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John T. Andrews

PRESENT AND PALEO-CLIMATIC INFLUENCES ON THE GLACIERIZATION AND DEGLACIERIZATION OF CUMBERLAND PENINSULA, BAFFIN ISLAND, N.W.T., CANADA

J. T. ANDREWS AND R. G. BARRY

With Contributions From

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FINAL REPORT to the U.S. Army Research Office, Durham, North Carolina
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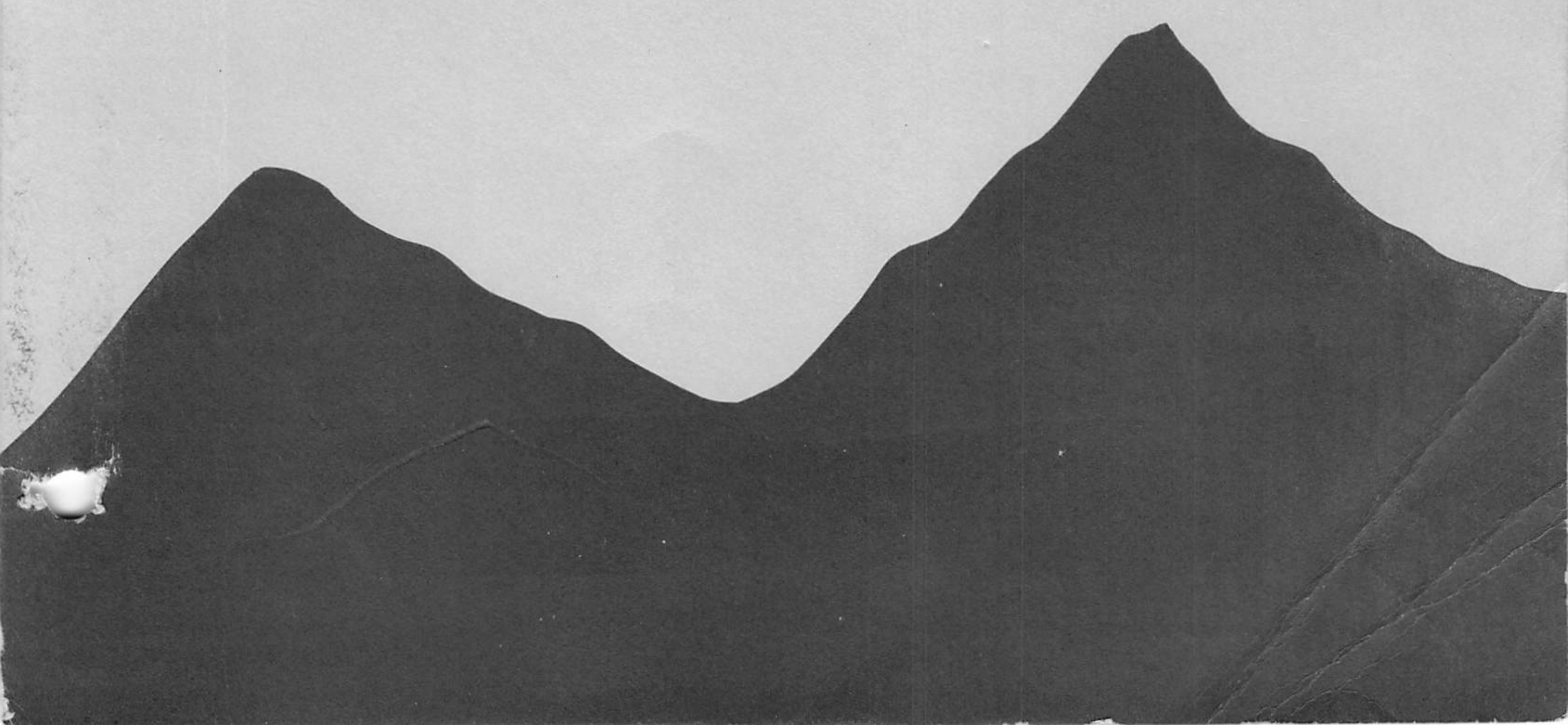

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Final Technical Report

J.T. Andrews and R.G. Barry

1972

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13. ABSTRACT
Cumberland Peninsula lies between latitudes 65° and 68°N and longitude 61° and 68°W. It is dominated by the 6,000km^2 Penny Ice Cap but there are many other local ice bodies. Maximum elevations approach 2,000m and in its central part the area is extremely alpine in character. To the west relative relief decreases until it gives way to the rolling interior plateau of the island at about 600m a.s.l. The purpose of the research was an interdisciplinary study of present climate and its affect on glacier mass balance at both the micro- and topoclimate level and use of these data to develop a concept of past climate and glaciology for the Peninsula for the latter part of the Quaternary, basing this model on changes in area and volume of Laurentide and local ice bodies over the course of the last 120,000± years. Detailed field work involving micro-meteorology, climatology, glaciology, and Quaternary geology was carried out in the vicinity of the "Boas" Glacier in 1970 and additional research in 1971. Overall mapping of the glaciation limit and local glacier ELA's was carried out for Baffin Island, and corrie floor elevation, orientation and global radiation receipts were derived for features along the northern half of the Peninsula. Our conclusions are that: 1) the area is critically sensitive to small climatic change, 2) that the development of major ice bodies (and the start of the Laurentide Ice Sheet!) is related to increased snowfall which occurs during the early climatic deterioration, 3) thereafter the area becomes increasingly arid and ice volumes decrease (in explaining glacierization, scale factors-spatial and temporal are very important), and 4) a pronounced climatic change during the last decade is associated with an increase in winter snowfall and a decrease in summer temperatures that enforce each other to give the possibility (as in the 1969-1970 mass balance year) of a positive balance over entire glacier surfaces.

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A special work of appreciation is due to the National Air Guard units of Wyoming and Minnesota who flew men and equipment from Denver to Frobisher Bay and return in 1970.

Finally, we acknowledge the help and assistance rendered by the U.S. Army Research Office, Durham, and particularly Drs. A. Dodd and F. Bronner who have helped in many aspects of the grant research reported in this document. At the same time we acknowledge various forms of assistance provided by the Institute of Arctic and Alpine Research, University of Colorado and to Dr. J.D. Ives, for his continual encouragement of the program.

1. INTRODUCTION

J.T. Andrews and R.G. Barry

This report is based on research funded by two grants from the Army Research Office, Durham, North Carolina. The title of the funded grants was "Present and Paleoclimatic Factors Affecting the Glacierization of Northern Cumberland Peninsula, East Baffin Island, N.W.T., Canada" and the grants were awarded to J.T. Andrews and R.G. Barry as the two Principal Investigators in 1969-1971 and 1971-1972.

The project was initiated in 1968 when H. Kryger, University of Colorado, obtained funding to visit two glaciers lying between the head of Narpaing and Quajon fiords, east Baffin Island (Figure 1-1). His visit was intended as a primary reconnaissance survey and he succeeded in establishing a Hoinkes storage precipitation gauge in Smirling Valley and lichen growth stations on the outer moraine of the Akudnirmuit Glacier (Figure 1-1).

The survey and installations were extended in 1969 by Williams and Weaver (see Appendix I) and in 1970 a full-scale expedition was landed on the large lake in Smirling Valley by an Otter aircraft on skis. A total of 16 graduates and faculty were engaged that year in detailed climatological, micro-meteorological, glaciological and geological studies (Andrews and Barry, 1971). Major logistic support was given by National Air Guard units from Wyoming and Minnesota who flew men and equipment from Denver to Frobisher Bay.

The detailed studies of 1970 were replaced in 1971 by visits to the Boas Glacier in June and August. These visits enabled the 1970-1971 mass balance to be computed and it added further information on the temperature regime of the study area.

Of considerable assistance to our research has been the fact that the grant research reported here has been complemented in space and time, by other research activities of the Institute of Arctic and Alpine Research in the region of Cumberland Peninsula. Our knowledge of the Quaternary chronology has gained much from the research conducted under grants from the National Science Foundation to J.T. Andrews (GA-10992) that ran from 1969-1971 and to J.T. Andrews and J.C. Harrison (GA-28003) that started in 1971. Information on the regional energy budget is being investigated by R.G. Barry and J.D. Jacobs under grants from the Polar Programs Office, National Science Foundation, as grants GV-28220 and GV-28218. These grants have included analysis of satellite data, instrumented aircraft flights and ground stations.

Research Objectives

The purpose of the research discussed in this report was to attempt an integrated analysis of the past and present climates of the northern Cumberland Peninsula region with specific attention focused on the links between glacier distribution and fluctuations and the climate. The final objective of the research is to attempt to model the paleoclimate of the region during the late Quaternary.

These broad objectives can only be realised when a number of separate discipline topics have been researched. In general it seems to us quite clear that a complete understanding of climatic change and Quaternary history will only come about by a vigorous dialogue between climatologists, meteorologists, geologists, geographers, and others concerned with past environmental interpretation Too often is research on climatic change confined to a specific aspect, such as the dating of moraine sequences, and the interrelationships to glaciological/ climatological conditions are either ignored or inferred on a somewhat tenuous basis.

The original impetus for the project came after reconnaissance air photo mapping of glacial deposits and an analysis of the present distribution of glacier ice on the Okoa Bay 1:250,000 Map Sheet (Andrews, Barry and Drapier, 1970). These analyses showed that the present glaciers were nearly exclusively oriented toward the NW-N-NE whereas large end moraines delimited former large corrie glaciers flowing from south-facing corries. In a number of instances the moraine distribution suggested that the former south-facing glaciers had been considerably more extensive than their neighbouring north-facing counterparts. This general problem was, fortunately for us, typified in one rather small area between the heads of Narpaing and Quajon fiords where two glaciers presently face toward the north but extensive moraine systems delimit once larger glaciers that flowed southward from two basins cut into the southern flank of the mountain (Figure 1-1). Both of these south-facing features are currently ice-free. Low, gentle end moraines lie very close to the margins of the north-facing Boas and Akudnirmuit[1] glaciers (Figure 1-2).On the basis of air photograph interpretation these moraines were considered correlative to the major end moraine systems that marked the margins of the two glaciers lying in Sulung and Itidlirn valleys.

This small and relatively accessible area thus became the focal point for our research but to gain a wider, regional perspective, other facets of the project were conducted over a much larger area of Cumberland Peninsula.

Basically then our approach was to examine the various aspects of the problem on a number of different spatial and time scales. Below we identify the broad areas that we have investigated under the terms of the grants:

1. Large and medium scale mapping and analysis of the pattern of present-day glacier distribution both for Okoa Bay, but also for Cumberland Peninsula in particular and, in general, for the entire area of Baffin Island.

[1]Due to a typographic error this is shown as Akudlermuit in some of our publications arising from this research program.

2. Mapping of past glacier distribution and computations of past snowlines on both the medium and large spatial scales.
3. Mass balance studies of one representative glacier over the course of two years.
4. Macro-scale climatological analysis of conditions during the 1969-1971 balance years and for the 10 years or so prior to those dates.
5. Topo-climatic conditons in the Narpaing-Quajon area, and correlation analysis between that region and the DEWline weather station on Broughton Island.
6. Micro-meteorological conditions governing ablation processes during the 1970 summer on the Boas Glacier and analysis of conditions at sea level in 1971.
7. Climatological evidence for climatic change in the Cumberland Peninsula over the course of the period for which records are available.
8. Late Quaternary history of the Boas/Sulung area.
9. Late Quaternary history of the northern fiords of Cumberland Peninsula.

Table 1-1 represents a 2 x 2 table of space and time aspects of our study and itemizes those categories for which information is available, or has been gathered, and those where information is required. The table indicates the way our research design is structured. We are especially interested in the topo-climates and macro-scale climate over periods of 10 to 1,000 years.

The problem of linking the products of a glacial response, i.e. the moraine sequences, to the associated changes in mass balance and macro-scale, topo-climatic scale and micro-scale climatic changes has been stressed and discussed by Meier (1965). The links in the process-response model of climatic and mass balance changes and glacier response are not well understood. However, to some extent we feel that current studies on these links may overstress the lack of concordance in the association of glacier response and mass and climatic changes. The time scale of detailed observations on glaciers is only 1 to 2 decades and we suggest that this record is very noisy as observations have been undertaken during a period of climatic change.

The geological evidence for the timing of glacier advances during the last 5,000 years (Porter and Denton, 1967; Mercer, 1967) is that although the magnitude of a particular glacier response might vary from area to area that by and large the major advances are correlative. This apparent synchroneity of glacier response (Denton and Karlén, 1972) is of course in part an artifact of the uncertainty in the geological or radiometric dating of these deposits. This has the effect of imposing a filter on the glacier response record of ± 50 years over the last 300 years and ± 200 years on deposits older than this. From a geological viewpoint there are very few possibilities of improving, i.e. shortening the length of this filter because although, stratigraphically, a more detailed history might be preserved, available dating methods do not have sufficient resolution for determining the age of small scale events.

This smoothing of the glacier response record suggests that we will not be able to decipher, nor is there need to, the micro-scale and micro-time interactions of Table 1-1. Accordingly, we are primarily interested in those categories marked (?) on Table 1-1.

A number of papers have already been published on one or other of the 9 listed areas of interest, and others are in various stages of preparation. Hence it is not the purpose of this report to exhaustively cover the same material that is presented in these papers. Rather we have tried to take these various papers and present the main observations and conclusions. A list of papers published, in press, or in near final manuscript stage, is given below and this information should be available to the reader, prior to the main body of the report.

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Meier, M.F. 1965: Glaciers and climate. In: Wright, H.E., Jr., and Frey, D.G. (eds.) Quaternary of the United States, Princeton Univ. Press, 795-805.

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TABLE 1-1

Information available or required in terms of a space/time model on the response of glaciers to climatic/meteorological changes

TIME	SPACE: Micro-scale	SPACE: Topo-climate	SPACE: Macro-scale
1-2 Years	0	0	0
10^1 x 10^2 Years	X	(0)(?)	0
x 10^3 Years	X	?	?

0 Obtained

? Required

X Not Required

LIST OF PAPERS WHOLLY OR IN PART SUPPORTED UNDER THE TERMS OF THE GRANT

(Numbers in brackets at end refers to the appropriate section of the list above)

Andrews, J.T., Fahey, B.D., and Alford, D. 1971: Notes on correlation coefficients derived from cumulative distributions with reference to glaciological studies. J. Glaciol., 10(58), 145-147. (7)

Andrews, J.T., 1970: University of Colorado: 1969 summer field season in east Baffin Island. Arctic, 23(1), 62-63.

Andrews, J.T., and Barry, R.G. 1971: University of Colorado: 1970 summer field season in east Baffin Island. Arctic, 24(2), 145-146.

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Andrews, J.T., and Miller, G.H. 1972: Quaternary history of northern Cumberland Peninsula, Baffin Island, N.W.T., Canada: Part IV: maps of the present glaciation limits and lowest equilibrium line altitude for northern and southern Baffin Island. Arct. Alp. Res., 4(1), 45-59. (1,2)

Andrews, J.T., and Dugdale, R.E. 1971: Quaternary history of northern Cumberland Peninsula, Baffin Island, N.W.T: Part V: factors affecting corrie glacierization in Okoa Bay. Quat. Res., 1(4), 532-551. (2)

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Peninsula, Baffin Island, N.W.T. Part I: Late and Neoglacial deposits of the Akudnirmuit and Boas glaciers. Can. J. Earth Sci., 9(4), 403-414. (8)

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Miller, G.H., and Andrews, J.T. 1972: The Quaternary history of northern Cumberland Peninsula, Part VI: preliminary lichen growth curve for Rhizocarpon geographicum. Bull. Geol. Soc. Am., 83, 1133-1138. (8,9)

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Williams, L.D. 1972: Some factors influencing cirque glacierization on eastern Cumberland Peninsula, Baffin Island, Canada. Unpublished Master's Thesis, Univ. Colorado. (1,2)

Andrews, J.T., and Miller, G.H. 1972: The Quaternary history of northern Cumberland Peninsula, east Baffin Island, N.W.T. Part X: radiocarbon date list. Arct. Alp. Res., 4(3), 261-277.

CHAPTER ONE - FIGURE CAPTIONS

Figure 1-1: Upper: Topographic map of the primary field area between the heads of Narpaing and Quajon fiords. The rectangle refers to the area in which detailed computations have been made on the topographic effects on global radiation receipts.

Lower Left: Location of the area of northern Cumberland Peninsula relative to Baffin Island, N.W.T.

Lower Right: Location of Boas Glacier at environs of Broughton Island.

Figure 1-2: Schematic map of the ice bodies and moraines within the primary field area. Compare for location, elevation, etc. with Figure 1-1 (Upper).

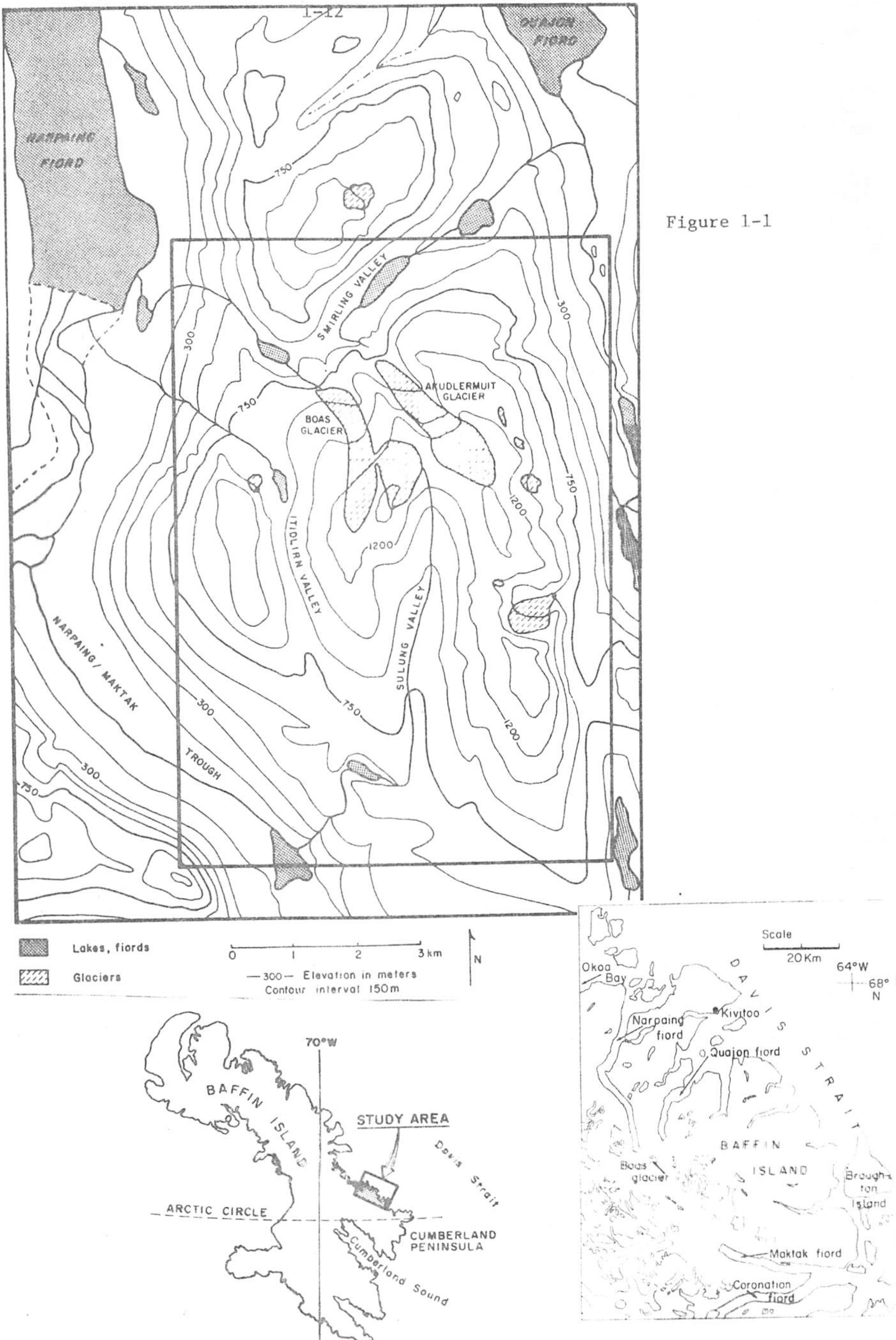
QUAJON FIORD
NARPAING FIORD
SMIRLING VALLEY
AKUDLERMUIT GLACIER
BOAS GLACIER
ITIDLIRN VALLEY
SULUNG VALLEY
NARPAING / MAKTAK TROUGH
300
750
1200
Lakes, fiords
Glaciers
0 1 2 3 km
—300— Elevation in meters
Contour interval 150m
N
70°W
BAFFIN ISLAND
STUDY AREA
Davis Strait
ARCTIC CIRCLE
CUMBERLAND PENINSULA
Cumberland Sound
Scale
20 Km
64°W
68° N
Okoa Bay
DAVIS STRAIT
Kivitoo
Narpaing fiord
Quajon fiord
BAFFIN ISLAND
Boas glacier
Broughton Island
Maktak fiord
Coronation fiord

