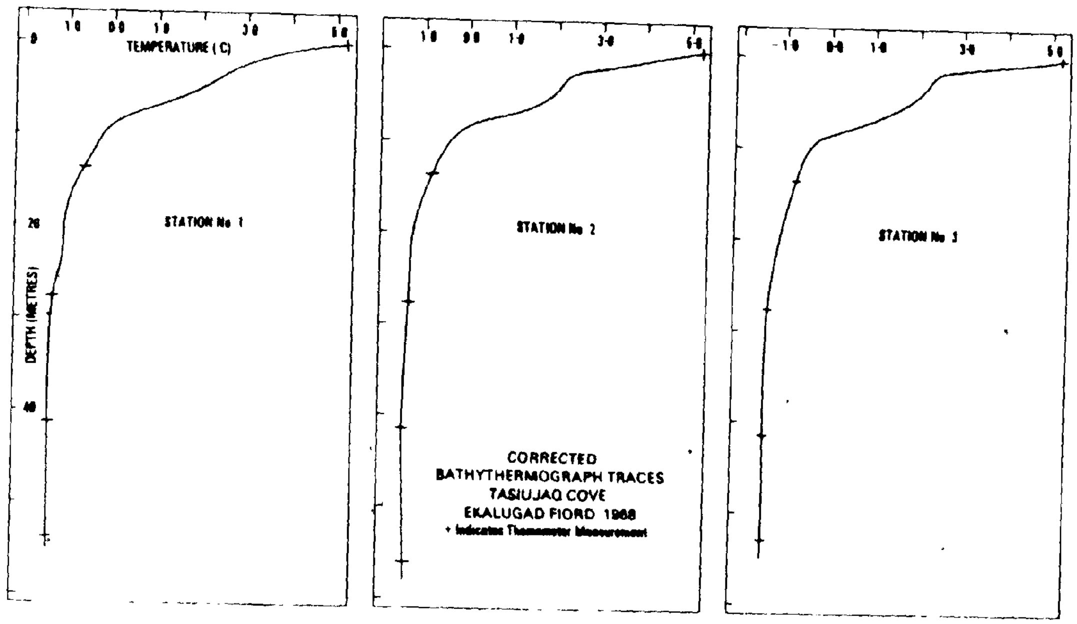




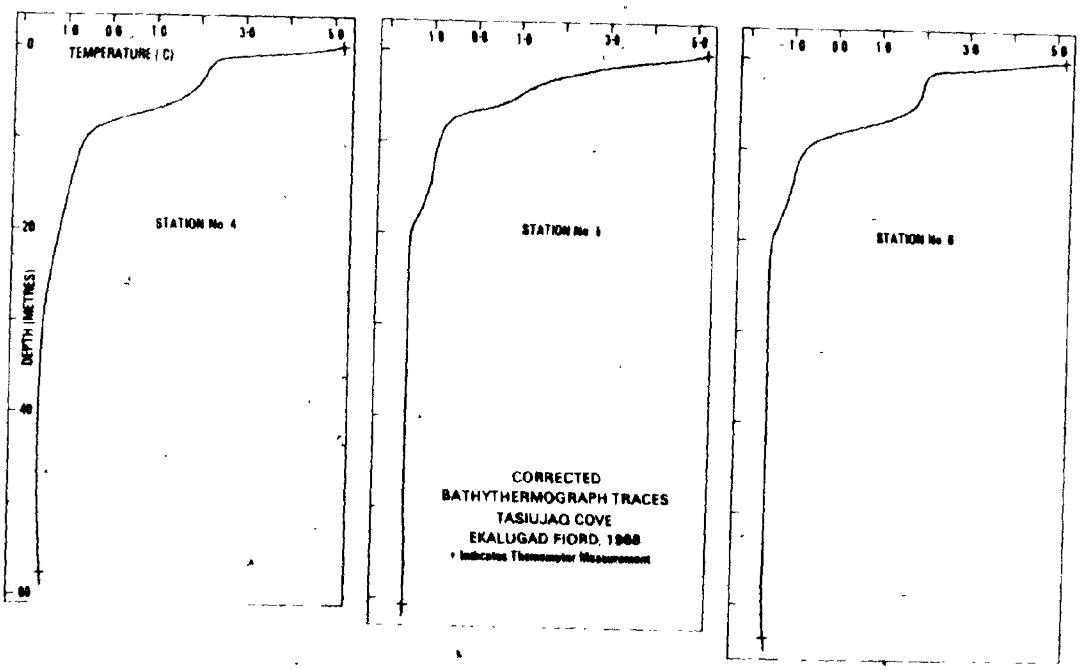


TASIUJAQ COVE

The number on each refers to the station number of Table 3 and of Figure 5.