Figure 1-1

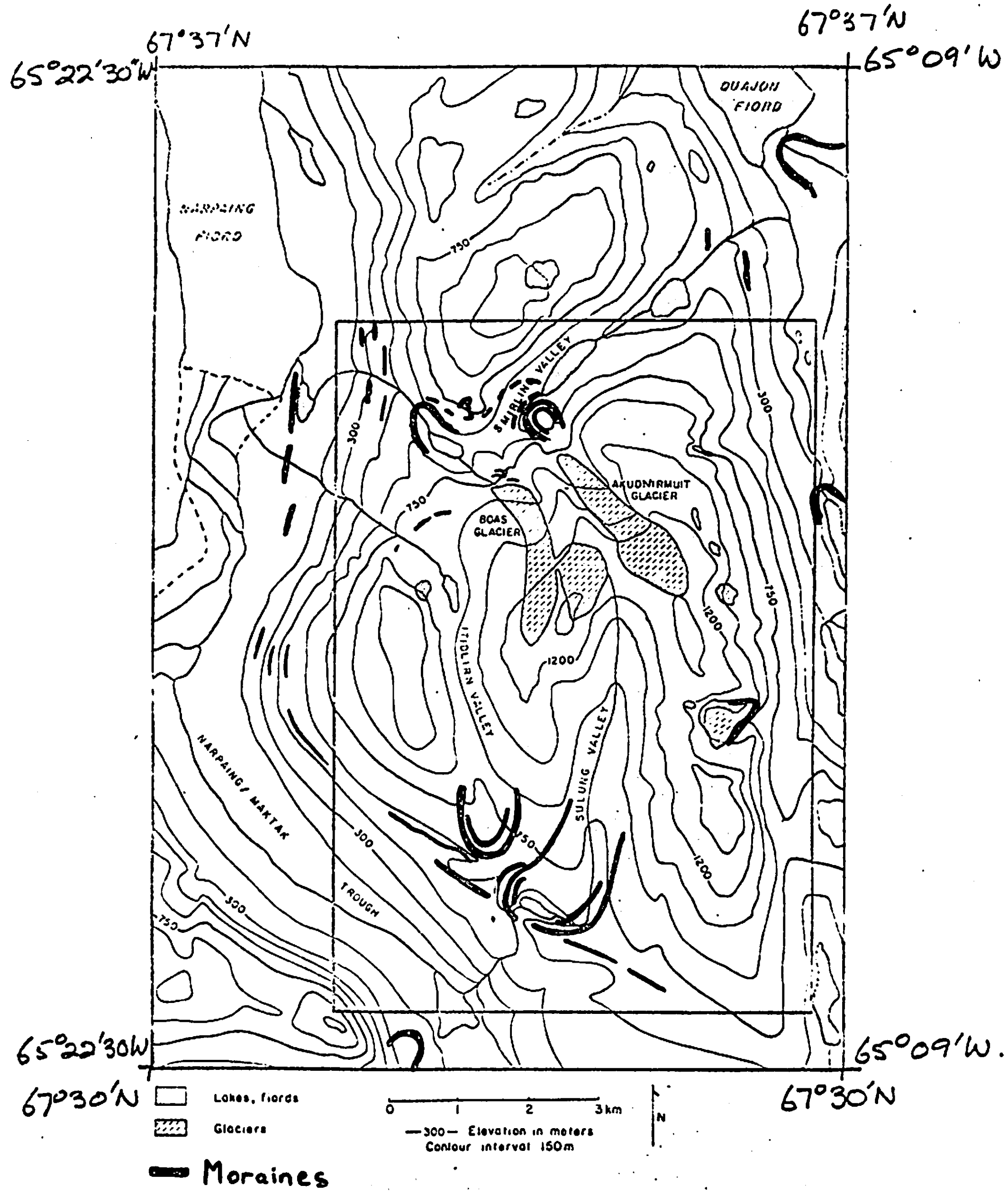
67°37'N
65°22'30"W
67°37'N
65°09'W.
QUAJON FIORD
NARPAING FIORD
750
300
SMIRLING VALLEY
AKUDNIRMUIT GLACIER
BOAS GLACIER
1200
ITIDLIRN VALLEY
SULUNG VALLEY
NARPAING/MAKTAK TROUGH
65°22'30W
65°09'W.
67°30'N
67°30'N
Lakes, fiords
Glaciers
0 1 2 3km
N
—300— Elevation in meters
Contour interval 150m
Moraines

Figure 1-2

2. GENERAL BACKGROUND TO FIELD AREA

J.T. Andrews and R.G. Barry

The purpose of this chapter is to briefly present background data on northern Cumberland Peninsula. This information includes aspects of location, geology, topography, ice distribution, climate and previous work. These data will serve to set the stage for the more detailed discussions of our research which will commence in the next chapter.

LOCATION, GEOLOGY, TOPOGRAPHY AND ICE DISTRIBUTION

Cumberland Peninsula is a rectangular block of land, jutting at an angle from the general SE-NW trend of the east coast of Baffin Island. The Peninsula itself is aligned with its long axis oriented NW-SE and it is 260 km long by 150 km wide. In the south it is bounded by the large Cumberland Sound where maximum water depths reach -600 m. On the east side of the block the coast fronts Davis Strait, and on a clear day Greenland can be viewed from the high mountains along this extreme eastern coast of Baffin Island. Davis Strait has a depth of about -700 m. In the west, the mountains of Cumberland Peninsula grade gently into the broad interior plateau of Baffin Island with summit elevations of about 600 m which in turn decline in elevation toward Foxe Basin (Figure 2-1). The northern coast of Cumberland Peninsula extends NW from Durban Island to the vicinity of Nedleuksuak Fiord at which point the coast turns slightly and continues its normal trend to Home Bay (Figure 2-1).

The Peninsula has recently been mapped at a reconnaissance level by the Geological Survey of Canada (summer 1970) but as yet no map is available. Riley (1957) studied the geology around parts of Cumberland Sound and Clarke and Upton (1971) have discussed the distribution and age of a volcanic suite that lies along the outermost coast between Cape Dyer Island and Cape Searle. These rocks were probably extruded as the result of the rifting apart of Greenland and eastern Baffin Island. The present area of Baffin Bay and Davis Strait was probably produced as a result of this rifting. The basalts on either side of Davis Strait date from late Cretaceous to early Tertiary in age. The development of Baffin Bay and Davis Strait was accompanied by isostatic uplift of the margins of the rift. Estimates of the amount of uplift vary between 400 and 600 m. During the rise of the margin extensive faulting took place both parallel to, and normal, to the margin. Cumberland Peninsula appears to be major horst bounded on the south by the Cumberland Sound graben. The majority of fiords have considerable stretches that parallel the major fault trends. The latest large scale episode might have been a submergence of the coast (Le Pichon, et al., 1971). Although the fiords have certainly been cut to a degree by large outlet glaciers the presence of major troughs, extending across the Continental Shelf in areas where the vertical extent of the ice can be shown to have been limited to less than 600 m a.s.l. (Clarke and Upton, 1971), indicate that they cannot be explained as ice erosional forms.

The climate during the Paleogene was warm temperate as shown by the flora of the 'Rimrock Bed' in central Baffin Island (Andrews, Guennel, Wray and Ives, 1972). No evidence has yet been found that indicates when glaciation of Cumberland Peninsula began. However, the rifting and isostatic uplift of the rim probably occurred during the early mid Tertiary and with the evidence for climatic deterioration and even glaciation in Alaska during the Miocene (Bandy, et al., 1969; Denton and Armstrong, 1969) it is considered probable that Miocene glaciation could have occurred. The topography at that time was no doubt different from that seen today and a pre-Quaternary glaciation might account for anomalous erratic blocks located within areas that show no other evidence of glaciation (see Chapter 9). Immediately below this well-weathered zone (Weathering Zone I) occurs a weathered till sheet that Pheasant (1971) calls Weathering Zone II. The boundary between Zone II and the youngest zone, Zone II, is frequently marked by long stretches of lateral moraine. The age of Zone III apparently spans the entire Wisconsin Glaciation (Pheasant, 1971; Pheasant and Andrews, 1972) whereas Zone II is clearly much older and may mark a Glaciation 500,000 years ago.

Cumberland Peninsula is one of the more heavily glacierized regions along the east Baffin Island coast. A discussion of the present ice distribution in the northern part of the Peninsula is given in Andrews, Barry and Drapier (1970) and map s of glaciers and other ice bodies have recently been made available by the 'Glacier Inventory Project' of the Canadian Glaciology Subdivision. The area is physically dominated by the 6,000 km^2 mass of the Penny Ice Cap. In the eastern part this ice cap flows through spectacular alpine topography but to the west it ends on the gently sloping interior plateau (Figure 2-1). Continuation of topographic elevations beneath the ice sheet suggest that it has a thickness of 600-700 m in the center where it rises to nearly 2000 m a.s.l. Smaller individual ice caps, transection glaciers, valley glaciers and corrie glaciers lie to the north and east of the Penny Ice Cap. They can be divided into major groups on the basis of large topographic blocks separated by significant, fault-controlled fiords and valleys. An example of such a group is the area lying between North Pangnirtung and Pagnirtung fiords to the west and Kangert and Padle fiords to the east. Our primary interest is in the features bounded on the west by through trough that limks the heads of Narpaing and Maktak fiords and on the south and north by Maktak and Narpaing fiords proper. The Penny Ice Cap and/or the Laurentide Ice Sheet were channelled around the periphery of this area and glaciation of the block was primarily the result of the expansion of local ice fields and the growth of local valley glaciers.

Our primary field research area lies on the northwestern limit of the area - the Sulung Valley directly overlooks the Narpaing/Maktak through trough (Figure 1-1). To the east of the Sulung/Boas area, a high col at 600 m a.s.l. or so connects the through trough with Quajon Fiord. The divide between the Sulung/Itidlirn valleys and the Boas and Akudnirmuit glaciers lies at about 1250 m a.s.l. The south-facing interfluves drop gradually toward the through trough but become oversteepened at about 750 m a.s.l. and then drop steeply into the trough. The former Sulung and Itidlirn glaciers extended 4 to 6 km from the corrie headwalls and stopped on a broad bench at 600 m a.s.l. Immediately below the bench the valley sides are steep (Figure 1-1). On the northern side of the divide the topography is much more rugged and both present glaciers descend steeply into the east-west trending Smirling Valley. Smirling Valley links the heads of Narpaing and Quajon fiords - the lip overlooking Quajon is 356 m and that above Narpaing is 604 m a.s.l. Access to the field area involves ascending either the side wall in Quajon or Narpaing fiords. Although the route from Narpaing is steeper it has the advantage that the route into the Boas Glacier or Base Camp is relatively smooth whereas the traverse from the lip overlooking Quajon along the large rocky talus toward Base Camp can be treacherous. The maine lake in Smirling Valley (Figure 1-1) is large enough to land a single-engine De Haviland Otter with full pay load on skiis.

The Boas and Akudnirmuit glaciers descend from the local divide where part of their upper accumulation zones are in the form of thin ice patches. The Akudnirmuit Glacier drops over a steep backwall at 1120 m a.s.l. into the cirque basin before flowing a steep gradient into Smirling Valley. At the ice fall the glacier is crevassed. The Boas Glacier is considerably gentler in gradient and hence less hazardous to work on. The glacier has two accumulation areas that are referred to as the East Bowl and West Bowl although the West Bowl is in reality a direct continuation of the main glacier (Figure 1-1). On the Glacier Inventory of Canada the Akudnirmuit and Boas glaciers have the unique identification 46204J67 and 46204J68 respectively. The Boas Glacier has an elevation range from 800 to 1300 m a.s.l. whereas the Akudnirmuit Glacier extends from about 530 m to 1300 m a.s.l. From the top of the local divide the Penny Ice Cap forms the horizon to the west and south at a distance of 15 km. The main divide of the Penny Ice Cap is 45 km away at 2000 m a.s.l.

WEATHER STATIONS AND RECORDS

A partial list of past and present weather stations for Baffin Island is included in this report as Table 11-1. In terms of the local Cumberland Peninsula region weather stations currently operate on Broughton Island and Cape Dyer, but in the past records have been kept at Kivitoo, Durban Island, Padloping Island, Pangnirtung and the head of Cumberland Sound. All these sites are, of course, coastal. The DEW Line stations are located on hilltops and they thus have a considerably different location from stations attached to settlements, such as Pangnirtung - which are all near sea level.

The difference in station elevation is important in view of the frequent presence of low-level temperature inversions in the Arctic. Bradley (in press) draws attention to this factor with reference to the determination of the glaciation limit from the estimated altitude of the mean July 0°C isotherm (Andrews and Miller, 1972). At Clyde, the mean inversion depth is 340 m in July according to Bilello (1966).

Data collected by INSTAAR in Broughton "village" at a site near sea level since 5 June 1971 provide a comparison with the station at the DEW Line site (580 m a.s.l.). The Broughton village station (10 m a.s.l.) has maintained a continuous record of temperature, relative humidity and pressure, as well as twice-daily observations of wind, state of sky, snowdepth and sea ice conditions[1].

Preliminary comparisons for June and July 1971 show the following:

	Broughton Village (10m a.s.l.)			Broughton Island (580 m a.s.l.)		
	Mean Daily Temp. (°C)	Precip. (cm.w.e.)	Precip. days	Mean Daily Temp.(°C)	Precip. (cm.w.e.)	Precip. days
June	(0.6)	(0.54)	(3)	1.1	0.47	2
July	3.8	1.17	9	6.0	1.01	8

There is little difference in the precipitation totals but temperatures are lower at the village station, especially in July. This reflects the presence of sea ice in the harbour when the snow cover had gone from the upper site. In addition, there is a frequent low stratus deck over the village. The DEW Line site reported "fog below the station" on 13 days in June and 20 days in July. These differences are of great importance for the radiation regimes and must clearly be taken into account in studies of the relationships between sea ice and climatic data for stations in this area.

A further comparison was made betwen temperatures in the area of the Boas Glacier and on Broughton Island. For 1-18 August, 1970, a correlation of +0.9 was obtained between daily mean temperature at Base Camp (497 m a.s.l.) and the Broughton Island DEW Line station. However, there was a poor correlation between Base Camp and the Broughton village station in summer 1971. If it is substantiated the upland climate (at 'open' sites) in this region shows rather little areal variation, it will be of considerable importance for future

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[1]Copies of these records are deposited with the Climatological Division, Atmospheric Environment Section, Department of Environment, Downsview, Ontario as well as at INSTAAR.

assessments of conditions, past and present, based on the coastal records.

PREVIOUS RESEARCH IN THE AREA

Despite the obvious scenic and scientific attractions on Cumberland Peninsula and its relative accessibility by commercial airline (from Frobisher Bay to Pangnirtung or Broughton Island), there has been relatively little research on the Quaternary geology, present and past climates, and glaciology. The great proportion of existing literature arose as the result of a single, successful expedition to the Penny Ice Cap in 1953. The expedition was sponsored by the Arctic Institute of North America (AINA) and included a variety of studies on glaciology, geomorphology, and biology. A comprehensive study of southern Baffin Island has been presented in a RAND report (McGill University, 1963) and members of the Canadian Government have made brief halts at various sites. In 1970 the Geological Survey of Canada mapped the area during its "Operation Penny" but as of writing this report a preliminary geological map has not yet been published. Geological papers on the area of Cumberland Sound included the Ph.D. thesis of Riley (1957) and the volcanic series around Cape Dyer has recently been discussed by Clarke and Upton (1971) The only Quaternary geology reported for the area is included in Thompson (1954); Ives (1963); Riley (1957); Clarke and Upton (1971) and Ives and Bornes (1971).

Since 1967-1968 studies on the Quaternary geology, past and present climates, glaciology and air/ice/sea interactions have been pursued vigorously by graduates and faculty at INSTAAR.

The studies on the Penny Ice Cap in 1953 indicated (Ward and Baird, 1954) that the density for the snow pack in early spring was 0.33 g cm^{-3} and that accumulation amounted to about 1 m of snow. Total accumulation (winter and summer) averaged 0.41 m H_2O. At an elevation of 400 m a.s.l. the ablation that year amounted to

2.68 m H_2O compared to 0.1 to 0.15 m H_2O at 2050 m.

Observations on the marine limits by Robitaille (in Ives, 1963) and Thompson (1954) indicated that they were low. The latter author reported elevations of 15 and 10 m respectively for the upper marine features in North Pangnirtung and Pangnirtung fiords and Ives (1963) indicates observations suggest that the marine limit is 0 m (i.e. submergence) at Cape Dyer and 18 m or so at Broughton Island and Padloping Island. In retrospect these figures could have been used to imply a limited ice load over the area during the late Quaternary (see Chapter 9).

Individual papers will be referred to in specific chapters but the bibliography below includes the major prior references:

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THE FOLLOWING CORRECTIONS SHOULD BE INSERTED

Page	Line	Correction
v	16	Delete work, insert word.
2-1	20	Delete Nedleuksuak, insert Cape Dyer
2-1	26	Delete Island
2-3	2	Delete Zone II, insert Zone III
2-4	19	Delete maine, insert main.
3-9		Insert minus signs for Base Camp, July, min = -0.3 Boas Glacier, July, min = -2.3
6-1	8	Delete m^2, insert m^3
6-4	9	Delete Thus, insert This.
6-7	5	Delete They, insert The net balance.
6-7	21	Delete reasonably, insert reasonable
6-8	16	Insert (800 m a.s.l.) and 1050 m a.s.l., but
7-2	12	Insert $\bar{U}$ = 2.04 m yr^{-1}.
7-7	22	Delete 1C/20 m insert 1°C/20 m.
9-2	17	Delete in pre-Quaternary, insert during the stadials.
9-3	21	Delete statials, insert stadials.
9-6	12	Delete 134,000 to 768,000 BP, insert 134,000 to >68,000 BP.
9-12	10	Delete Figure 9-3 and 9-5, insert Figure 9-3 and 9-4.
9-15	15	Insert (Abstracts).
9-20	Title	Delete of their surfaces, insert on their surfaces
10-2	19	Insert 700 m a.s.l. and 1,000 m a.s.l.
10-9	10	Delete The area under the ELA was ca. 8,900 m, insert The cross-sectional area below the ELA was ca. 8,900 m.
10-13		Insert, Flint, R.F. 1971: Quaternary and Glacial Geology. John Wiley and Sons, New York. Terasmae, J., Webber, P.J., and Andrews, J.T. 1966: A study of late-Quaternary plant bearing beds in north-central Baffin Island. Arctic, 19, 296-318.
11-1	9	Delete northern Hemisphere, insert Northern Hemisphere
11-1	26	Delete shows an elevation, insert show an elevation
11-2	1	Delete 1.2°C, insert 1.4°C.
11-2	18	Delete Figures 11-1 and Table 11-1, insert Figure 11-1 and Table 11-1
11-8	7	Delete Allectoria, insert Alectoria.
11-11	20	Insert, Falconer, G. 1966: Preservation of vegetation and patterned ground under a thin ice body in northern Baffin Island, N.W.T. Geogr. Bull., 8, 194-200.
11-12	16	Delete Bull. Geol. Soc. Amer. (in press), insert Nature, 237, 386-389.
12-11	17	Delete lives, insert lived.
12-12	12	Insert of 200 - 450 m
12-13	3	Delete relation, insert related.
12-14	20	Delete (submitted), insert 1970
12-14	21	Add Arct. Alp. Res., 4(4), 359-360.

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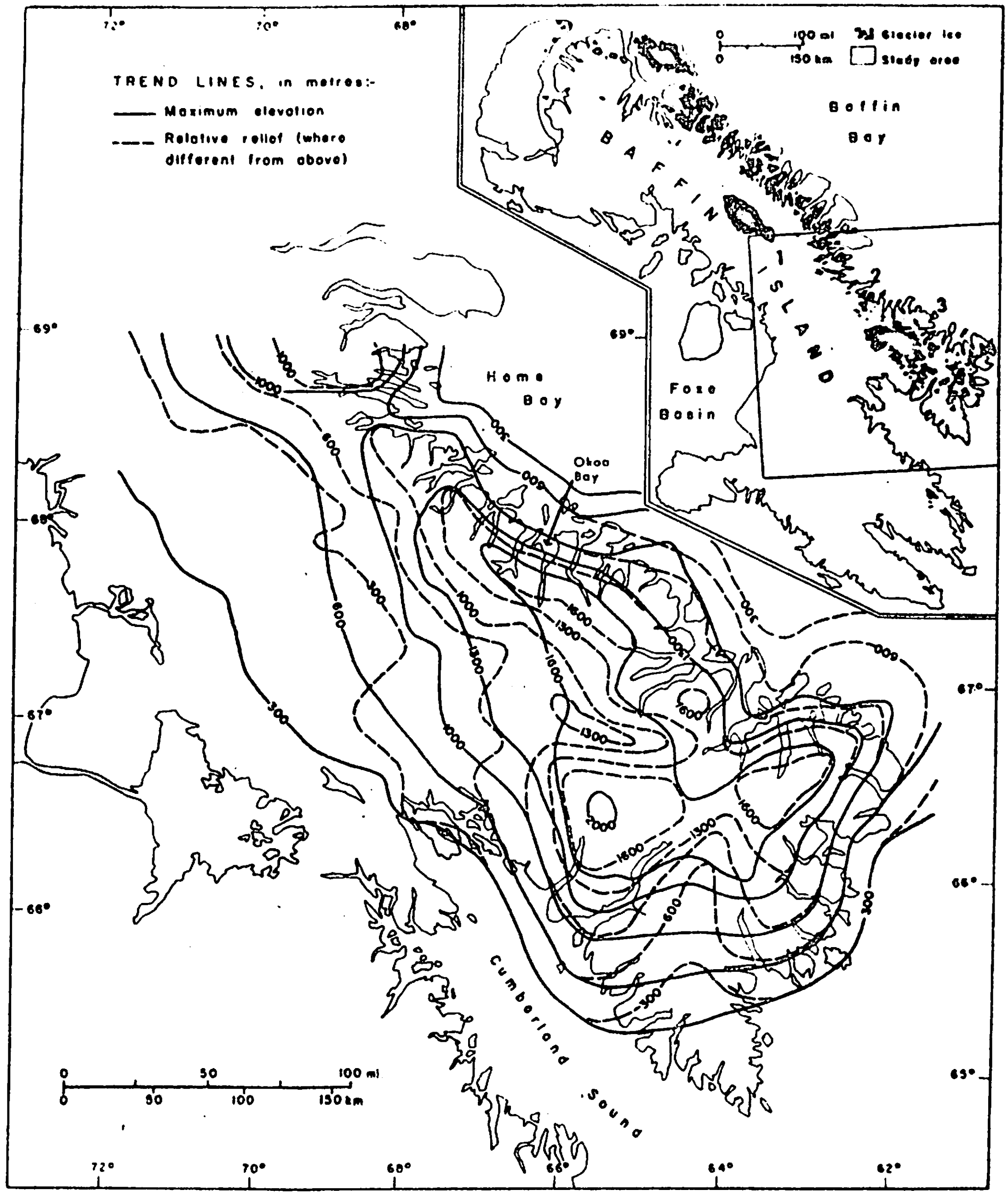

2-1: *Contours of maximum summit elevations and maximum relative relief for the area between Home Bay in the north and Cumberland Peninsula in the south. Inset map shows the distribution of ice masses in Baffin Island and location of weather stations 1 Dewar Lakes, 2 Cape Hooper, 3 Broughton Island, 4 Cape Dyer, 5 Frobisher.*

3. METEOROLOGY AND CLIMATOLOGY

R.G. Barry

OBJECTIVES

The objectives of the meteorological program were as follows:

1. to determine the "regional climate" of the area and its local variability.
2. to establish the energy budget of the Boas Glacier for summer 1970.
3. to correlate the meteorological data with the circulation regimes identified on synoptic weather maps.
4. on the basis of (2) and (3) to assess what regimes might account for the past distributions of ice in the area.

METHODS

Procedures implemented to meet the first two objectives were the establishment of a micrometeorological observing program on the glacier and the installation of a network of other recording stations visited at weekly to monthly intervals (Figure 3-1). The micrometeorological program relating to objective (2) is discussed in Chapter 4 by Jacobs. The data available from the climatological station network summarised in Table 3-1.

The clear-sky global radiation receipts over the area on 21 August are determined in Chapter 10 (Figure 10-7). This map delimits the theoretical extremes of summer radiation climate. Although this map was not available when the climatological stations were located it is, nevertheless, clear that most of the probably range of local climatic variation due to slope, aspect and topographic shading is represented in the data collected.

The methods used in connection with objectives (3) and (4) involved classifying the daily MSL synoptic weather maps for 00 GMT (1900 EST the previous day) according to a scheme developed for the area 55° - 75°N, 50° - 100°W (Barry, 1972). The classification is shown schematically in Figure 3-2.

WEATHER CONDITIONS, SUMMER 1970

Temperature

The general character of the summer is illustrated by Table 3-2. Temperatures were several degrees below normal and July was particularly cold in eastern Baffin Island. This may be compared with summers 1969 and 1971 which were close to average even for individual months. Temperatures at Base Camp (Figure 3-3) remained low until the end of June. From 10-30 July maxima fluctuated around 7°C on average, and minima were close to 0°C. The latter appear to reflect the effect of the lake. The rather low maxima are caused by the short "day-length" due to topographic screening as well as by the occurrence of stratus in the valleys. This was observed below the Glacier Station at 16% of observations in July (13% for the whole observation period).

Comparisons between stations are hampered by the fact that bias occurred in the maxima particularly as a result of heating effects in the screens. This is a common problem over snow surfaces with a high albedo but it was aggravated by the design of the lightweight aluminum screens which, in their original version as used in 1970, were unpainted and had a single floor (see Chapter 5). The mean differences between the thermometer in the screen and psychrometer readings at the Boas Glacier station (ΔT) were:

Time (LST)	00	04	08	12	16	20
ΔT (°C)	-0.5	(0.0)*	1.1	2.2	1.3	0.4
Number of observations	33	--	44	47	49	44

*Interpolated graphically

For midday the difference was 1.8°C for 8-10/10 cloud cover and 3.0°C for 0-4/10 cloud cover.

Table 3-3 compares the mean daily maxima and minima at each temporary station and also at Frobisher, Cape Dyer, and Broughton Island. The maxima have (where indicated) been adjusted by -2.2°C for the approximate periods of snow cover around the instrument screens. The Boas Glacier station was similar to Broughton Island in July in spite of its 560 m higher situation.

Comparisons between the stations are most easily considered in terms of potential temperature; that is, the temperature of the air brought to 1000 mb pressure by a dry adiabatic (isentropic) process. Pressure at mean sea level is 1013 mb on average although in individual months and locations it may only be within ±10 to 15 mb of this. Nevertheless, the differences are immaterial for purposes of relative comparisons within the same area. The 13 mb difference between m.s.l. mean pressure and 1000 mb is equivalent to approximately 110 m elevation.

Table 3-4(a) shows potential temperatures for July mean daily average and mean daily maximum calculated on this basis. The means are between 10°-11°C except at Base Camp and Broughton Island which are some 4°C colder. With respect to maxima, East Corrie is similar to Base Camp and these two are 3°C colder than the stations on the Boas Glacier. Again Broughton Island is the coldest station.

Another means of comparison is the "lapse rate" between the stations (Table 3-4(b)). In June there is no difference between the mean daily maximum at Base Camp and the Boas Glacier, but the difference increases through the summer, probably for the most part in response to increased shadowing effects at Base Camp. The mean daily minimum shows a similar but less pronounced trend.

The "lapse rate" for the mean daily temperature between Mid-Glacier station corresponds to the mean environmental lapse rate, but that for the Boas Glacier-Divide is 2-3 times too large which may point to calibration problems with the thermohygrograph at Divide Station. However, the mean potential temperature at Divide is in line with the other values.

Precipitation

Precipitation totals for the summer 1970 were close to normal in the area, although June received above the average amount - 128% of normal at Broughton Island (see Table 3-4). Within the limits of these measurements, there appears to be little variation over the research area (Table 3-5). However, there is evidence that totals over the year are highest in the north-facing valleys. This is discussed below.

In July, the total precipitation in the research area was comparable with that at Cape Dyer rather than Broughton Island. Moreover, for the period 9 June - 15 August there were 34 days with measurable precipitation at the Boas Glacier compared with only 11 at Broughton Island. Half of the total at the latter station during this period fell on only 2 days.

Records have been obtained from Hoinkes storage gauges in the Smirling Valley, north of the Boas Glacier and in Sulung Valley to the south since winter 1968-1969, in the former case. The totals are listed on Table 3-6.

Assuming that the catches are not biased by any special site effects, the differences may by the result of a disparity in the intensity and/or frequency of storms from northerly and southerly direction.

Analysis of the 1.5 m level wind records at Boas Glacier (based on 10 minute runs at 4 hourly intervals, omitting 0400 EST) by J.D. Jacobs, shows the following results for 287 observations between 9 June to 4 August 1970:

Direction	N	NE	E	SE	S	SW	W	NW	Calm ($<0.3\ m\ s^{-1}$)
Percent-frequency	20.9	13.2	9.4	24.7	12.5	10.4	0.4	7.6	1.0

On the basis of this distribution, precipitation days have been grouped into northerly and southerly flows according to the predominant wind direction for the day:

	Precipitation Days	Average Precipitation (cm H_2O)
Northerly flow	29	0.15
Southerly flow	10	0.13

There is only a small difference between the two groups in terms of precipitation intensity, but storms with northerly airflow were three times as frequent as those with southerly airflow. Comparison of northerly and southerly flow for days with and without precipitation by χ^2 indicates that the difference between the two wind direction groups is significant at 2.5% level. These results suggest that north-facing "slopes" in the area will have higher precipitation than south-facing ones, at least in years with a similar pattern of cyclonic activity to summer 1970. As shown on Table 3-7, July-August 1970 had a circulation regime very similar to the mean conditions for July-August 1961-70.

SYNOPTIC CHARACTERISTICS OF SUMMER 1969, 1970 AND 1971

Table 3-7 shows the striking contrasts between the summer of 1969, 1970, and 1971 in terms of cyclonic activity in the area between Foxe Basin and Davis Strait, and Hudson Strait to latitude 70°N, and also with respect to 700 mb height departure from normal. Cyclonic control of the MSL circulation was pronounced in summer 1970.

The circulation type frequencies are given in Table 3-8. The frequencies of the two main groups clearly differentiate 1969 from the two succeeding years. In 1969 anticyclonic circulations, and particularly the 900 group of high cells centered over Baffin Island, accounted for 68% of days in July-August. Although 1970 had as many anticyclonically controlled circulations as 1971 it should be noted that 14.5% of the total occurred in the last 9 days of August 1970, too late to be of real significance for ablation on the Boas Glacier. Thus, the frequencies for 1970 are, from this standpoint, not quite as close to "normal" (1961-70) as they might appear.

Precipitation Synoptic Climatology

The large variability in precipitation amounts in time and space makes it unwise to consider synoptic events of a single season in too great detail. However, the synoptic catalogue for Baffin Island (Barry, 1972) provides a basis for longer term evaluations. Table 3-9 summarizes a synoptic climatological analysis of precipitation at Broughton Island for July-August 1961-70.

The dominance of lows centered over Baffin Island in the contribution they make to the total precipitation is readily apparent. However, in summer 1970 these situations gave mostly rain, rather than snow, on the Boas Glacier. Snow occurred with a variety of situations including those with a low in the Davis Strait and those with a major low to the south or southwest accompanied by a trough or old primary center over Baffin Island.

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TABLE 3-1

DATA AVAILABLE FROM CLIMATOLOGICAL STATIONS

The location of climatological stations in the field area

	Temp.	Rel. Hum.	Precip.	Wind Run	Radiation
Boas Glacier	x	x	x	(Profile)	Daily and instant. (short and net wave)
Mid-Glacier	7-day	7-day	x		
Base Camp	x	x	x	x	
Divide	7-day	7-day	x	x	Weekly actinograph
South Corrie	31-day	31-day	x	x	
East Corrie	31-day	31-day	x	x	
Sulung	7-day		x		

TABLE 3-2

CHARACTERISTICS OF SUMMERS 1970 AND 1971

		(°C) Mean Temperature Departures			Precipitation Departures %		
1970		June	July	August	June	July	August
Broughton Is.	581 m	-1.5	-3.4	-2.2	+28	-94	+7
Cape Dyer	368 m	-1.7	-3.3	-2.0	+42	+16	-15
1971							
Broughton Is.		+0.5	+0.6	-1.1	-64	+180	-64
Cape Dyer		+0.4	-0.4	-0.9	-68	+190	-40

Departures based on 8-9 year means.

Table 3.3: MEAN DAILY MAXIMA AND MINIMA AND DEPARTURES FROM "NORMAL", 1970. (°C)

9-day binomially weighted running mean temperatures at Base Camp

STATION	ALTITUDE m.	JUNE Max.	JUNE Min.	JUNE Departures		JULY Max.	JULY Min.	JULY Departures		AUGUST Max.	AUGUST Min.	AUGUST Departures	
FROBISHER	21	4.3	-1.1	-2.4	-1.6	9.5	2.3	-2.2	-1.8	9.8	3.1	-0.7	-0.4
CAPE DYER	368	1.8	-4.5	-1.6	-2.4	5.1	-0.3	-4.5	-3.2	6.1	-0.3	-1.8	-1.1
BROUGHTON IS.	581	0.2	-4.8	-1.7	-1.2	4.6	-1.5	-3.7	-3.5	4.4	-1.2	-2.0	-2.0
BASE CAMP	490	0.8	-4.9			7.0	-0.3			(5.9)	(-0.1)		
MID-GLACIER	986					5.4*	-1.8						
BOAS GLACIER	1142	(-1.5)* 8-30 June	(-7.1)			3.9*	~2.3			(0.4)* 1-16 Aug.	(-3.8)		
DIVIDE	~1300	(-3.6)* 12-15,17-30 June	(-7.1)			1.1* 24 July missing	-4.0			(-3.5)* 1-15 Aug.	-6.1		
E. CORRIE	~1150					0.6*	-2.0						
S. CORRIE	970					6.4	-1.7			(2.4) 1-16 Aug.	(-2.8)		
SULUNG	610					8.0 (17-31 July)	1.3			(5.0) 1-16 Aug.	(0.1)		

* Maximum adjusted -2.2 C during snow cover period.

TABLE 3-4

TEMPERATURE DIFFERENCES BETWEEN STATIONS

(a) Potential Temperature at 1000 mb

STATION	ALTITUDE (m)	July Mean (°C)	July Max. (°C)
Base Camp	490	6.8	10.8
Mid-Glacier	986	10.4	14.0
South Corrie	970	10.8	14.8
Boas Glacier	1142	10.9	14.0
East Corrie	1150	10.1	10.8
Divide	1300	10.2	12.8
Broughton Island	581	6.2	9.2

(b) "LAPSE RATES" (°C/100m)

Base Camp - Boas Glacier (497 - 1142m)

	Max.	Min.	Mean
8 - 30 June	0.00	0.34	0.17
1 - 31 July	0.48	0.40	0.44
1 - 16 August	0.85	0.57	0.71

Mid-Glacier - Boas Glacier (986 - 1142m)

	Max.	Min.	Mean
1 - 31 July	0.95	0.32	0.63
13 June - 16 August	0.94	0.21	0.58

Boas Glacier - Divide (1142 - 1300m)

	Max.	Min.	Mean
1 - 31 July	1.77	1.08	1.42
1 - 15 August	2.47	1.45	1.96

TABLE 3-5

PRECIPITATION - 1970 (cm)

% Snow in Parentheses

STATION	ALTITUDE (m)	JUNE	JULY	AUGUST	SUMMER
Frobisher	21	6.2 (85%)	7.9 (10%)	5.1 (0)	19.2
Broughton Island	581	4.2 (100%)	0.1	2.7 (46%)	7.0
Cape Dyer	368	7.0 (100%)	2.7 (39%)	6.1 (54%)	15.8
Base Camp 9 June - 16 August	497	2.4	1.9	(1.9)	6.2(71 days)
Boas Glacier 6 June - 16 August	1142	2.6 (100%)	1.9 (22%)	(2.1)(90%)	6.6(72 days)
Mid-Glacier 12 June - 16 August	986				5.3(66 days)
Divide 11 June - 15 August	1300				4.3(66 days)
South Corrie 14 June - 10 August	970				4.8(58 days)
Sulung 17 July - 14 August	610				3.1(29 days)

TABLE 3-6

TOTAL PRECIPITATION (cm H_2O)

	Smirling Valley Gauge 1	Smirling Valley Gauge 2	Sulung Valley period (Gauge 3)	Sulung Valley Gauge 3	Sulung Valley period (Gauge 4)	Sulung Valley Gauge 4
Elevation (approx.) m	546	613		664		590
Winter 1968-69 (August 1968 to 18 August 1969)		30				
Winter 1969-70 (18 August 1969 to 23 June 1970)	7.6	15	(23 August 1969 to 22 July 1970)	13.8	(23 August 1969 to 17 July 1970)	5.8
Winter 1970-71 (18 August 1970 to 9 June 1971)	34.3	33.8				
Summer 1971 (9 June to 4 August 1971)	3	11			(17 July 1970 to 4 August 1971)	20.7

Gauge 1. below Akudnirmuit Glacier
Gauge 2. below Boas Glacier

Gauge 3. West end of Sulung
Gauge 4. West end of lake below Sulung valley

TABLE 3-7

COMPARISON OF 1969, 1970 AND 1971

(a) Days with cyclones centered in the area Foxe Basin-Davis Strait

	May	June	July	August	September
1969	7	7	2	0	7
1970	7	11	10	9	9
1971	12	3	8	10	-

(b) 700 mb height anomalies (g.p.m.)

	May	June	July	August	September
1969	-40	-60	+20	+30	-40
1970	-80	-70	-40	0	-30
1971	-70	+60	0	-45	-30

Sources: Climatological Data. National Summary (Asheville, N.C.) Volumes 20-22, Nos. 5-9, 1969-1971.

Monthly Weather Review (Washington, D.C.), Volumes 98-100.

TABLE 3-8

COMPARISON OF CIRCULATION REGIMES, JULY-AUGUST 1969-1971

		Frequency (%)			
Types[1]	Description	1969	1970	1971	1961-70[1]
100,101,102,110,130	Central low/trough	3.2	16.1	19.4	15.6
500,510,542	Davis Strait low	6.5	11.3	9.7	9.4
200,210,520,532,540,702	Baffin Bay low	4.8	3.2	8.1	4.7
600,610	Low to SW	11.3	4.8	4.8	10.2
612,620,630,640	Low to SW or S with other lows	6.5	9.7	4.8	9.2
	Cyclonic control	32.3	45.1	46.8	49.1
900,910,920,930	Anticyclone	32.3	17.7	6.5	10.0
120,800,810	Ridge situations	3.2	6.5	9.7	8.4
300,310,320	Ridge, low to S	9.7	11.3	16.1	10.3
400,410,420,430	High in E, low to W	9.7	11.3	9.7	9.8
511,521	Ridge, Baffin Bay low (NE flow)	9.7	3.2	0	3.7
220,531,701	Ridge, Baffin Bay low (N-NW flow)	3.2	4.8	11.3	8.7
	Anticyclonic control	67.8	54.8	53.3	50.9

[1] from Barry, 1972.

TABLE 3-9

SYNOPTIC CLIMATOLOGY OF PRECIPITATION AT BROUGHTON ISLAND,

JULY-AUGUST, 1961-70

Description	Days with Measurable Precipitation	Total (cm)	% of total for 1961-70
Central low/trough	26		37.6
Davis Strait low	11		14.5
Baffin Bay low	9		6.2
Low to SW	8		9.9
Low to SW or S with other lows	9		10.5
Cyclonic Control	63		78.7
Anticyclone	5		6.0
Ridge situations	6		4.0
Ridge, low to 5	10		6.4
High in E, low to W	2		1.6
Ridge, Baffin Bay low (NE flow)	2		1.3
Ridge, Baffin Bay low (N-NW flow)	9		2.1
Anticyclonic Control	34		21.4

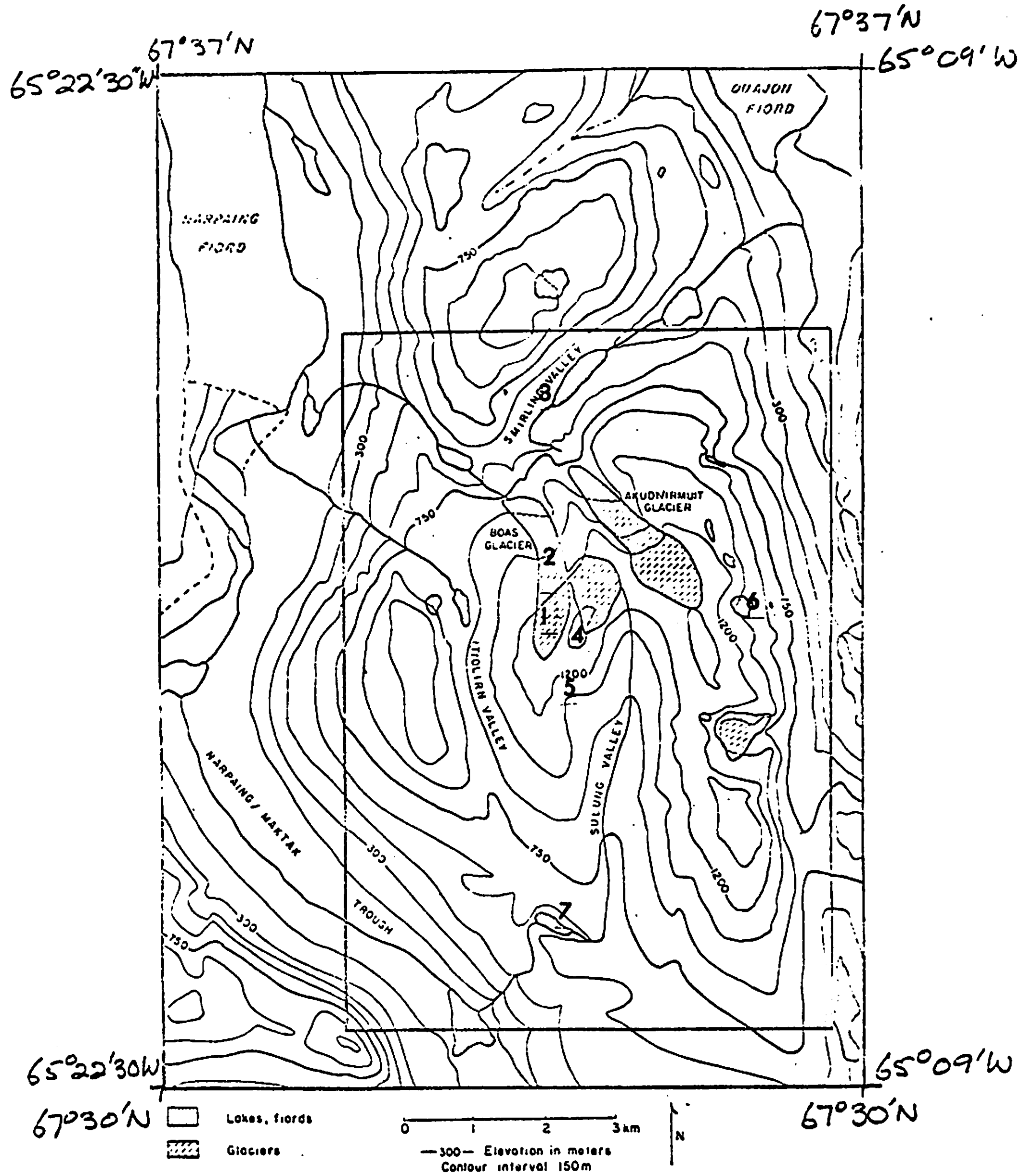


Figure 3-1: Map showing the location of weather stations within the primary field area. 1 = Boas Glacier, 2 = Mid-Glacier, 3 = Base Camp, 4 = Divide, 5 = South Corrie, 6 = East Corrie, 7 = Sulung. See Table 3-1 for list of instrumentation at each station.