CORRECTED
BATHYTHERMOGRAPH TRACES
TASIUJAQ COVE
EKALUGAD FIORD 1968
+ Indicates Thermometer Measurement



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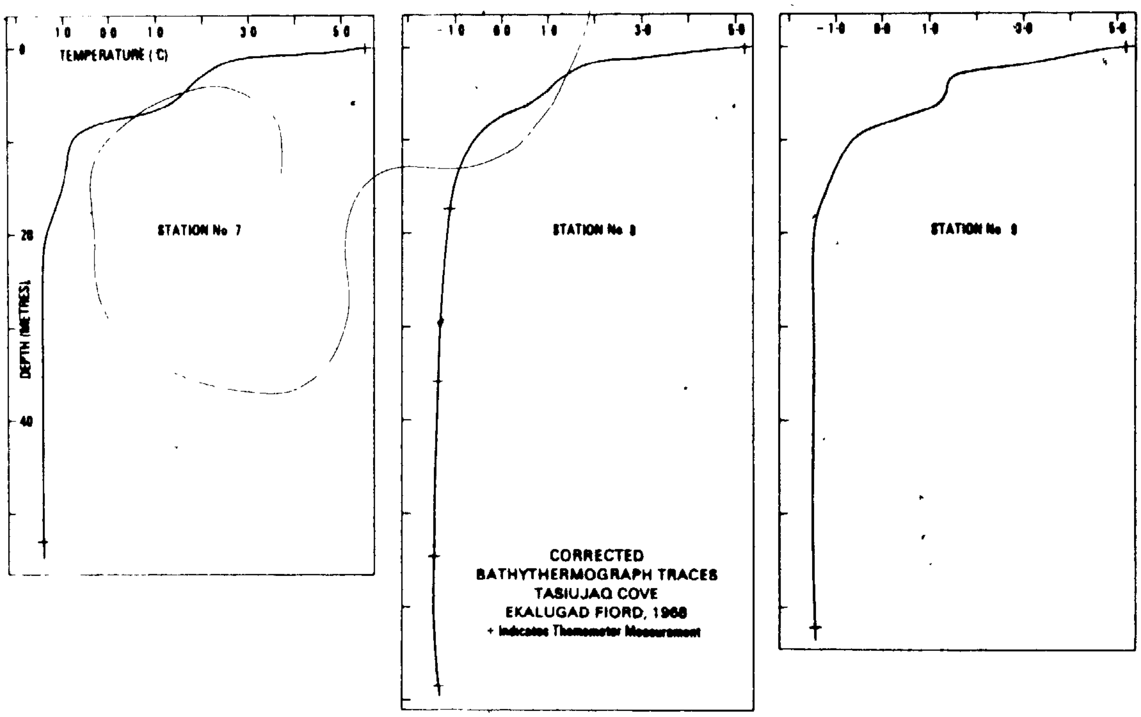

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+ Indicates Themometer Measurement

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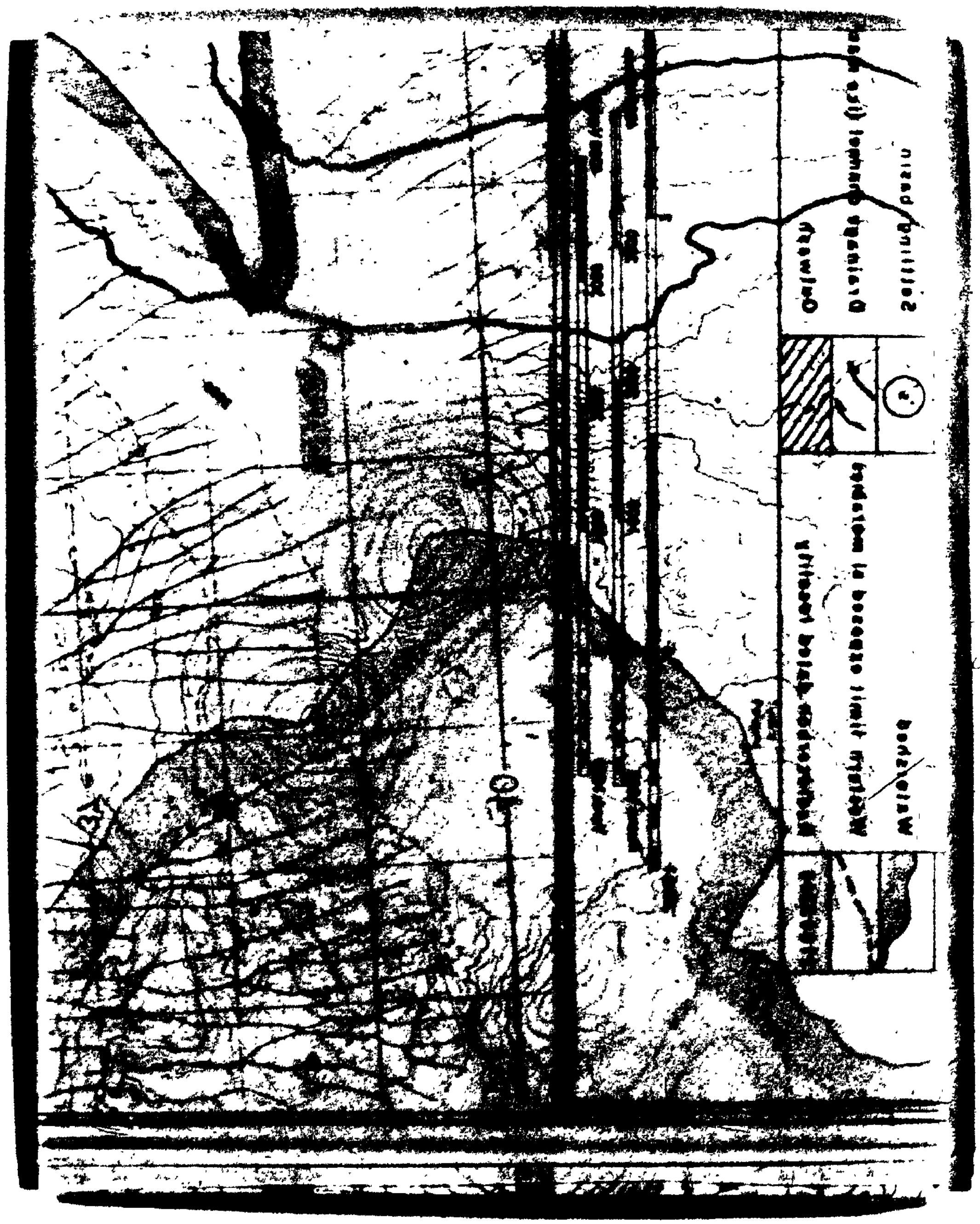