FIGURE 3-2

Classification scheme for MSL pressure maps of Baffin Island and adjacent areas (from Barry, 1972)

(a)	Outline of the area to which the classification area relates.		
(b)	Types	100 - 130	Central low or trough
(c)	Types	200 - 220	Baffin Bay low
	"	300 - 320	Ridge, low to south
(d)	Types	400 - 430	High in east, low to west
	"	500 - 510	Davis Strait low
(e)	Types	520 - 540	Baffin Bay trough
	"	542	Davis Strait low
(f)	Types	550 - 560	Inverted low
	"	600 - 620	Low to southwest (with other lows)
(g)	Types	630 - 640	Low to south or southwest with other lows
	"	701 - 702	Northern Baffin Bay low, ridge to southwest
	"	800 - 810	Ridge situations
(h)	Types	900 - 930	Anticyclone

FIGURE 3-2(a)

Outline of the Area to which the Classification Area Relates

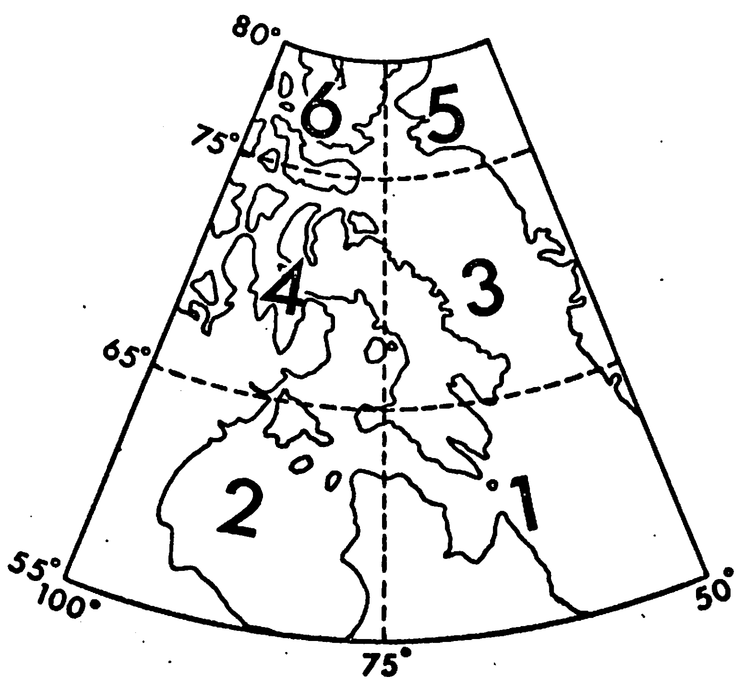


FIGURE 3-2(b): Types 100-130. Central low or trough

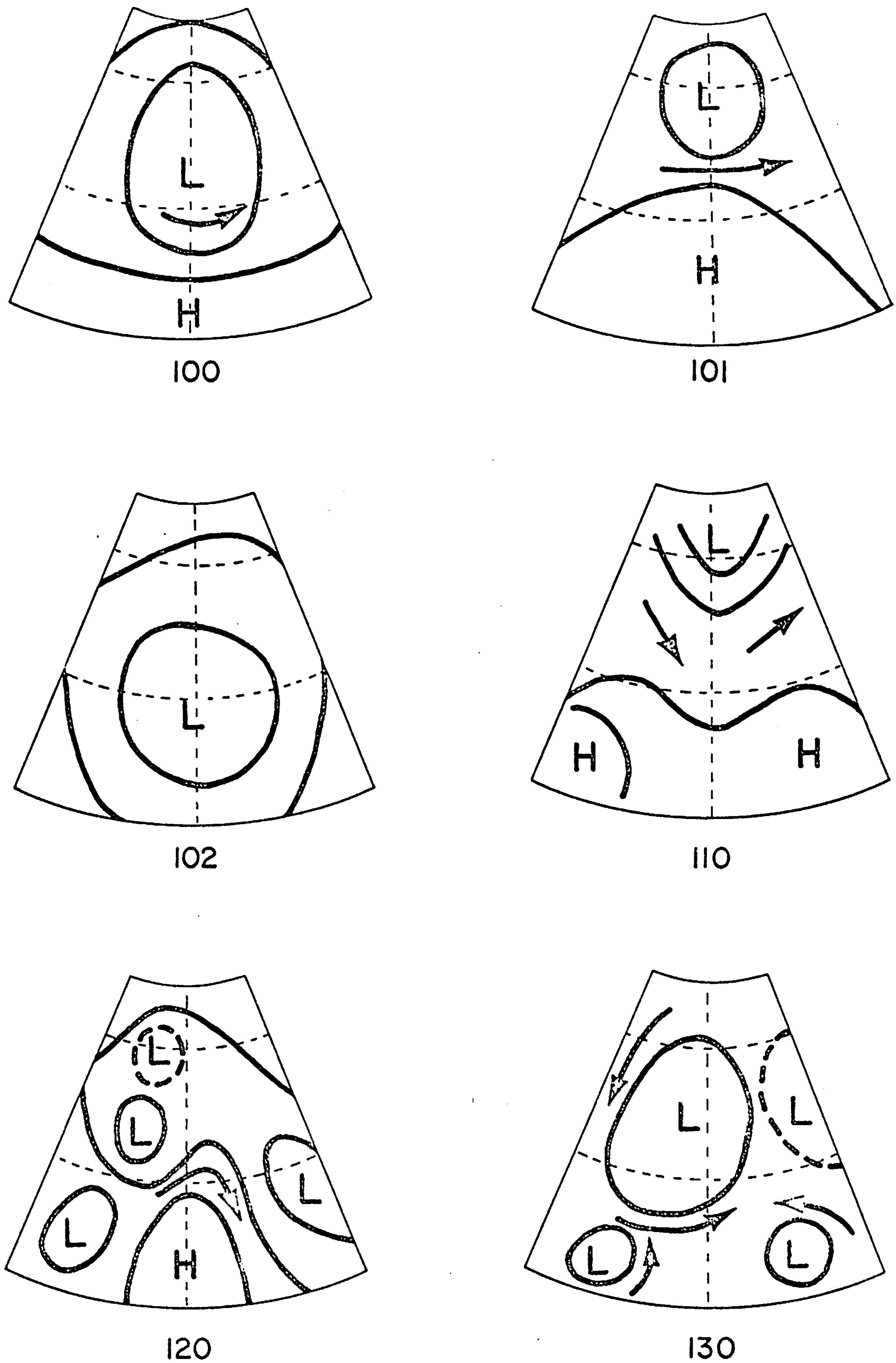
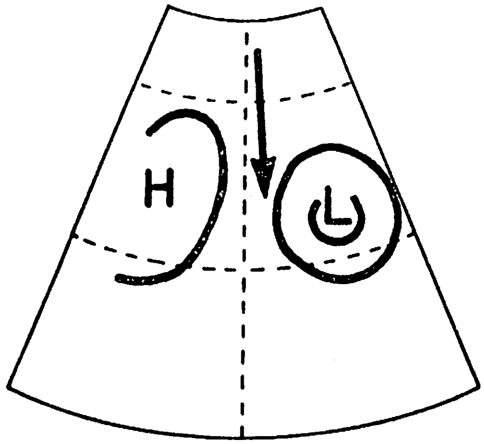


FIGURE 3-2(c): Types 200-220. Baffin Bay low
Types 300-320. Ridge, low to south



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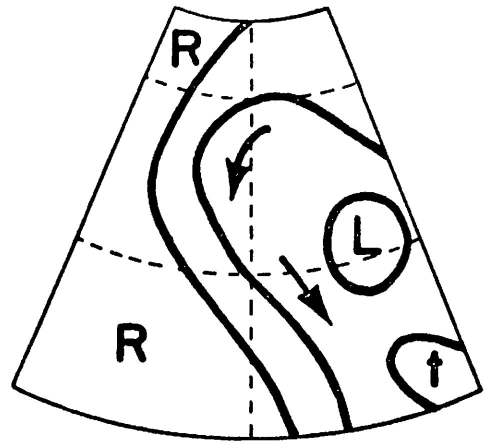

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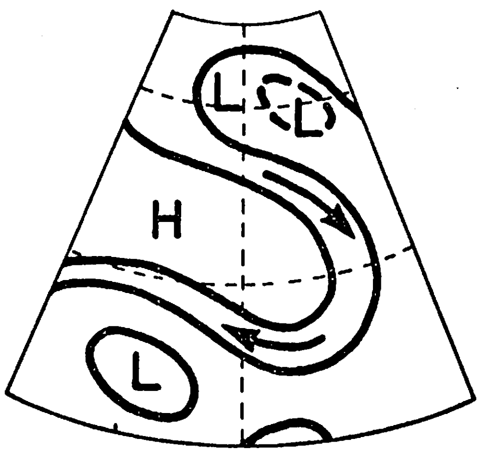

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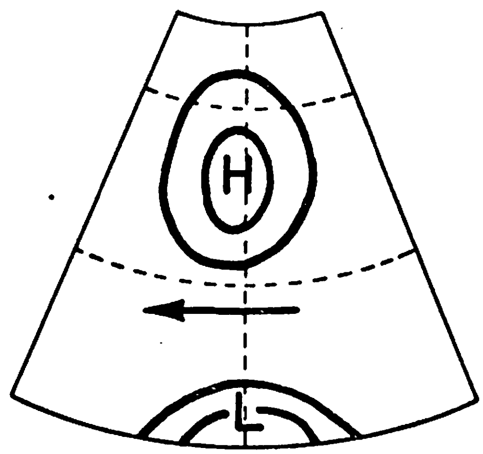

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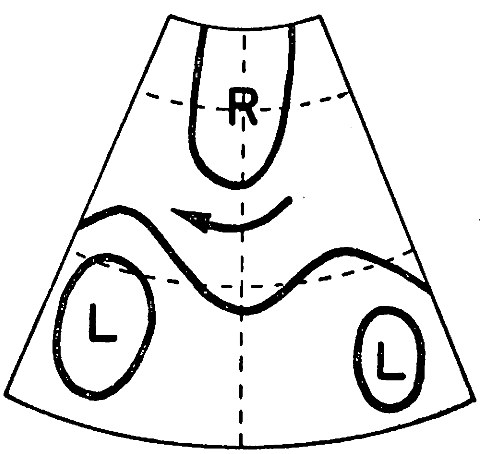

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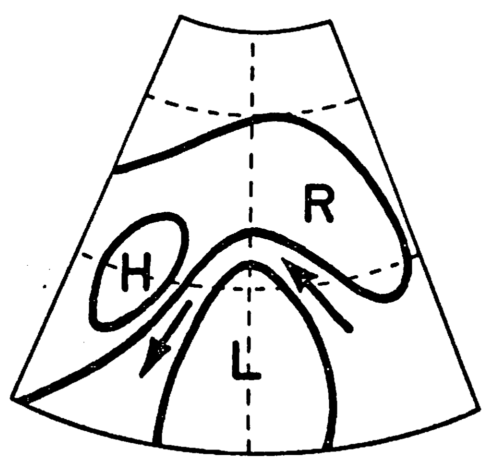


320

FIGURE 3-2(d): Types 400-430. High in east, low to west
Types 500-510. Davis Strait low

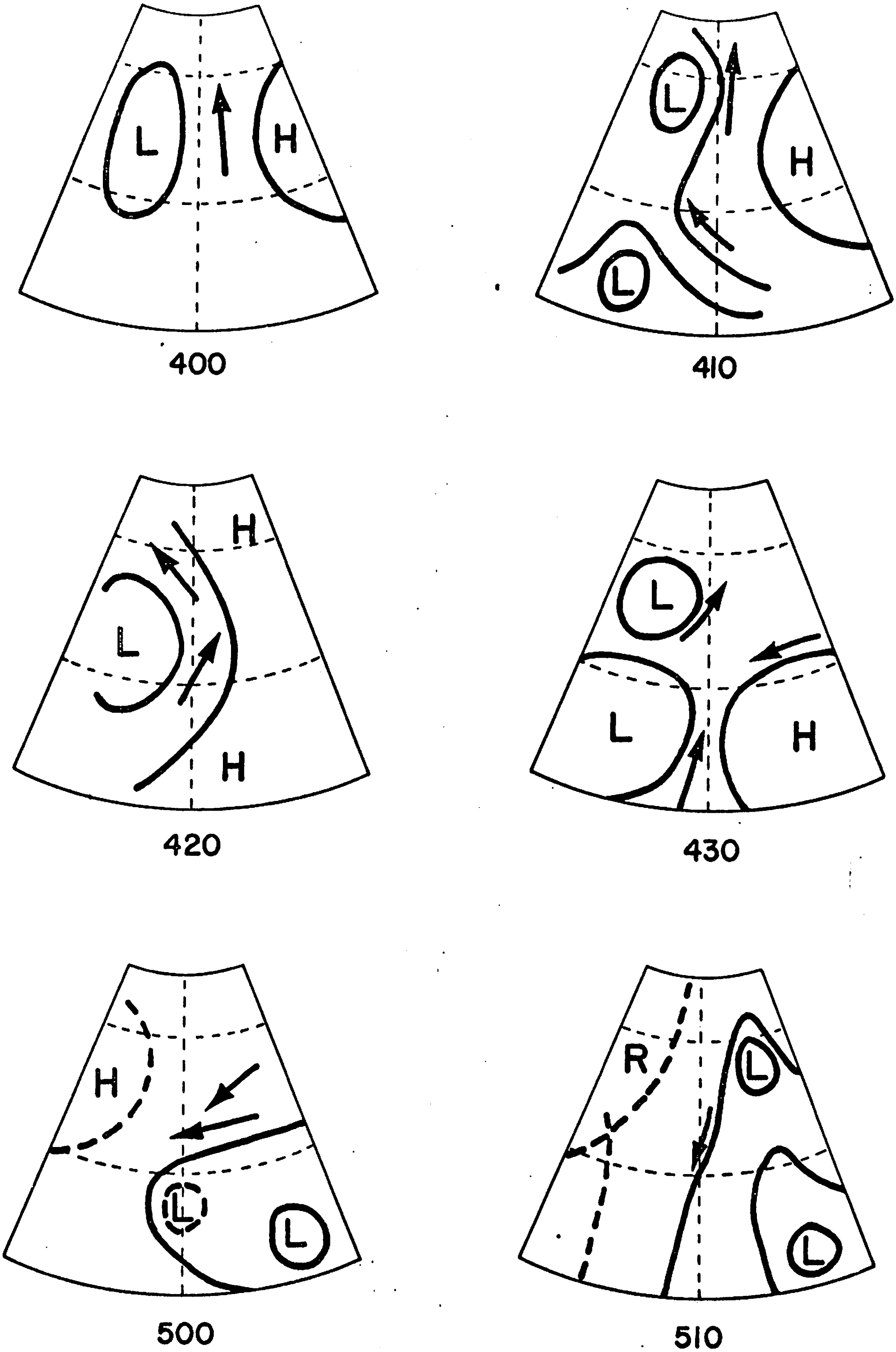
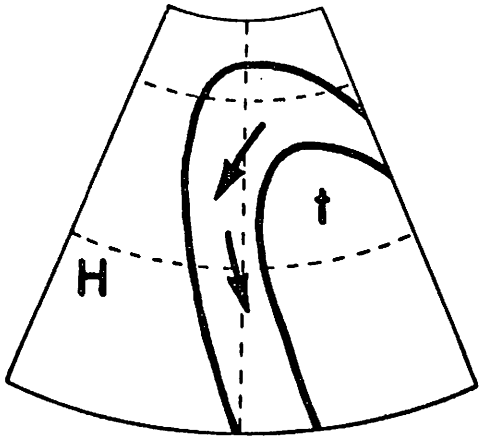
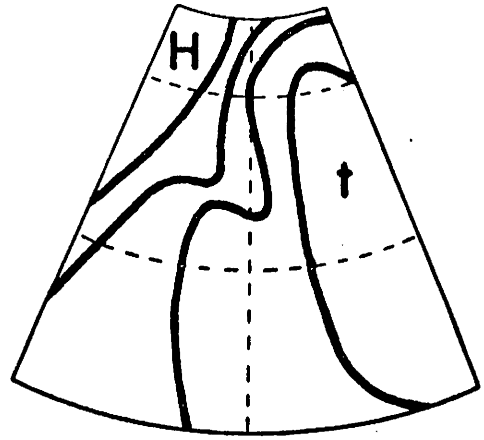


FIGURE 3-2(e): Types 520-540. Baffin Bay trough
Types 542. Davis Strait low



520



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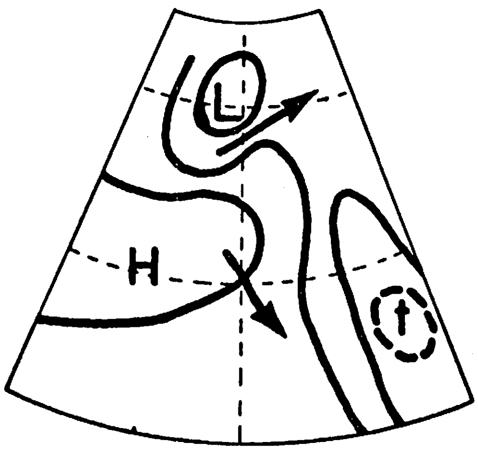
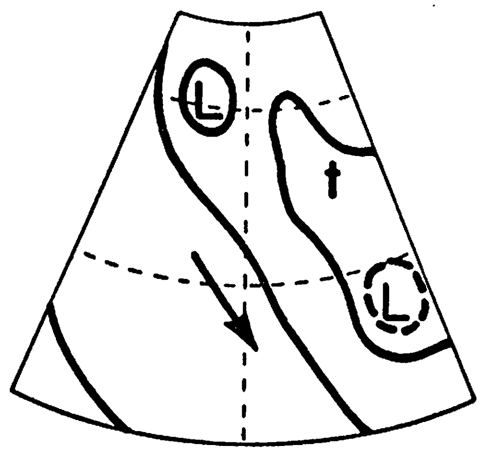


531



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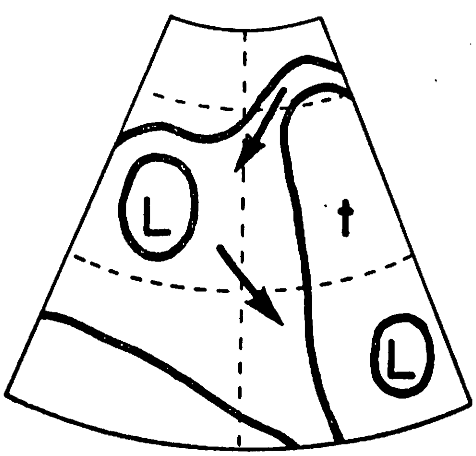

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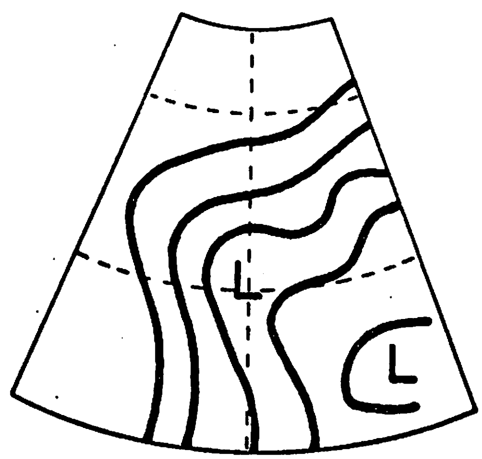


542

FIGURE 3-2(f): Types 550-560. Inverted low
Types 600-620. Low to southwest (with other lows)

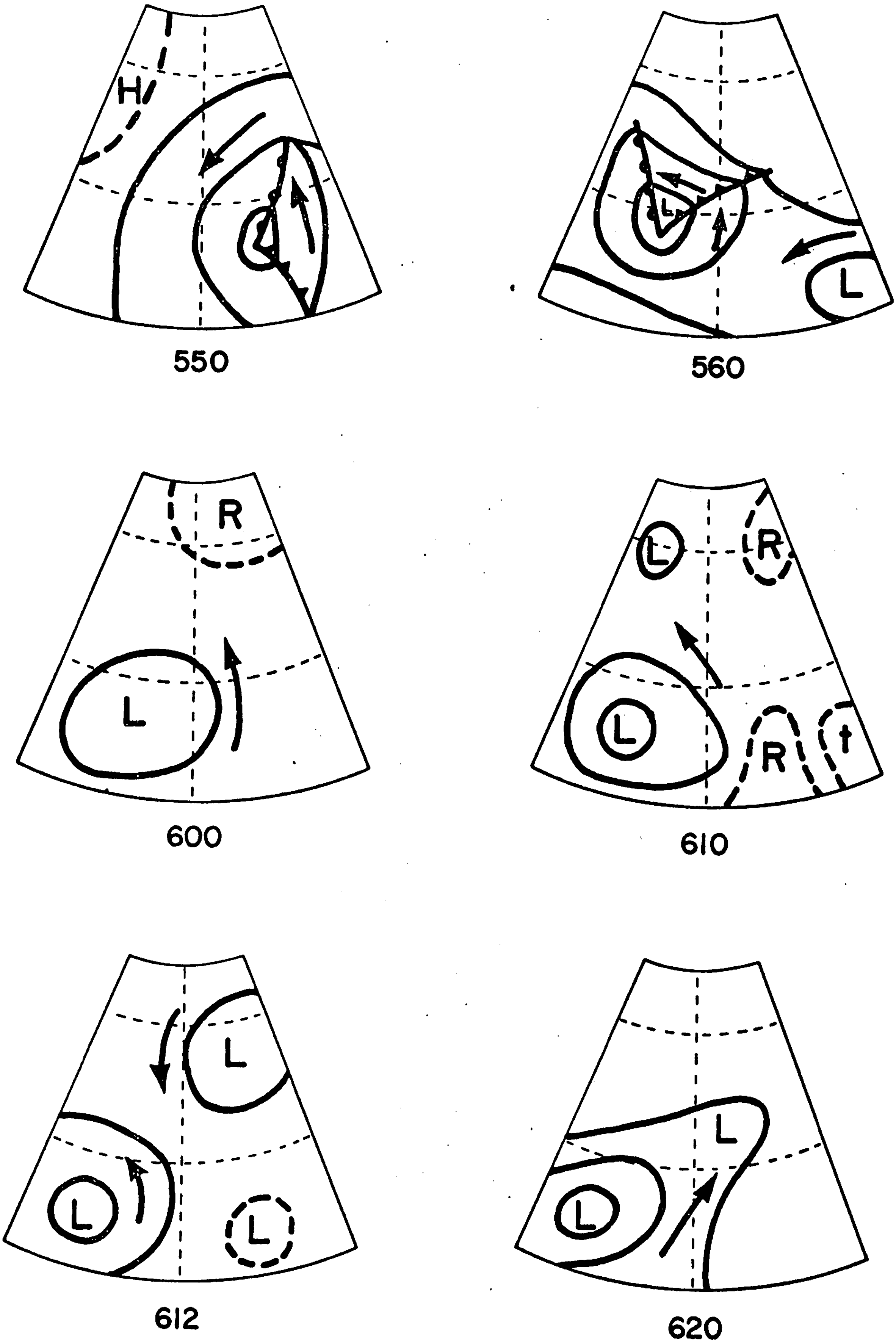
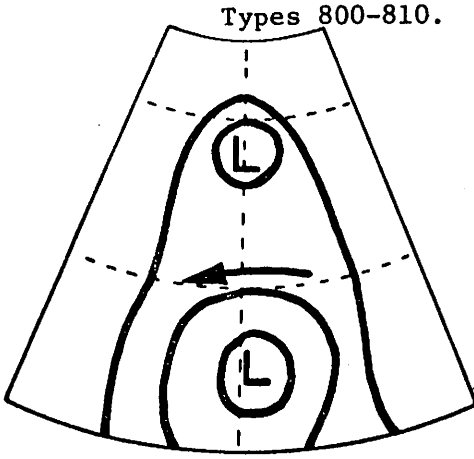
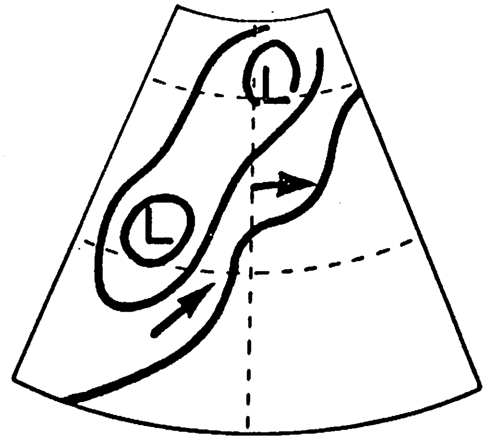


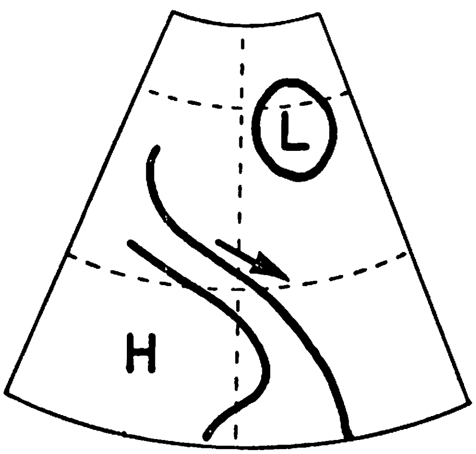
FIGURE 3-2(g): Types 630-640. Low to south or southwest with other lows
Types 701-702. Northern Baffin Bay low, ridge to southwest
Types 800-810. Ridge situations



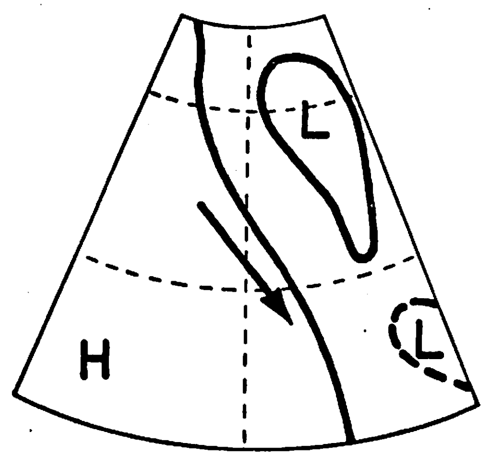
630



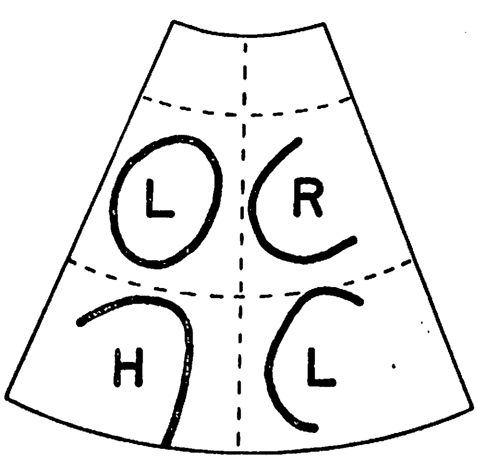
640



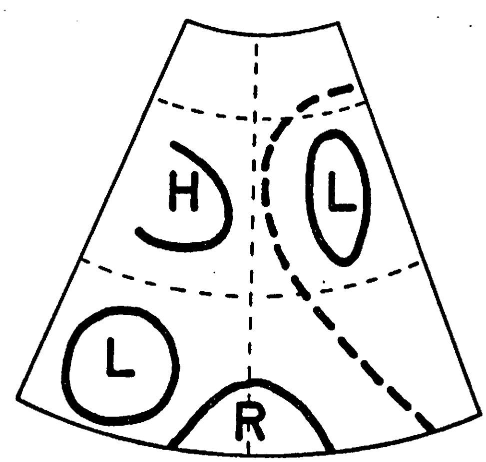
701



702

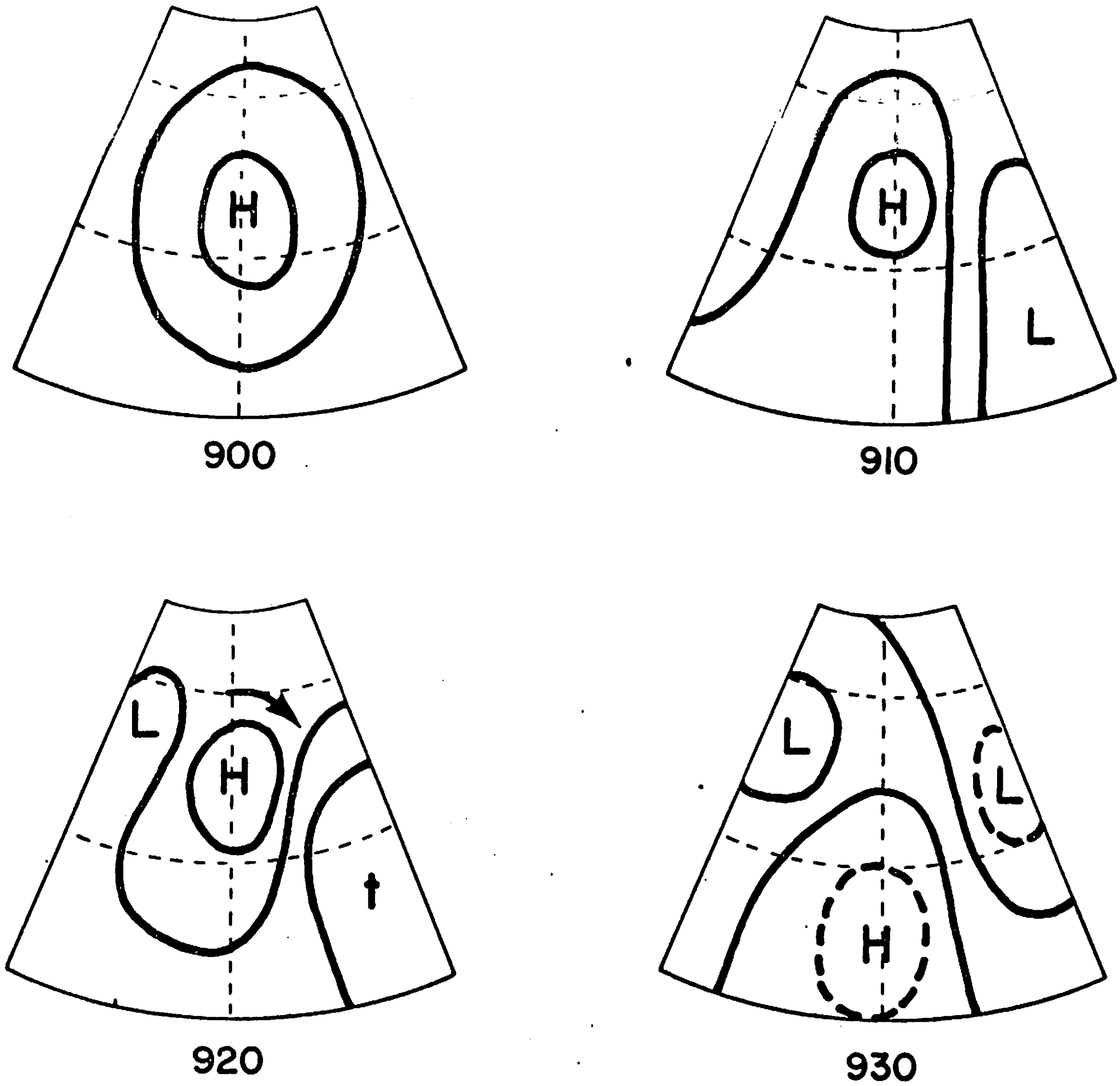


800



810

FIGURE 3-2(h): Types 900-930. Anticyclone



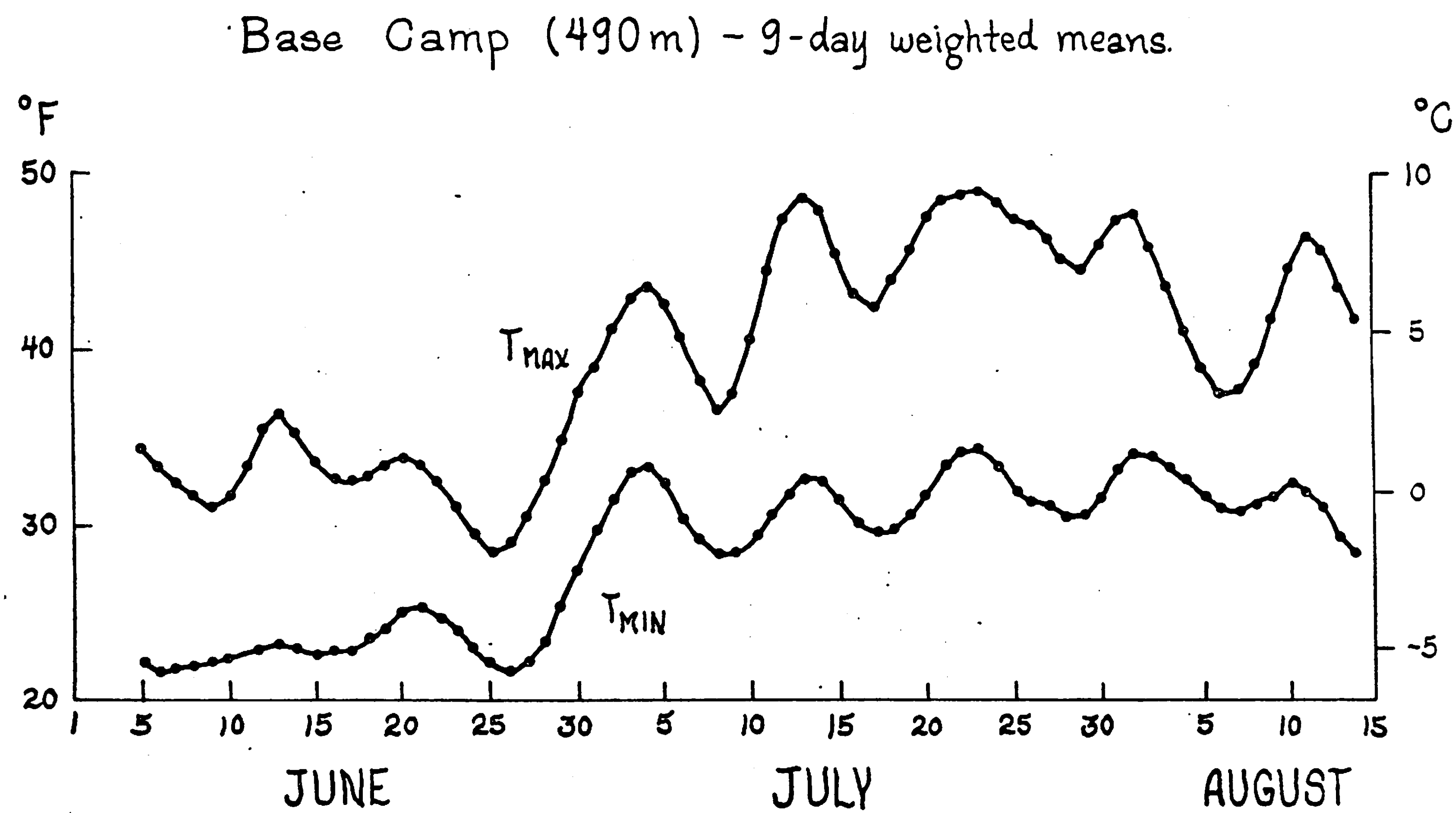


Figure 3-3: Temperature record from Base Camp for the summer of 1970.

4. GLACIAL METEOROLOGY

J. D. Jacobs

PROGRAM

The meteorological program conducted on the Boas Glacier from 9 June to 15 August, 1970, was intended to provide the basis for energy budget calculations for comparison with mass budget measurements, and thus permit determination of the principal factors affecting heat and mass budgets of the glacier and its surroundings. The observations were carried out at an elevation of 1140 m.a.s.l. and on surface of some ten degrees slope with NW aspect. With the exception of the NE quadrant, where the adjacent ridge rises about 18 degrees, there were no obstructions to view. Figure 4.1 shows a horizon diagram and representative sun paths.

The meteorological program consisted of four-hourly synoptie observations and a variety of micrometeorological measurements. The latter included continuous recording of global radiation determined with a Kipp-Zonen pyranometer (in addition to daily actinograph totals), net radiation recorded from a Fritschen radiometer, and temperatures at 40 and 160 cm above the surface. Wind and humidity profiles were measured periodically, as were albedo, subsurface temperatures, and precipitation. A continuous record was obtained over a period of 60 days. During this same period there were 30 days in which melting was observed on the surface, and at the end of the ablation period nearly 0.5 m of snow cover remained. Thus, for the 1970 season, observations relate to the micrometeorology of a glacial snowpack and to the explanation of a positive mass budget year.

Results

Energy budget calculations have been based on measurements at the Glacier Camp site; little variation is expected over the upper section at least of this relatively small and uniform glacier. Results for the season are shown in Table 4.1. Of the various measurements, those pertaining to the radiation fluxes were most satisfactory. The accuracy is estimated to be 10% for global radiation, 20% for net radiation. Difficulties were encountered in obtaining reliable near-surface vapor and temperature profiles. Hence, the values for the turbulent transfer terms are only best estimates based on limited measurements.

TABLE 4-1 ENERGY BUDGET OF THE BOAS GLACIER DURING THE 1970 ABLATION SEASON (9 June - 7 August)

Sources (cal cm^{-2})		Sinks (cal cm^{-2})	
Net Radiation	3700	Warming of snowpack and Glacier	1300
Sensible Heat	1100	Melting of 0.15 m w.e. of snowpack	1200
Latent Heat (condensation)	500	Evaporation and sublimation (residual)	3650
Latent Heat (superimposed ice)	850		
	6150		6150

Of particular interest is the estimate of heat of evaporation and sublimation derived as a residual in the energy balance equation. Thie high value is supported by the fact that small but clearly negative vapor pressure gradients were measured on 65 per cent of those synoptic intervals during which melting was observed. In addition, direct measurements of evaporation from porous cup atmometers indicated a high evaporative potential. Further calibrations are necessary before actual evaporation rates can reliably be inferred from such measurements. The large potential sink which the evaporative term represents, shows that such work is of great importance. In view of the fact that net radiation accounts for about 90 per cent of the energy input to the glacier, consideration of the relationship between radiative flux and mass balance is suggested. On a short-term (daily) basis, two kinds of glaciological data were available for comparison. One was surface lowering as measured with an "ablatometer", an array of rods placed two meters into the ice with a horizontal bar establishing a fixed reference plane. The other daily data were totals of runoff measured with a stream guage at the base of the glacier.

The results of the first comparison are shown in Figure 4.2. A high correlation between surface lowering and net radiation is obtained. With some 50 per cent of the variance in lowering explained by net radiation, the latter quantity becomes a good predictor of ablation. A similar relationship is to be expected for a snowfree surface, with allowance made for differences in albedo.

Runoff provides a better measure of conditions over the glacier as a whole. A stepwise regression on 13 days' data during the ablation maximum shows 66 per cent of the variance in runoff (r) being accounted for by net radiation (Rn), global radiation (s), and mean daily temperatures (T).

$$r = 38.2 + 0.66\ Rn + 0.02\ S - 2.38\ T$$

where r is $m^3\ dy^{-1}$, Rn and S are cal $cm^{-2} day^{-1}$ and T is in ^oC.

Examination of the correlation matrix (Table 4-2) suggests that net radiation and temperature are of greatest importance.

TABLE 4-2 CORRELATION MATRIX FOR GLACIER RUNOFF AND METEOROLOGICAL VARIABLES

	R_n	S	T
Runoff (r)	0.54	0.08	0.57
Net radiation (R_n)	-	-0.43	0.05
Global radiation (S)		-	0.63

These results suggest that an estimate of glacial mass balance, or at least ablation, can be made from temperature and radiation data. A preliminary attempt at this was made in the subsequent summer (1971), using temperatures from an unmanned station in the vicinity of the glacier. Surveys were carried out at the beginning and end of the season which revealed that substantially more ablation has occurred than during the previous year. The temperature record revealed that nearly twice as many melting degree-days occurred between 9 June - 4 August 1971. Assuming a linear relationship between temperature and ablation (with a value of 0.3 cm per degree-day), a proportionaly larger amount of ablation would have occurred in 1971. This compares quite well with actual measurements in 1971 although the relationship is probably non-linear.

TABLE 4-3 AVERAGE TEMPERATURES, MELTING DEGREE-DAYS, AND ABLATION, BOAS GLACIER FOR 9 JUNE - 4 AUGUST 1970 and 1971

	1. Mean temperature (°C)	2. Number of Degree days (0°C)	3. Estimated Ablation (cm H_2O)	4. Measured Ablation (cm H_2O)	5. (4-3)
1970	-1.7	43	-	15 ± 3	-
1971	(1.1)[1]	83	29 ± 6	39 ± 4	-1to 20

[1]Based on regression of minimum temperature between Base Camp and Glacier Camp for 1970.

Net radiation was measured in 1971 at a sea level site on Broughton Island some 50 km east of the glacier. In order to be able to extrapolate from such measurements at a point to the glacier and other surfaces, it is necessary to know how net radiation varies with surface type and synoptic conditions -- the latter being the most readily available areal data. The 1970 observations provide the basis for a tentative system of radiation-day types based on cloud conditions. Cloud height and amount were found to be the principal factors affecting net radiation (Table 4.4). Simply stated, while increased cloud cover results in less global radiation, it has a positive effect on the net longwave term. However, the magnitude of this effect decreases with increasing cloud height. Thus, middle and high clouds, while diminishing the incoming shortwave flux, have little compensating effect in terms of longwave flux. Five distinct radiation-day types based on cloud cover were recognized during the 1970 summer season (Figure 4-3 to 4-7). An analysis of variance on net and long-wave terms demonstrated the significance of this classification.

The radiation-day approach makes it possible to estimate a radiation budget using calculated clear-sky global radiation, estimated longwave flux, and cloud cover data. For example, in the extreme case of no clouds during the 60 day period of observation, a calculated net radiation of 4700 cal cm^{-2} compares with the 3700 cal cm^{-2} measured under actual cloud conditions, which averaged 6 tenths for the period. This approach is being developed more fully using a combination of satellite, synoptic, and upper air data in conjunction with field observations. Some applications to palaeoclimatic reconstruction are discussed by Barry in Chapter 12.