Figure 4-2: Map of Generator Lake basin showing the localities, sublacustrine contours and their interpretation and spatial relationships of features.

POCKET ITEM

GENERATOR LAKE

DISTRICT OF FRANKLIN

NORTHWEST TERRITORIES

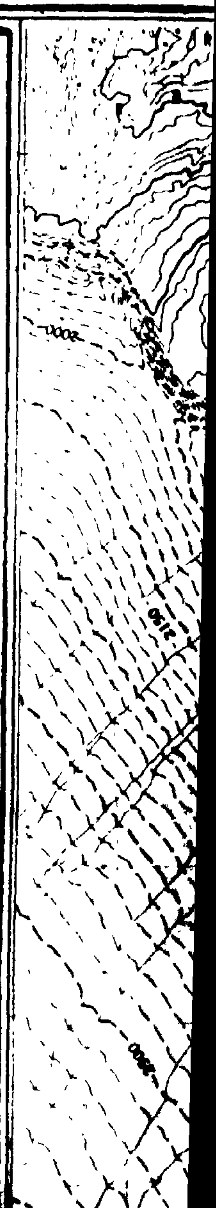
1 of 1

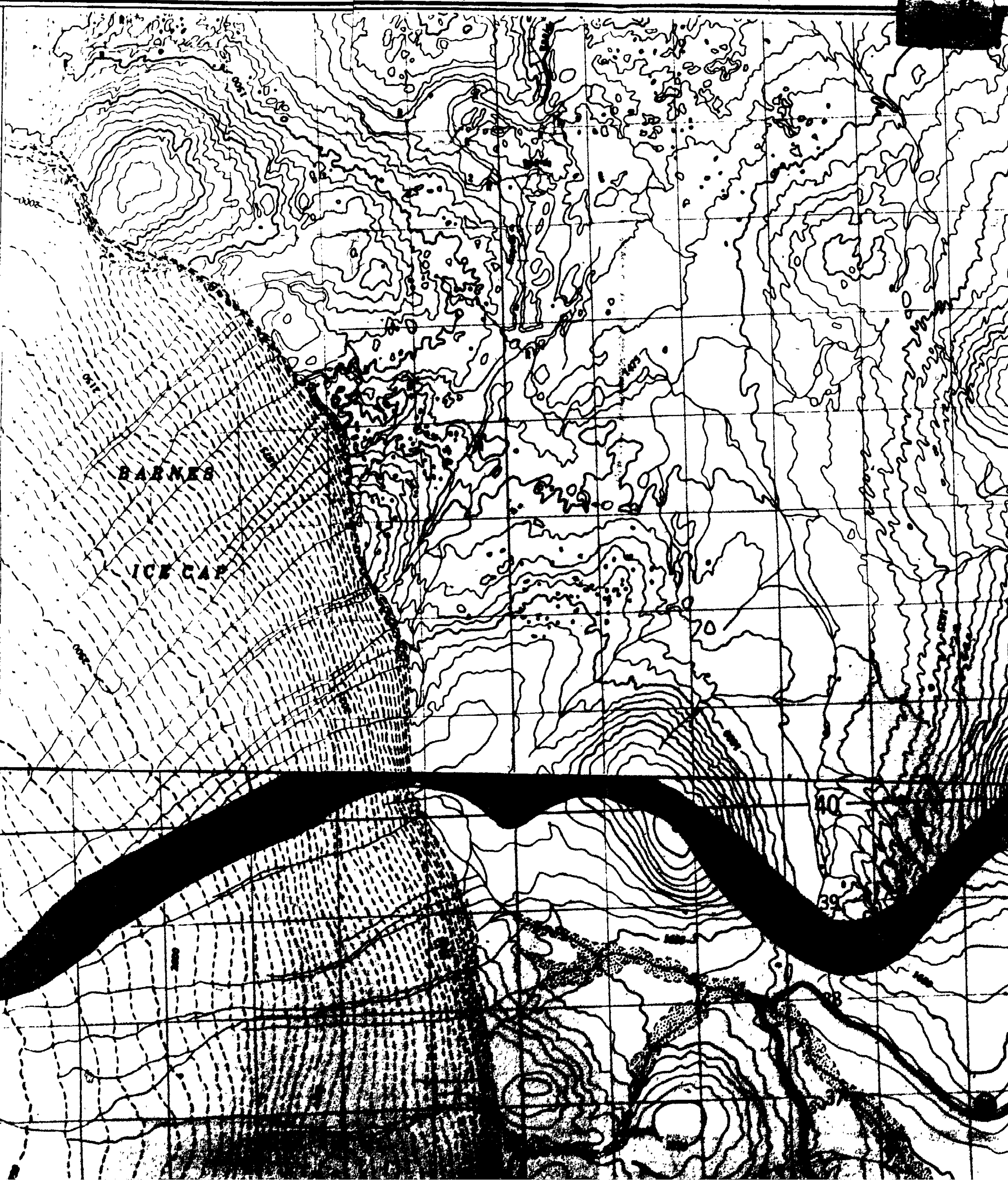
▶	Stream gauging site		Overflow channel
(33)	Beach pebble sample site	DP	Delta profile
	Water washed bedrock	SP	Shoreline profile
C	Till volume cross section	TF	Till fabric site
	Sublacustrine contours (metres)		Sounding profile
	76 m shoreline	d d	Drainage divergence
	Watershed	(s)	Settling basin
	Western limit exposed sl moraines		Drainage channel (ice marginal)
2180±240	Radiocarbon dated locality		Outwash