Table 4.4 Radiation Data -- Boas Glacier, Summer, 1970

Daily totals given in langleys

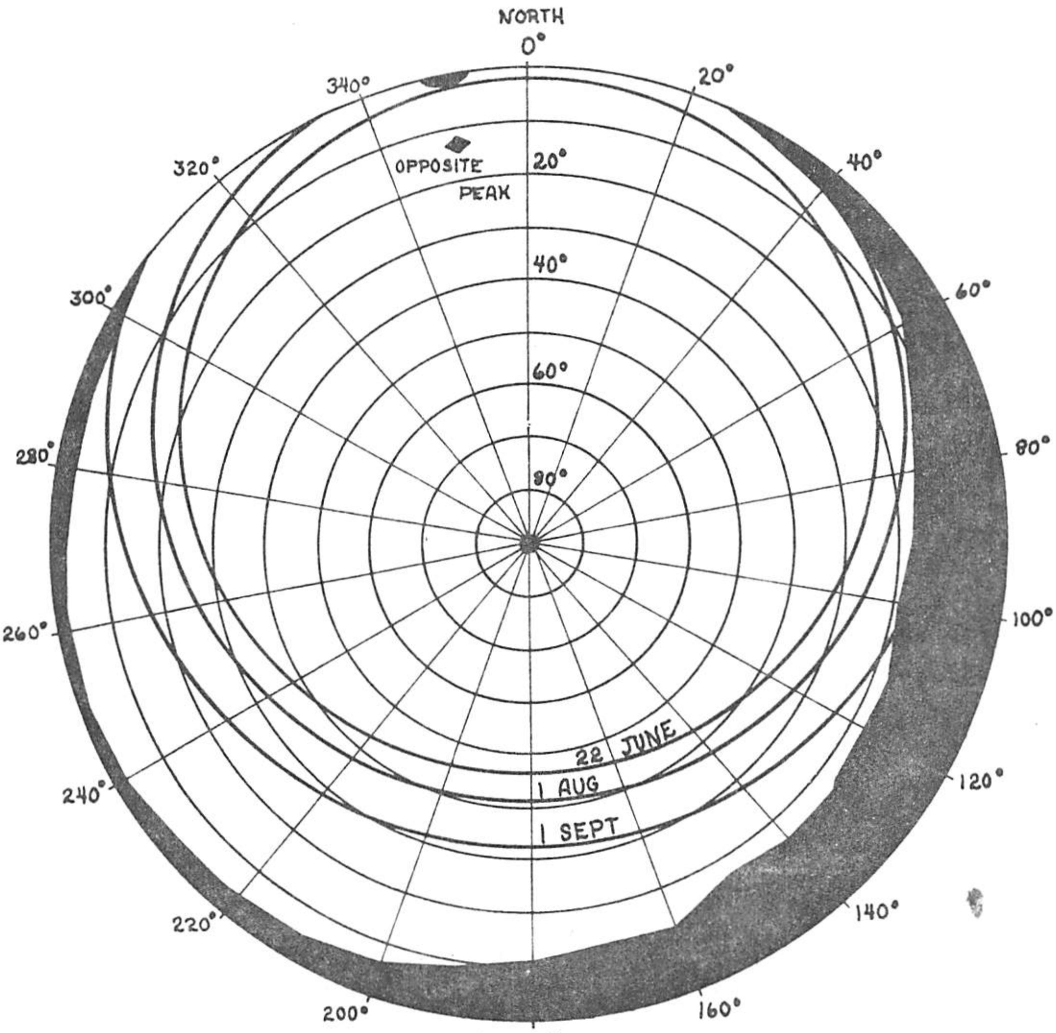
Date	S	a(%)	S(1-a)	R_n	L_n	Cloud	Weather	Radiation Type
June 9	505	90	50	64	14	10L	s	A
10	371	90	37	51	14	10L	s	A
11	470	90	47	47	0	10L		A
12	551	76	132	36	-96	01M		B
13	550	77	126	57	-69	05M		
14	464	78	102	100	- 2	09LM	s	
15	451	90	45	29	-16	10L	s	A
16	406	90	41	43	2	10L	f	A
17	684	75	172	121	-51	06L		
18	348	90	35	21	-14	10L	s,f	A
19	576	83	98	49	-49	07M		C
20	484	80	96	56	-40	08M		C
21	333	88	40	-11	-51	08M		C
22	430	90	38	83	45	08M		C
23	362	85	54	80	26	08M		C
24	400	90	40	35	- 5	10L	s,f	A
25	569	86	80	25	-55	08L	f	
26	385	90	38	51	13	10L	s,f	A
27	497	87	65	68	3	10LM		A
28	695	75	174	23	-151	02L		B
29	577	64	207	61	-146	01LM		B
30	511	73	138	100	-38	06LM		
July 1	389	90	38	49	4	10L	s,f	A
2	431	81	82	44	-38	08MH		C
3	385	90	39	47	8	10LM	f	A
4	643	69	199	97	-102	00		B
5	489	76	118	107	-11	07MH	s	C
6	347	90	31	48	17	10L	s,f	A

Date		S	a(%)	S(1-a)	R_n	L_n	Cloud	Weather	Radiation Type
	7	385	78	85	37	-48	09LM	s,f	
	8	380	90	38	34	- 4	10LM	s	A
	9	354	90	35	22	-13	10L	s,f	A
	10	424	88	52	9	-43	09L	s,f	
	11	669	85	100	-30	-131	00		B
	12	664	78	146	71	-75	01H		B
	13	600	82	109	95	-14	02MH		B
	14	550	67	182	79	-103	08H		E
	15	576	62	218	36	-182	02H		B
	16	578	66	196	106	-90	07M		C
	17	602	75	151	94	-57	02MH		B
	18	480	83	90	106	16	07M		C
	19	306	90	61	51	-10	06LM	f,s	
	20	444	71	129	125	- 4	04MH		
	21	531	70	160	96	-64	02M		B
	22	428	62	162	92	-70	06M		C
	23	488	73	132	130	- 2	03M		
	24	236	81	45	70	25	10M	s	D
	25	398	86	56	24	-34	10L	s,f	A
	26	574	70	172	71	-101	02M		B
	27	565	73	153	49	-104	02H		B
	28	508	70	152	51	-101	03LMH		
	29	468	77	107	50	-57	07M	f	C
	30	546	61	212	51	-161	00		B
	31	558	66	189	56	-133	00		B
Aug	1	463	73	125	84	-41	06M		C
	2	353	70	106	46	-60	07H		E
	3	214	82	38	72	34	10M	s	D
	4	276	75	69	104	35	09L		
	5	194	75	50	54	4	10L	r,s	A
	6	242	75	59	82	23	10L	f	A
	7	287	75	71	78	8	10L	r,s,f	A
Totals	60	27555	-	5698	3676	-2029			
Average		459	79	95	61	-34			

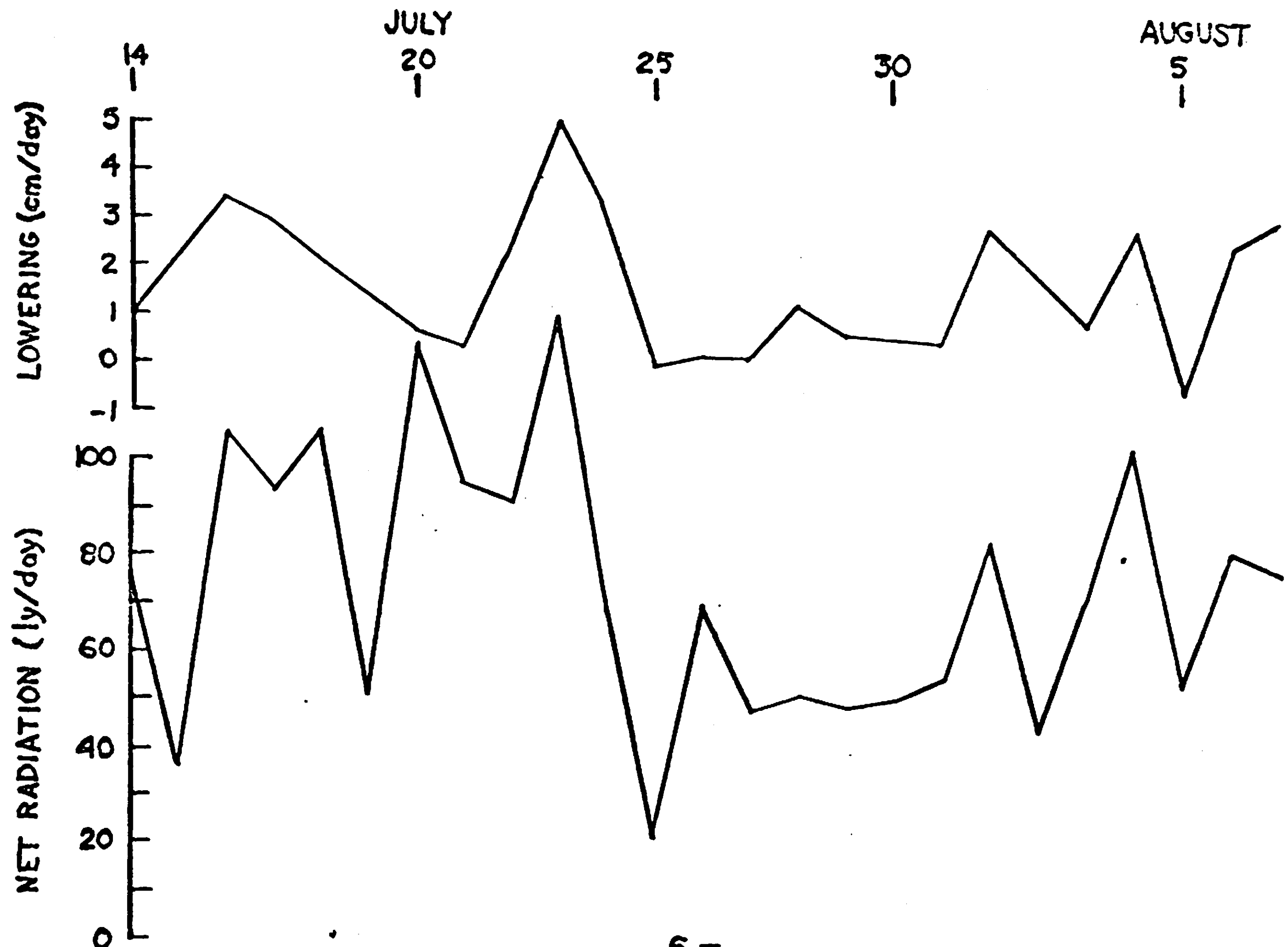
Notes:

1. Measured radiation values (S,R_n) assumed correct.
2. Albedo usually that measured at 1200EST relative to glacier slope.
3. Cloud data refer to predominant conditions for that day.
4. s = snow, r = rain, f - fog

Figure 4-1: Horizon diagram for the main Glacier Camp meteorological station



Horizon Diagram
Glacier Camp
Boas Glacier 1140 m



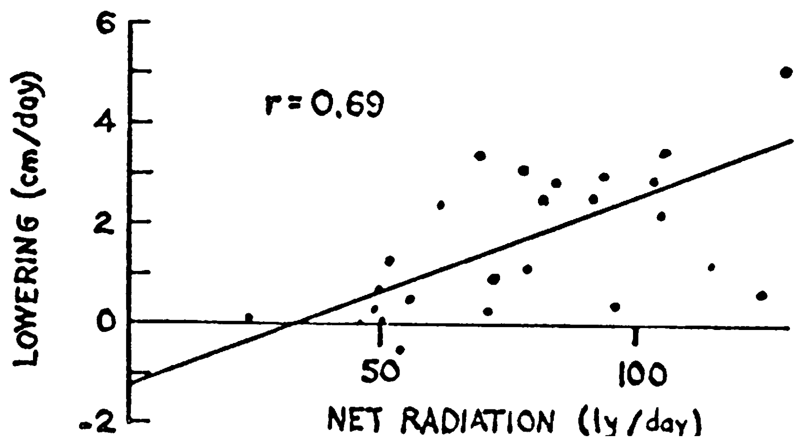


Figure 4-2: Daily surface lowering (cm/day) and net radiation (ly/day) at Glacier Camp, 14 July - 7 August 1970. The lower diagram shows the correlation (r = +0.69) between lowering and net radiation