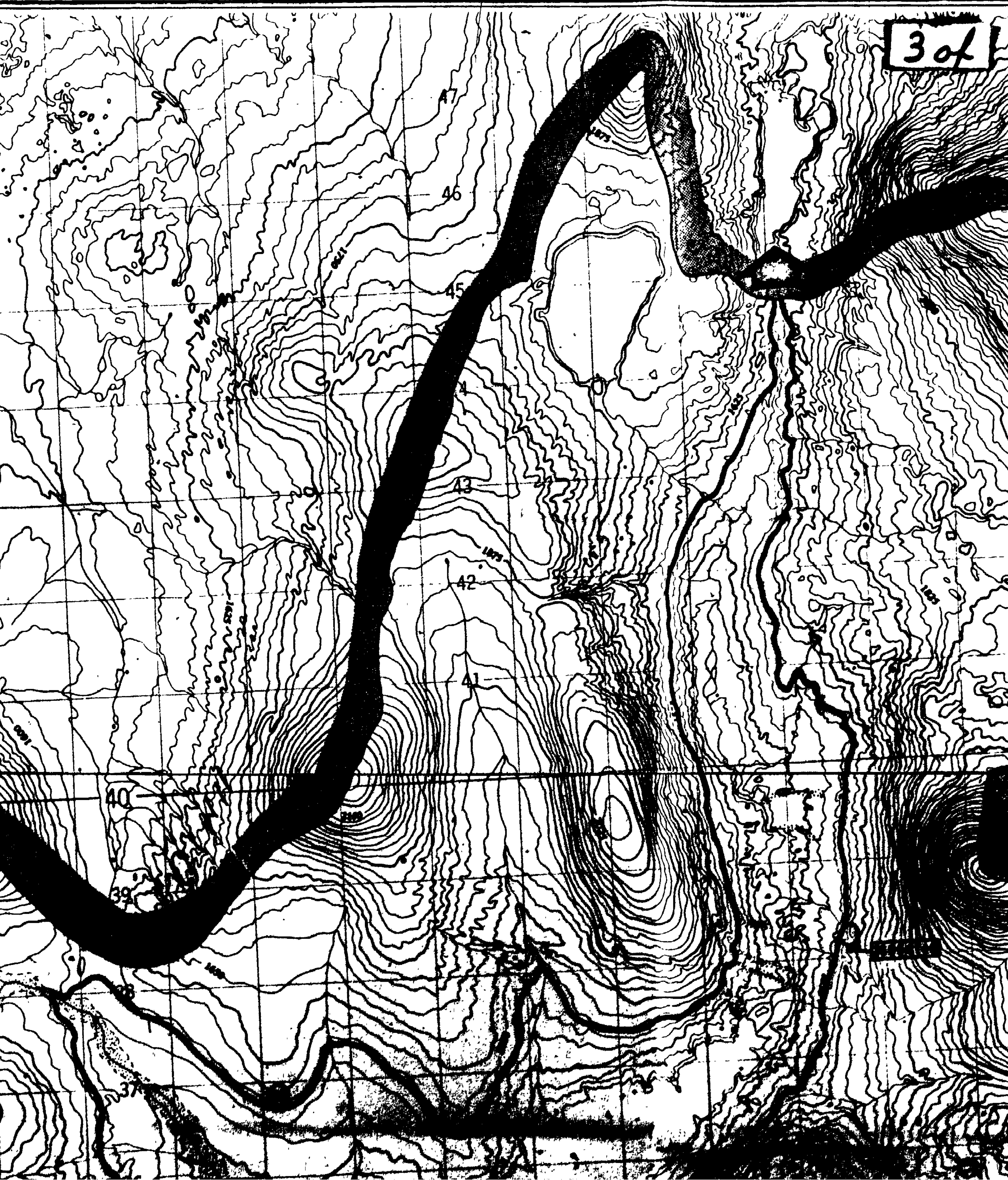
Miles 1 0 1 2 3 Miles

Metres 1000 0 1000 2000 3000 4000 Mètres

Yards 1000 0 1000 2000 3000 4000 Verges




BARNES
ICE CAP
40
39
38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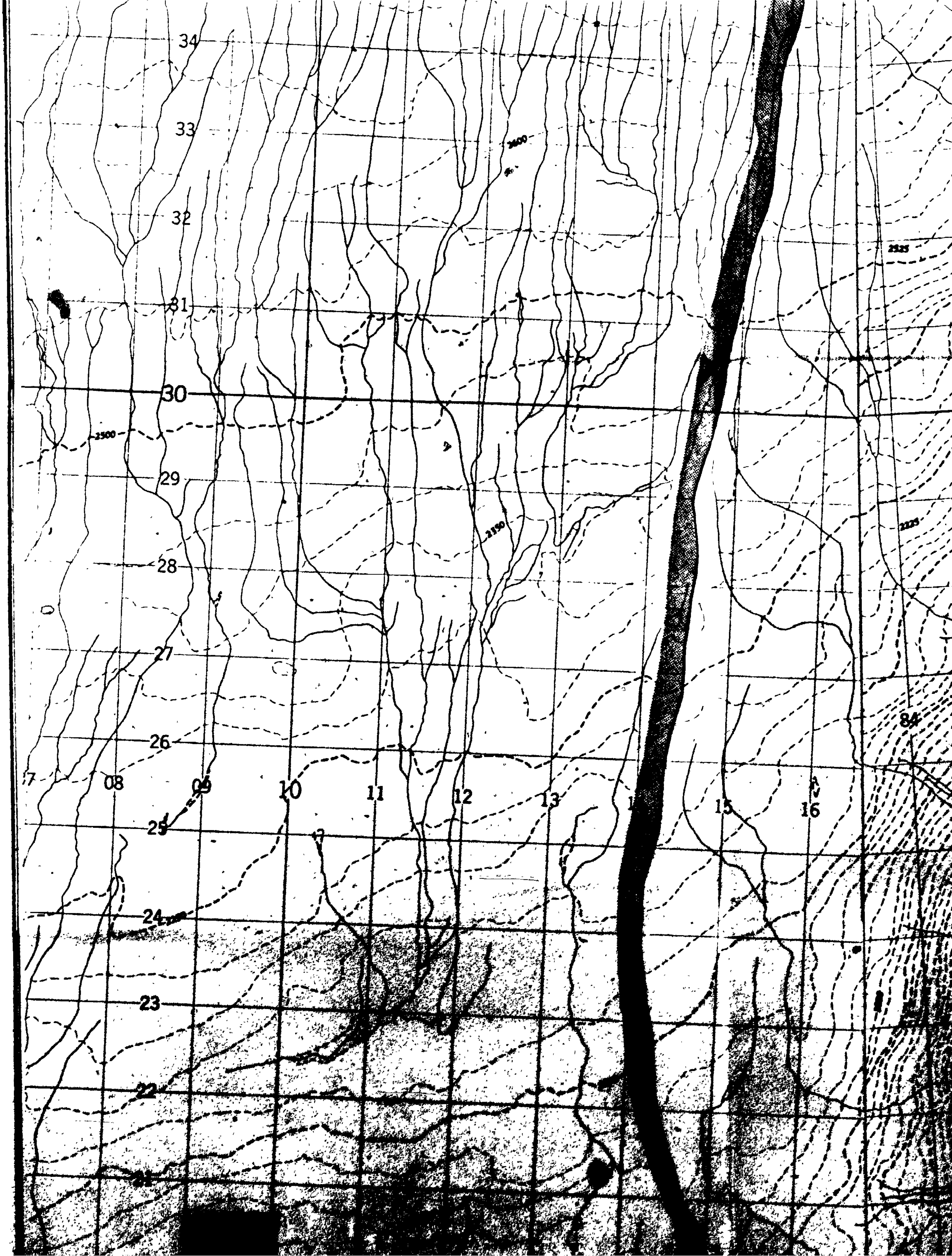
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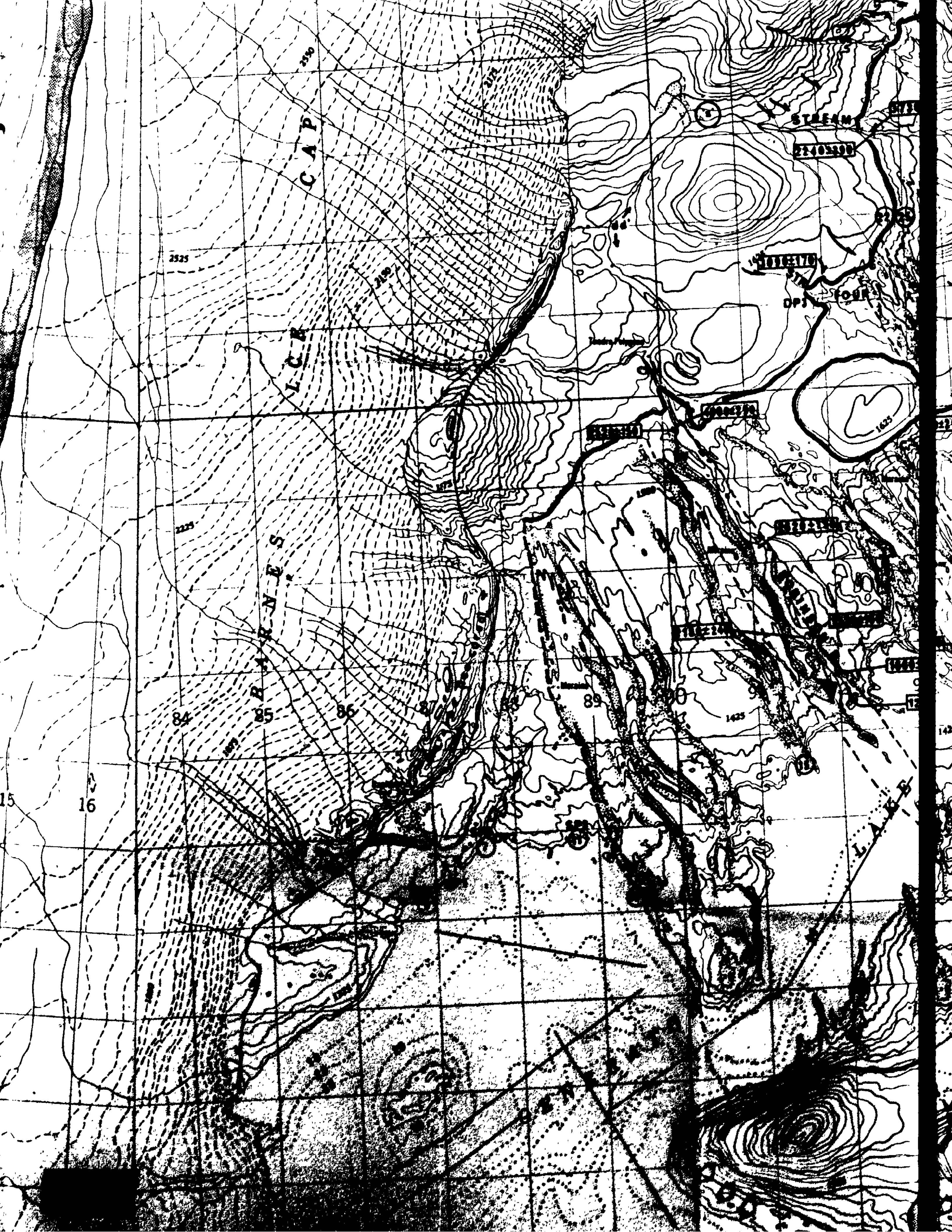