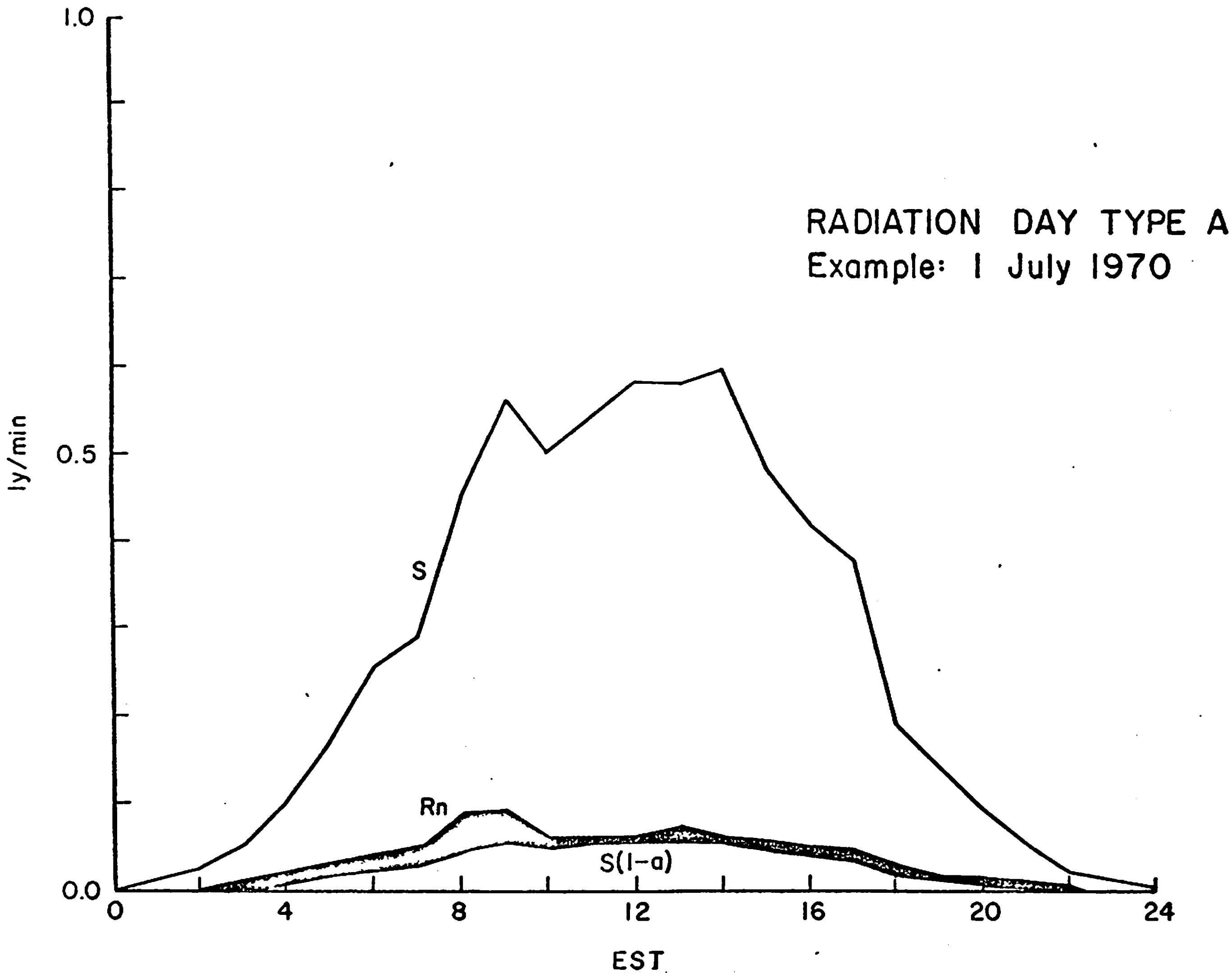


Figure 4-3: Radiation day type A

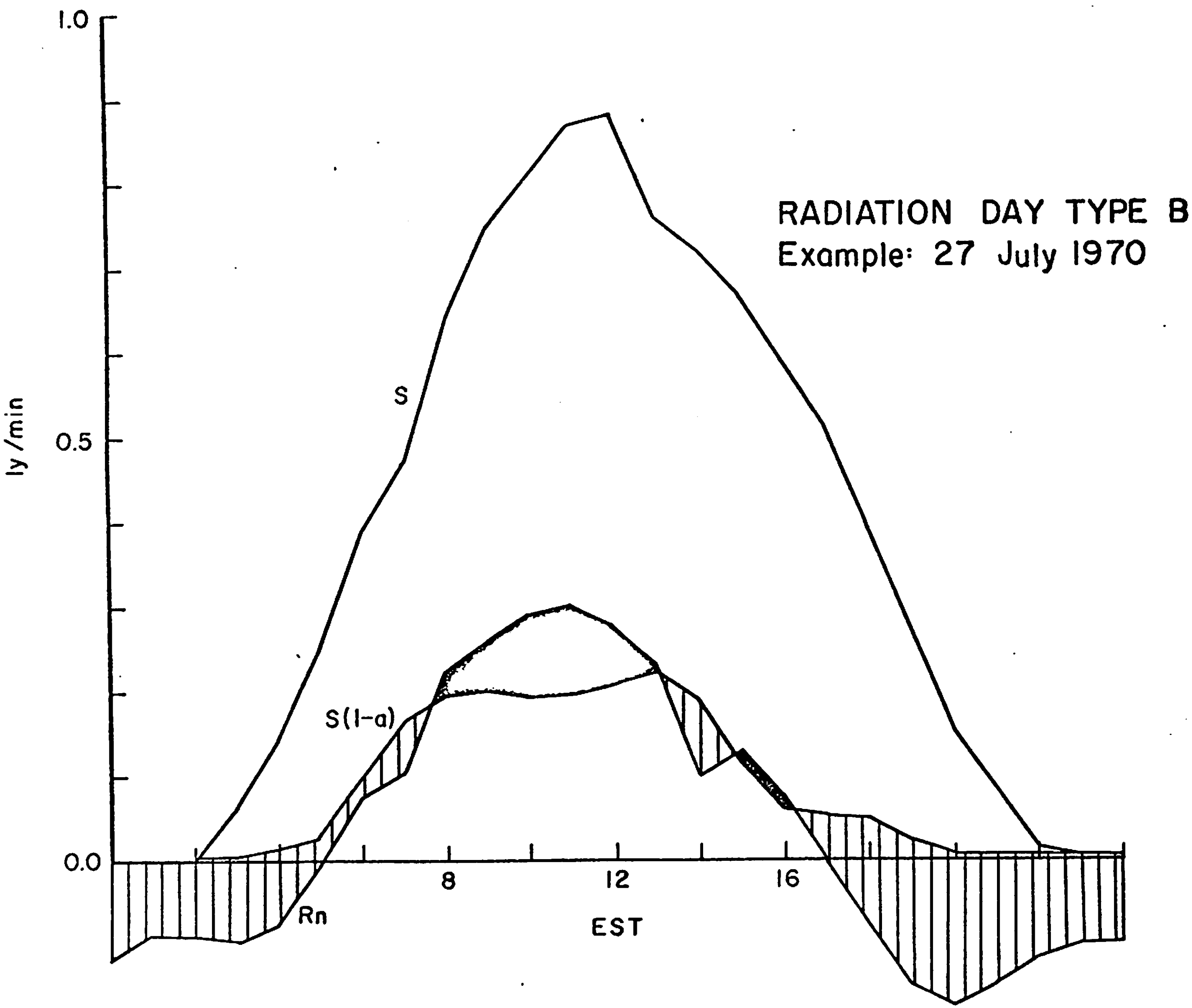


Figure 4-4: Radiation day type B

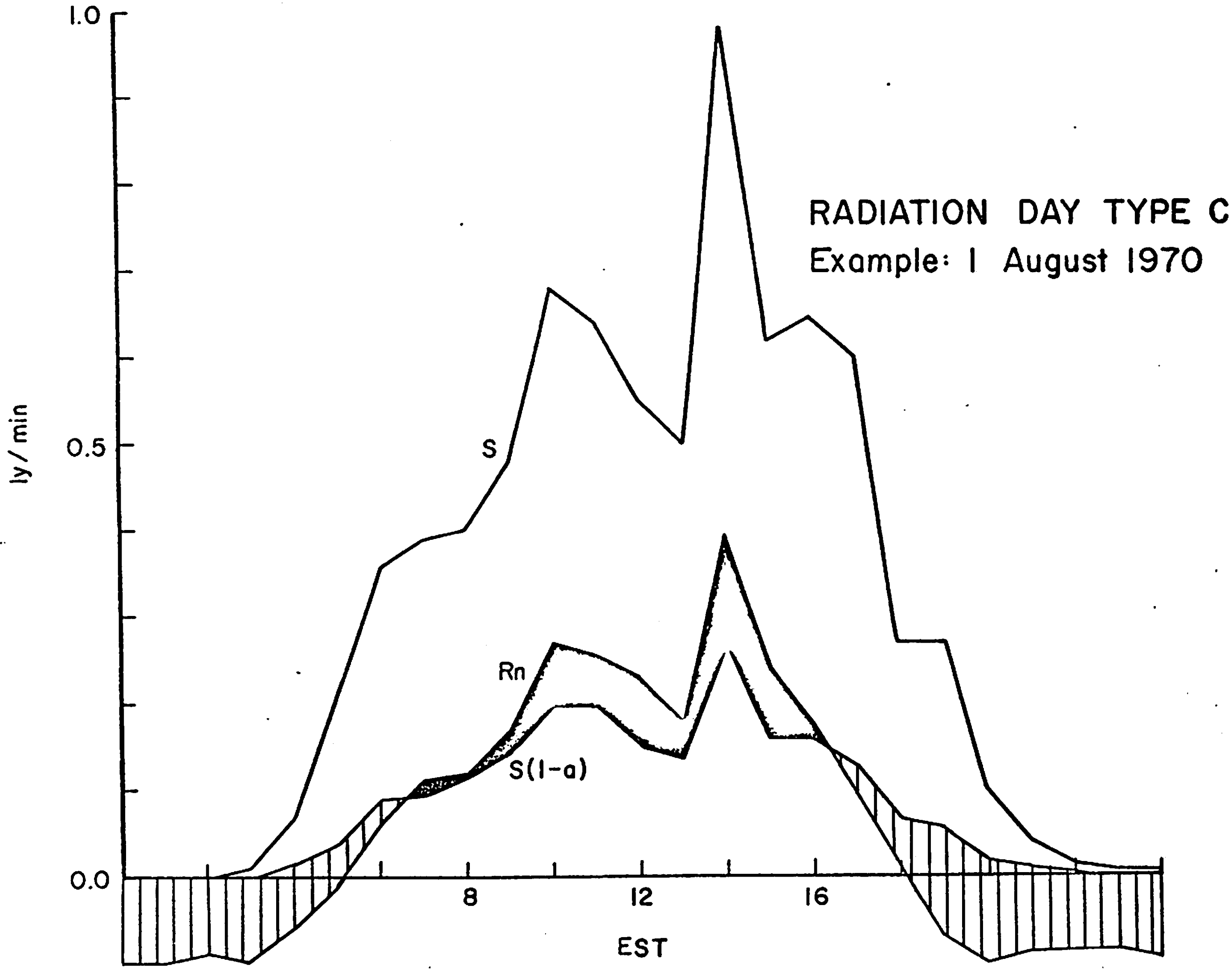


Figure 4-5: Radiation day type C

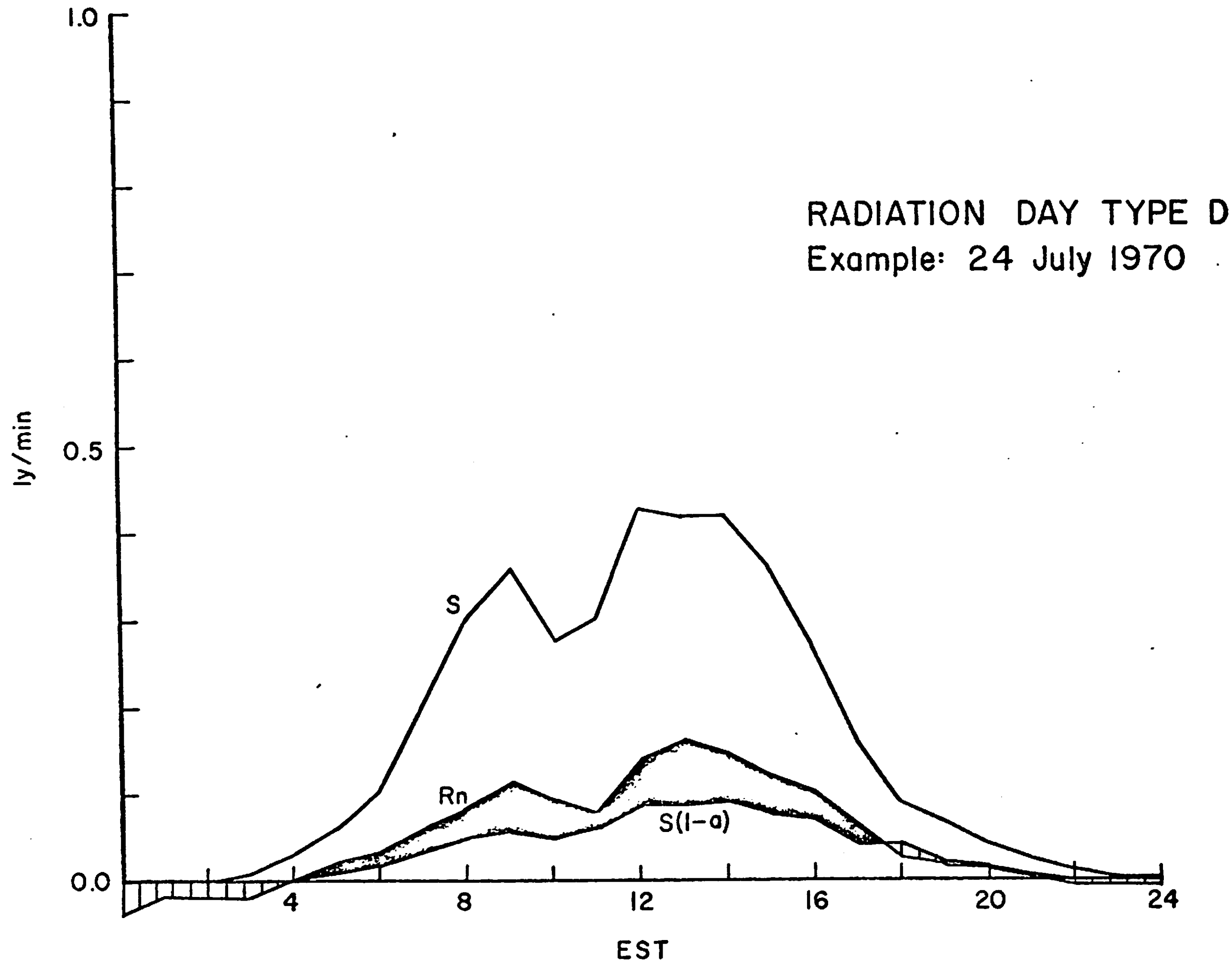


Figure 4-6: Radiation day type D

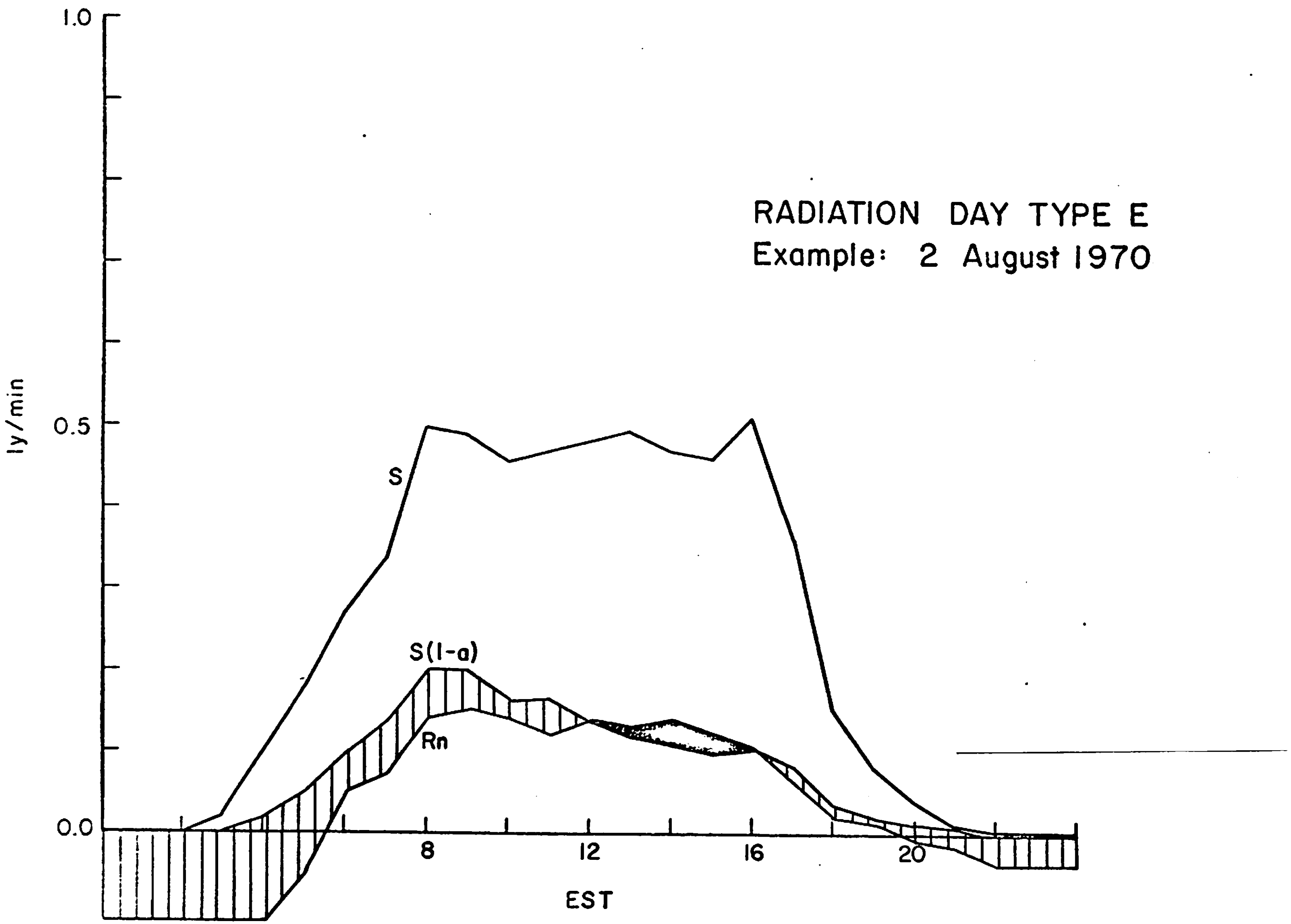


Figure 4-7: Radiation day type E

5. CLIMATIC DATA SUMMARY

R. G. Barry, J. D. Jacobs, R. Weaver and L. D. Williams

The basic climatological data on temperature, humidity and precipitation for the seven stations are summarized in Tables 5.1-5.2. Temperature and humidity data are daily maxima and minima determined from hygrothermographs except at Base Camp and Boas Glacier where the screen temperatures are from maximum and minimum thermometers. These were read at 20.00 EST at Base Camp, but at Boas Glacier they refer to the period 00-24.00 EST. These daily tabulations are uncorrected for radiation errors (see discussion in Chapter 3).

Precipitation amounts were collected in "Tru-chek" gauges. At Mid-Glacier, Divide and South Corrie these were read at about weekly intervals on servicing or other visits. At Base Camp, Boas Glacier and Sulung daily totals were obtained. These refer to the period ending 20.00 EST at Sulung and 24.00 EST at Boas Glacier where the observations were four-hourly.

At Boas Glacier the June precipitation, which all fell as snow, was determined from snow depth x 0.1. (Attempts to construct a device for weight assessments of net precipitation were unsuccessful.) For July and August snow depth and gauge measurements were made. The comparison between the two sets of data were good for July (gauge: 0.75 in; snow-fall x 0.1: 0.71 in) when almost all of the precipitation was of frozen form. The tables show the gauge data at Boas glacier for July and August; rain and snow were distinguished where possible from the daily weather records.

The Tru-chek gauges very probably underestimate the true precipitation, especially for snowfall. In the more exposed sites (Mid-Glacier, Boas Glacier and Divide) the snow was often wind-blown which would certainly reduce the catch in the small orifice of these gauges. Blowing and drifting snow may on other occasions have caused a positive error. This seems a likely reason for the large total at Mid-Glacier between 28 June-3 July compared with Base Camp and Boas Glacier over the same period. The absence of catch at Divide between 26-30 June is also problematic in view of the precipitation at Base Camp and Boas Glacier.

At the more remote stations evaporation and sublimation must have removed some precipitation from the gauges between the storms and the time of reading. Even when air temperature remained below $0^{\circ}C$ the gauges were sufficiently heated during periods of strong insolation to melt rapidly the ice within them.

The temperatures for Base Camp, which are less subject to radiation errors than those on the glacier, show daily maxima $\geqslant 43^{\circ}C$ (the conventional growing season threshold) on 3 days in June, 18 days in July and 9 days from 1-18 August.

Table 5.3 shows 164 melting degree days (mean daily temperature $>32^{\circ}F$) for Boas Glacier between 9 June-14 August compared with 179 at Broughton Island for the same period, although the former figure is probably too high due to the radiation errors.

The incoming solar radiation at Boas Glacier (Table 5.3), based on daily actinograph charts, averaged 515 cal cm^{-2} day^{-1} for 9-30 June and 511 cal cm^{-2} day^{-1} in July. These may be compared with averages of

483 and 484 cal cm^{-2} day^{-1}, respectively, from digitally integrated records on the Kipp Zonen pyranometer (see Table 4-1). The actinograph was carefully calibrated with respect to the Kipp subsequent to the field program at the 10,000 m climatological station of the Institute 25 miles west of Boulder and an explanation of the 6% discrepancy in the measurements in Baffin Island has not yet been determined. It is probably unrelated to the small temperature coefficient of the Kipp; the usual accuracy of actinograph totals is within about 10%.

TABLE 5.1 DAILY EXTREMES OF TEMPERATURE (^{o}F) AND RELATIVE HUMIDITY
(TEMPERATURES UNCORRECTED FOR RADIATION ERRORS)

JUNE	BASE CAMP (490m)				MID-GLACIER(986m)		BOAS GLACIER (1142m)			
1970	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	Tmax	Tmin	RHmax	RHmin
1	43	27	97	48						
2	43	21	96	40						
3	37	28	95	48						
4	41	25	100	40						
5	25	18	98	88						
6	36	18	96	54						
7	32	25	100	38						
8	37	23	90	45			32	14	100	60
9	27	21	100	78			27	16	100	70
10	23	21	100	89			30	17	100	72
11	30	23	100	57			31	18	100	70
12	41	23	70	37			35	18	85	30
13	50	27	95	32	49	24	46	30	100	24
14	33	21	100	80	38	22	31	20	100	50
15	27	21	100	85	31	20	29	17	(100)	(60)
16	30	25	98	77	32	21	34	18	100	58
17	34	21	100	85	34	21	30	22	100	79
18	31	21	98	52	32	19	29	17	100	81
19	36	25	68	37	34	19	41	16	100	42
20	36	27	71	43	37	18	27	17	96	75
21	31	27	65	52	32	20	27	17	96	79
22	37	25	86	52	40	21	33	20	96	74
23	33	25	99	80	44	24	41	22	100	75
24	27	23	100	93	31	18	30	18	100	80
25	25	21	100	01	33	18	32	17	100	72
26	23	19	100	74	21	17	21	15	100	100
27	33	21	97	68	38	16	36	16	100	75
28	33	21	100	70	41	16	29	16	100	77
29	37	26	100	73	51	23	48	20	100	52
30	35	28	100	82	45	25	44	26	100	62
MEAN	33.5	23.2	94.0	62.9	(36.8)	(20.2)	(33.2)	(18.6)	(98.8)	(66.0)

AUGUST	BASE CAMP (490m)				MID-GLACIER (986m)		Boas Glacier (1142m)			
1970	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	Tmax	Tmin	RHmax	RHmin
1	55	42	57	44	53	30	50	32	87	60
2	47	32	100	63	44	28	40	28	92	63
3	40	32	92	64	41	28	36	28	100	83
4	41	34	100	82	45	32	43	33	100	85
5	59	32	100	77	37	31	36	26	100	72
6	34	30	100	90	37	24	32	24	100	100
7	37	30	100	83	38	24	34	25	100	88
8	34	30	100	90		22	39	22	100	84
9	36	29	98	60	36	21	34	20	100	83
10	49	35	73	42	43	27	41	25	100	62
11	58	36	60	28	44	33	39	33	89	46
12	53	36	100	52	46	28	40	32	100	72
13	39	26	100	92	34	20	32	21	100	87
14	29	24	97	70	25	12	21	14	100	100
15	45	20	85	30	39	10	37	12	100	32
16	44	36	92	40	37	21	32	27	100	77
17	42	34	96	51						
18	44	34	89	48						
MEAN	(42.6)	(31.7)	(91.0)	(62.0)	(39.9)	(24.4)	(36.6)	(25.1)	(98.0)	(74.1)

JUNE	DIVIDE (1300m)				EAST CORRIE (1150m)				SOUTH CORRIE (970m)			
1970	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	RHmax	RHmin
12	30	18	88	32								
13	46	30	64	25								
14	27	20										
15	28	18							30	19	100	61
16	(24)	17							31	19	100	64
17	29	21										
18	29	17			25	17	100	76				
19	33	16			28	18	84	56				
20	26	18			29	20	88	60				
21	25	18			26	20	80	64				
22	32	20			32	22	78	62				
23	31	23			31	23	10	70				
24	28	18			34	28	99	83				
25	32	18	100	76	24	18	96	65				
26	20	15	100	100	18	16	94	65				
27	26	16	100	100	30	16	99	67				
28	28	19	100	85	31	19	95	67				
29	49	22	100	32	36	23	99	55				
30	41	22	100	54	29	23	99	74				
	(30.7)	(19.3)	(94.0)	(63.0)	(28.7)	(20.2)	(93.2)	(66.5)				

JULY 1970	DIVIDE (1300m) Tmax	Tmin	RHmax	RHmin	EAST CORRIE (1150m) Tmax	Tmin	RHmax	RHmin	SOUTH CORRIE (970) Tmax	Tmin	RHmax	RHmin	SULUNG (610m) Tmax	Tmin
1	33	24	100	84	29	24	100	86		27	100	56		
2	35	22	100	78	31	24	97	70	36	26	90	55		
3	37	22	100	68	37	28	90	70	47	30	100	53		
4	38	24	100	70	40	29	99	60	53	31	100	38		
5	45	27	100	55	38	29	98	58	53	30	100	36		
6	28	20	100	81	29	22	98	88	33	25	100	70		
7	34	21	100	79	30	22	98	78	43	25	100	52		
8	27	20	100	100	28	23	99	82	32	26	100	74		
9	24	20	100	95	26	22	99	82	28	25	100	77		
10	34	20	100	79	33	22	96	71	44	26	100	58		
11	35	28	100	60	37	28	98	50	46	31	88	42		
12	54	29	100	51	46	34	85	60	60	31	85	41		
13	44	34	84	48	47	40	67	42	56	31	85	46		
14	55	33	80	46	52	35	86	52	58	35	60	40		
15	49	29	92	54	42	27	100	70	48	25	84	52		
16	53	25	100	53	38	25	100	70	43	25	100	60		
17	32	25	100	68	33	26	99	58	43	26	100	57	36	31
18	41	25	86	52	42	32	96	56	53	27	100	60	48	31
19	35	27	100	66	38	30	94	64	44	30	100	66	47	35
20	39	28	94	50	40	31	70	52	46	30	90	42	52	36
21	38	28	80	58	42	32	74	59	43	30	80	48	51	40
22	42	32	85	48	39	34	80	62	45	37	78	56	50	40
23	40	30	83	58	44	34	80	54	48	37	75	48	55	41
24	39	26	100	76	40	28	100	75	43	30	100	66	46	37
25		24	100	100	28	27	98	82	31	28	100	80	37	32
26	39	24	88	46	40	29	90	47	44	30	100	38	51	32
27	38	26	86	50	41	30	90	49	46	32	78	40	52	38
28	32	22	86	72	34	26	93	72	39	29	80	40	41	30
29	28	22	94	75	30	24	100	78	35	26	100	68	32	28
30	34	24	86	58	33	29	96	42	38	25	100	55	44	28
31	38	24	70	38	42	36	60	40	36	31	64	29	54	37
MEAN	36.7	25.3	93.3	65.0	37.1	28.4	91.3	63.9	43.8	28.9	91.5	53.0	(46.4)	(34.3)

AUGUST	DIVIDE (1300m)				EAST CORRIE (1150m)				SOUTH CORRIE (970m)				SULUNG (610m)	
1970	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin	RHmax	RHmin	Tmax	Tmin
1	40	30	68	56	43	34	68	54	47	35	62	43	53	45
2	34	23	100	55	38	27	95	54	38	29	100	54	36	31
3	32	24	90	64	35	28	94	63	37	29	100	59	42	31
4	36	28	100	60	37	32	94	55	43	34	95	68	39	34
5	29	22	100	60	34	24	100	60	37	27	100	52	37	32
6	26	22	100	100	27	24	96	95	30	26	100	100	34	30
7	30	22	100	72	28	24	96	95	32	26	100	100	37	31
8	30	18	100	80	28	20	96	95	31	24	100	81	34	29
9	25	16	100	80	28	22	94	88	29	23	100	80	35	29
10	29	20	100	64	36	25	80	60	37	26	86	50	50	30
11	35	27	80	43	40	33	70	40	45	35	70	32	54	44
12	35	24	100	63	38	33	70	62	42	30	68	34	51	35
13	25	17	100	80					30	22	100	100	33	27
14	14	12	100	100					27	17	100	80	27	25
15	26	10	100	60					43	17	100	25	49	23
16									34	30	64	38	45	37
17														
18														
MEAN	(29.7)	(21.0)	(95.9)	(69.1)	(34.3)	(27.2)	(86.1)	(68.4)	(36.4)	(26.9)	(90.3)	(62.2)	(41.0)	(32.3)

TABLE 5.2 PRECIPITATION DATA, 1970 (inches)

JUNE	BASE CAMP		MID-GLACIER	BOAS GLACIER		DIVIDE	SOUTH CORRIE	COMMENTS
6		-	-		0	-	-	
7		-	-		0	-	-	
8		-	-		0	-	-	
9	S		-	S	.14	-	-	
10	S	.22*	-	S	.20	-	-	*Estimate from snow depth
11		0	-	S	t		-	
12		0	-		0		-	
13	r	.02	-	rS	.02		-	
14	S	.06	-	S	.04			
15	rS	.04	-	S	.02			
16	S	.05		S	.05	0.0		
17		0		S	t			
18	S	.31		S	.20			
19		0			0			
20		0			0			
21		0			0			
22		0		S	t	.30		
23	S	t		S	.04			
24	rS	.09		S	.05			
25	S	.01		S	t	.10		
26	S	.11		S	.28*			*Possibly affecte by drifting
27	S	.02	.50	S	t			
28		0			0			
29	S	t			0			
30	S	.03		S	t	0	.74	
TOTAL		.96 (9-30th)	.50 (16-27th)		1.04 (6-30th)	.40 (11-30th)	.74 (14-30th)	

PRECIPITATION DATA, 1970 (INCHES)

JULY	BASE CAMP		MID-GLACIER	BOAS GLACIER		DIVIDE	SOUTH CORRIE	SULUNG
1	rS	.05		S	.04	0		-
2	rS	t		S	.03			-
3	r	.03	.38*	S	.04			-
4		0			0			-
5	r	t		S	.02			-
6	S	.16	.12	S	.09			-
7	S	t		S	.01	.11		-
8	S	.05		S	.03			-
9	S	.16	.0	S	.20			-
10	S	.16		S	.06		.50	-
11		0	.01		0			-
12		0			0	.13		-
13		0			0			-
14		0			0		.0	-
15		0	.0		0			-
16		0			0			-
17		0			0	.0		.01
18	r	t		r	.01			.01
19	r	.10	.20	rS	.15			.17
20		0		rS	.02			.02
21		0			.12			0
22		0	.02	r	t			.01
23		0		r	t			0
24		0		r	t			0
25	r	.04		S	.05			.09
26		0			0	.50		0
27		0			0			0
28		0			0		.10	0
29		0	.09		0			0
30		0	.0		0			0
31		0	.0		0	.0		0
TOTAL		.75	.82 (28th June-31th July)		.75	.74	.72	(.31) (17-31)

* May include drifting.