B A R N E S
I C E
C A P
STREAM
FOUR
LAKE
2525
2225
1425
1625
84
85
86
89
15
16


STREAM
FOUR
DP3
2240±190
RIVER
B A F F I
STREAM ONE
34
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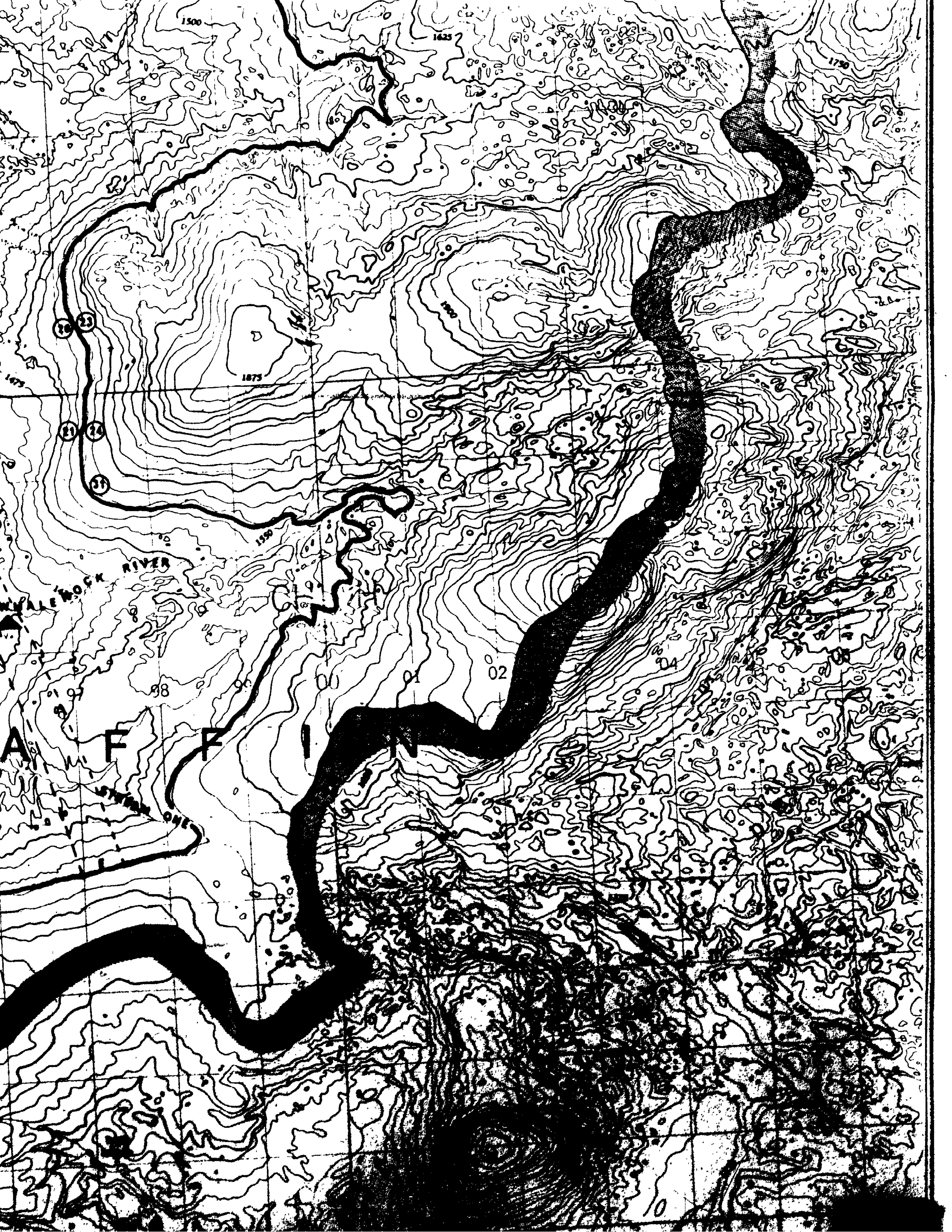
27
93
94
26
97
98
01
02
22
1625
1875
1550



1500
1625
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RIVER
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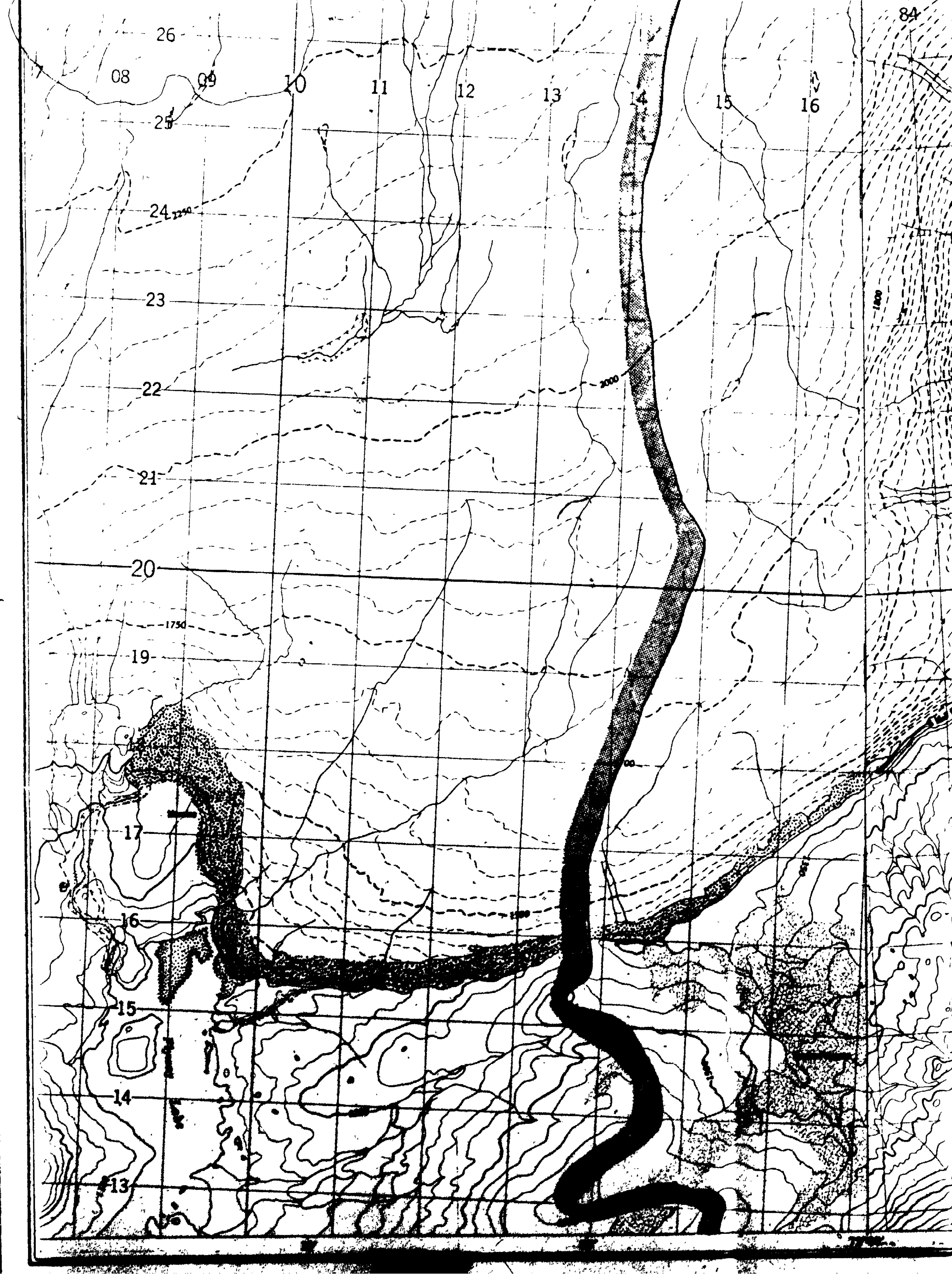
A F F I N



84
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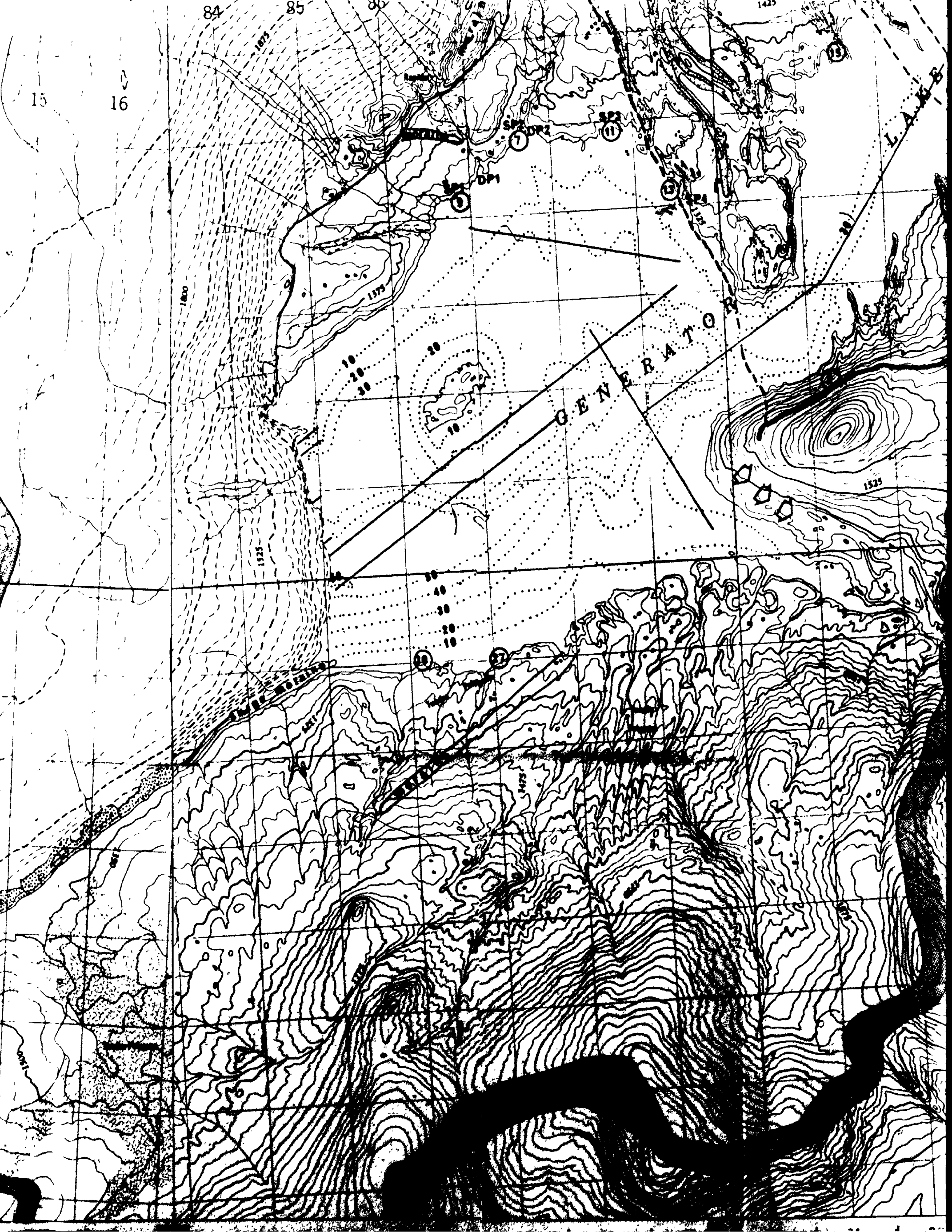

84
85
15
16
GENERATOR
LAKE
SP3
DP2
DP1
SP4
1375
1800
1525
1425


B A F F I
LAKE
ATOR
STREAM ONE
CLYDE RIVER
1425
1525
1725
1325
1375
26
22


STREAM ONE