PRECIPITATION DATA, 1970

AUGUST	BASE CAMP		MID-GLACIER	BOAS GLACIER		DIVIDE	SOUTH CORRIE	SULUNG
1		0			0			
2		0			0			
3	r	t		rS	.01			.01
4		0		r	t		.01	.01
5	r	.11		rS	.17			.10
6	r	.03		r	t			.07
7	Zr	t		S	.15	.12		.01
8	S	.21	.34	S	.15			.21
9	S	.05		S	.01			0
10		0	.13		0		.41	0
11		0			0		-	0
12	r	.01		S	.12	.18	-	.08
13	S	.11		S	.17		-	.22
14	S	.21		S	.06		-	.21
15		0			0	.24	-	-
16		0	.39		0	-	-	-
17	S	.01	-		-	-	-	-
18		0	-		-	-	-	-
19	rw	-	-		-	-	-	-
TOTAL		.74 (1-18th)	.76 (1-16th)		.84 (1-16th)	.54 (1-15th)	.42 (1-10th)	.92 (1-14th)

TABLE 5.3 SOLAR RADIATION AND MELTING DEGREE DAYS (MEAN -32) AT BOAS GLACIER 1970

	JUNE		JULY		AUGUST	
	S(CAL CM^{-2})	MDD	S(CAL CM^{-2})	MDD	S(CAL CM^{-2})	MDD
1	-	-	449.4	0	534.0	9
2	-	-	522.0	1	408.0	2
3	-	-	408.6	4	235.2	0
4	-	-	(667.2)	1	294.6	6
5	-	-	516.0	4	210.6	0
6	-	-	363.6	0	282.0	0
7	-	-	411.6	0	346.2	0
8	-	0	459.0	0	350.4	0
9	555.6	0	444.0	0	304.2	0
10	406.8	0	504.6	1	424.8	1
11	517.2	0	735.0	4	297.6	4
12	508.8	0	730.2	9	302.4	4
13	605.4	6	546.0	8	218.4	0
14	510.0	0	(661.2)	11	280.8	0
15	(495.0)	0	597.0	8		0
16	(466.8)	0	580.2	4		0
17	751.8	0	501.6	0		-
18	382.2	0	434.4	8		-
19	634.2	0	336.6	3		-
20	532.2	0	486.6	6		-
21	366.0	0	583.8	7		-
22	(474.6)	0	497.4	9		-
23	398.4	0	466.2	10		-
24	439.8	0	271.2	3		-
25	625.2	0	457.8	2		-
26	429.0	0	631.2	6		-
27	480.6	0	505.2	8		-
28	447.6	0	505.8	2		-
29	684.0	2	404.4	0		-
30	622.2	3	628.8	1		-
31			522.6	7		-
TOTAL	11,133.4	(11)	15,831.2	127	4,489.2	(26)
MEAN	(515.2)		510.7		(320.7)	

Appendix: Lightweight Instrument Shelter

A lightweight aluminum instrument shelter was developed for field use by John Clark in connection with the Baffin Island program. This is illustrated in the accompanying figure. The materials required are:

2 sheets 0.063" Aluminum 36" x 96" (semi-hardgrade)

1" x 1" Al tubing 1/8" wall thickness

10-24 machine screws with wing nuts and lock washers

2 pcs. piano hinge 3" long

1 transom catch

Misc: 1/2" Al angle for stiffening door edges and attachment of sides to legs

4 eye bolts and turnbuckles

Approximately 7 man-hours were required to louvre the sides and for fabrication (including the final assembly). The shelter weighs 10 kg.

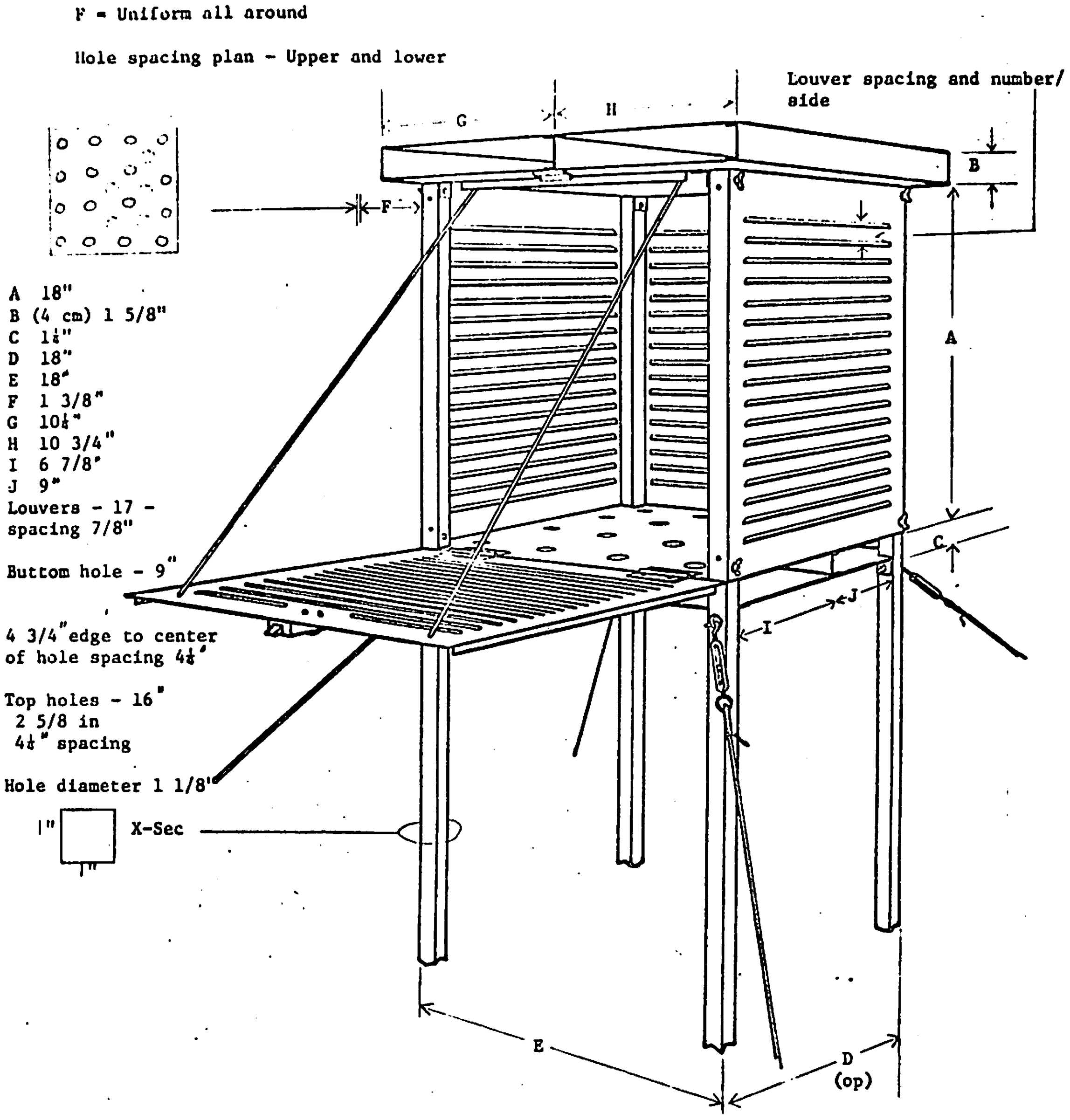


Figure 5-1: Perspective view of the light weight meteorological screen developed and used in this project. The screen has been modified on the basis of 2 years use and is now used extensively for remote mountain weather stations run by INSTAAR in Baffin Island, the Colorado Front Range and the San Juan Mountains of Colorado.

6. GLACIER MASS BALANCE (BOAS GLACIER) 1969-1970 AND 1970-1971

R. Weaver and J.T. Andrews

INTRODUCTION

The Boas Glacier is located at 67°35'N and 65°16'W and is numbered as 46204J68 in the Canadian Glacier Inventory. The elevation of the snout is 800 m a.s.l. and in its 2.4 km length the glacier rises to 1300 m a.s.l. The area of the glacier is 1.45 x $10^6 m^2$ (±) and, assuming an average thickness of 50 m, a volume of 70 x $10^6 m^2$. The glacier has two accumulation areas (Figure 6-1) - one is a direct continuation of the lower glacier whereas the East Bowl joins the main glacier through a very narrow, and possibly very shallow, channel. A thin ice cover links the Boas with the Akudnirmuit Glacier to the east (Figure 1-1). Both glaciers steepened appreciably at about 1120 m a.s.l. where they descend the backwalls of submerged corries. Both glaciers face toward N 20°W.

The Boas Glacier was selected for the glaciological program because of relative ease of access, its simple shape, and the lack of any appreciable hazards such as major crevasse zones. However, later studies on the radiation receipts and elevations of corrie glaciers within the Okoa Bay 1:250,000 map area (Williams, 1972 and Chapter Ten) indicate that the Boas lies close to mean values. Average calculated direct beam short-wave radiation of August 15 is 122 cal $cm^{-2}dy^{-1}$ for all glacierized corrie basins and the value for the Boas Glacier is 118 cal $cm^{-2}dy^{-1}$. Average elevations at the base of the backwall is 1014 m a.s.l. and the Boas lies at 1050 m a.s.l. Thus although we cannot claim that the Boas Glacier was selected by any rational process other than those itemized above, it is worth stressing that the glacier lies close to the average values on two important environmental parameters and hence the Boas Glacier is representative of the small glaciers in the area north and east of the Penny Ice Cap

INSTALLATIONS

In 1968 H. Kryger installed a Hoinkes storage precipitation gauge on the lip overlooking Quajon Fiord. In 1969 the gauge was emptied by Williams and Weaver (it had accumulated 0.3m H_2O in the year) and moved to the base of the Akudnirmuit Glacier and three other Hoinkes were installed (see Chapter Three). That same year 6 stakes were drilled in to the upper Akudnirmuit Glacier.

The main field thrust occurred in 1970. Stakes were installed along the length of the Boas Glacier (Figure 6-1), thermistors were installed in at 5 sites a velocity profile and strain diamond were installed and surveyed (Chapter Seven), and run-off was measured at two sites by colorimetry methods and the use of stream level recorders (Chapter Eight).

The Boas Glacier was revisited in June and August 1971 and sufficient measurements were made to compute the 1970-1971 mass balance. We anticipate a similar low level of maintainance during the 1971-1972 mass balance year.

MASS BALANCE 1969-1970

The terminology used in this report will follow that laid down in the UNESCO/IASH technical paper (1970) entitled "Combined heat, ice and water balances at selected glacier basins", specifically in Appendix 2 of that report which is called "Mass Balance terms". Some of the results below have appeared in the paper by Jacobs, _et al_. (1972).

1969-1970 Winter Balance:

The 1970 field season started its measurement program on the Boas Glacier on the 1st June. The winter balance was measured by a system of snow pits (Figure 6-1) and snow probes. A CRREL snow kit was used to measure snow density in the pits although the weighing was done on a small spring balance. The results are shown on Table 6-1. The mean density is virtually identical to that on the Penny Ice Cap in 1953 (Ward and Baird, 1954) where a value of 0.33 g cm^{-3} was reported and is also very similar to results from the Barnes Ice Cap in north-central Baffin Island

(Sagar, 1966; Anonymous, 1967; Løken and Andrews, 1966). No trend of density with elevation is apparent from these data (Table 6-1).

Snow probing on either side of the main stake line was conducted in June/early July prior to the start of the ablation season. The results are grouped into altitudional intervals on Table 6-2. A total of 292 probes were taken giving an average snowpack of 1.223 m with a standard error of the mean of $\pm$ 0.0187 m. The results of the snow probing (Table 6-2) appear to indicate that the snowpack decreases in thickness as a function of elevation. Using the full stake network (N=32) a regression indicates a relationship of the form:

$$\text{Snow depth (m)} = (-9.05 \times 10^{-4})(\underline{x}) + 2.187 \qquad (6\text{-}1)$$

where:

$\underline{x}$ is elevation in meters. The correlation coefficient, r, has a value of -0.56 and the amount of explanation is 32%.

The computation of specific winter balance, b_w, and areal winter balance B_w was greatly facilitated by the presence of glacier ice and superimposed ice at all elevations on the Boas Glacier. Thus the definition of the 1969 summer surface was unambiguous. Tables 6-3 and 6-4 give the data for $\bar{b}_w(z)$ and $B_w(z)$ by altitudional increments of 150 m (where z stands for the altitudional zone). The areas of the zones were planimetered from maps at 1:125,000 scale. The average specific net winter balance, $\bar{b}_w$, is 0.379 m H_2O or 0.363 m H_2O. The latter value is 'weighted' to take into account the snow depth at the main control snowpits relative to the probed values. These estimates can be compared with 0.390 m H_2O derived simply as the average snow density x average snow depth. The winter balance on the Boas Glacier is estimated at .533 x $10^6 m^3$ H_2O or .500 x $10^6 m^3$ H_2O.

The water equivalent of the snowpack on the Boas Glacier compares with the September to June (1969-1970) precipitation measurements on Broughton Island of .40 m. This value was 200% of the decadel average.

Summer balance and net balance 1969-1970:

The summer of 1970 was characterized by low ablation and considerable solid precipitation. No bare ice was exposed when the camps were vacated on August 20th and the Boas Glacier was mantled with a cover of snow. The net balance for the 1969-1970 balance year is calculated on the basis of snowpit data, observations on the growth of superimposed ice at the stake network, and probing. The growth of superimposed ice is especially critical and the combined results of stake and probe values (Figure 6-2) appears to have provided reliable estimates of this phenomena. Thus reliability is, however, in no small measure related to the presence of ice at the 1969 summer surface which meant that percolation into lower layers of the glacier was virtually impossible. A 5 m core at 1120 m a.s.l., well above any steady state equilibrium line altitude, showed no firn at all down to that depth.

Densities within the snowpack in mid-August had risen well above their June values. The average of 36 density determinations was 0.429 g cm^{-3} with a standard error of the mean of $\pm$ 0.001, whereas the average snow depth was still 0.662 m H_2O by mid-August. Clearly the net balance for the 1969-1970 year was highly positive! The net balance, $\bar{b}_n(z)$, on about August 18th 1970 is tabled on Table 6-5. The average net specific balance ($\bar{b}_n$) over the entire glacier is computed to be +0.389 m H_2O and the net balance B_n is computed to be .54 x $10^6 m^3$ H_2O (Table 6-6). Both these latter estimates are based on computations within altitudional zones. A simple specific net balance $\bar{b}_n$ based on the stake data alone (20 stakes) gives a positive balance of +0.374 m H_2O. Tables 6-5 and 6-6 should be compared with 6-3 and 6-4 which list the winter balance figures. Upon due consideration of all the errors involved in mass balance calculations (Anonymous, 1967) the two sets of data show no important change between the winter balance as of late June and the net balance as of mid-August. The glacier had received a moderate amount of precipitation during the summer, however, as discussed in Chapters 3 and 5. As the glacier did not change its mass balance appreciably during the summer it appears that the average specific net loss of mass over the course of the summer was of the order 0.06 - 0.03 m H_2O. The Boas Glacier was thus acting very much as a closed system during this particular balance year. Surface lowering was partly the result of changes in pack density and partly due to melting. However, this mass was not transferred out of the system but was stored at the base of the 1969-1970 snowpack in the form of superimposed ice. Run-off from the glacier during this year was accordingly very limited as discussed further in Chapter Eight. The effective change, e, during the year was basically 0.

Surface Lowering: Ablation Gradient:

Figure 6-2 illustrates the trend of surface lowering as a function of elevation and compares this with the net balance curve for the 1969-1970 year. The values suggest that there is an overall gradient in the surface lowering values equivalent to a 1 mm m^{-1}. This value can be compared to 1.6 mm m^{-1} computed from the data presented by Ward and Baird (1954).

Ice/Snow Temperatures:

Yellowspring thermistors were installed around June 12th at elevations of 831, 932, 1043, 1147 and 1229 m a.s.l. The thermistors were installed to depths of 2 m below the 1969 summer surface by the use of a CRREL 3" corer. At the 1147 m site near Glacier Camp a string of thermistors were installed at +.2 cm, 1969 surface, 1m, 2m and 5 m. The thermistors were read with a Yellow Springs telethermometer. Calibration checks in the field suggest the readings are accurate to within 0.5°C. Temperatures in the snowpack were routinely taken by the Weston dial thermometer, included in the CRREL snow kit.

Figure 6-3 shows temperatures at different periods and for the 2 m deep thermistors as well as snowpack temperatures at +.2 m above the 1969 summer surface. The gradients on the two 2 m profiles are essentially parallel and have a gradient of about 1°C/100 m elevation rise. The temperature at these depths changed by about +4°C over the course of the summer whereas snowpack was isothermal by late July although the severe weather of early/mid-August caused temperatures in the snowpack to drop well below 0°C.

AKUDNIRMUIT GLACIER 1969-1970

No detailed investigation of the mass balance of the Akudnirmuit Glacier was undertaken in 1970 although the 6 stakes established in 1969 were checked and do provide an estimate of the mass balance on the upper Akudnirmuit bowl.

1969-1970 Winter Balance:

Snow depths at the 6 stakes varied between 1.1 and 0.6 m. If the snow density is similar to that on the Boas Glacier, as is reasonable to assume, the winter net average specific accumulation is 0.256 m H_2O. This figure is nearly 0.1 m H_2O less than estimates from the upper Boas accumulation bowls (Table 6-5).

1969-1970 Mass Balance:

The stake network was visited 3 times in the course of the summer - the last visit being made on the 12th August, 1970. The surface lowering of the pack in the vicinity of the stakes varied very little with figures between .2 and .15 m. No superimposed ice was noted to have developed by July 12th and in the August visit no probing was undertaken, hence a precise statement on the mass balance is not possible. It is clear however that there was a positive mass balance and based on the results from the Boas Glacier stakes at comparable elevations the majority of the surface change is probably associated with compaction and the resulting change in snow density to near 0.38 and the growth of a few centimeters of superimposed ice.

MASS BALANCE 1970-1971

The mass balance for the 1970-1971 year was computed on the basis of visits to the glacier on June 9th and 10th and again on August 4th. Some ablation undoubtedly occurred in August but heavy snows fell on or about August 11th and this probably marked the end of the ablation season. During the June visit the stake network was remeasured and snowpits were dug. In August it was found that most of the glacier lay within the ablation or superimposed ice zone and hence they net balance could be estimated from the stakes. Figure 6-4 illustrates the changes in stratigraphy and level from June 1970 to August 1971 and formed the basis for the mass balance calculations. The snowpits proved essential to the study as the probing in many cases were clearly penetrating through the 1969-1970 pack to the top of the superimposed ice layer.

1970-1971 Winter Balance:

Three snowpits were dug near pole B2, B7 and B11 (Figure 6-1). They showed a 1970-1971 snowpack of between 0.9 and 1.04 m overlying the 1970 summer surface. Densities within the 1959-1970 snow were close to 0.45 g cm^{-3}. Densities within the 1970-1971 pack varied between 0.175 and 0.45 g cm^{-3} but averaged 0.314 g cm^{-3}, a figure virtually identical with that from the previous year (Table 6-1). The winter balance for altitudional zones ($\bar{b}_w[z]$) and the areal balance B_w are laid out as Table 6-7. The specific average net winter balance, $\bar{b}_w$, is 0.244 m H_2O. This is (0.244/0.37) x 100 = 34% lower than the winter accumulation of the previous year. Broughton Island received 0.176 m H_2O over the equivalent period or a near normal year. These data appear to suggest that there is a reasonably association between the Broughton Island precipitation record and the accumulation on the Boas Glacier. This makes precipitation analyses (Chapter 11) from Broughton Island considerably more significant, and together with the temperature correlations (discussed in Chapters 2,3,5) between Broughton Island and the Base Camp station opens the possibility of long distance monitoring of the glacier mass balance!

Figure 6-4 suggests, incidentally, that a limited amount of ablation or settling occurred after August 18th, 1970. The change varied between 0 and .3m of snow or between 0 and 0.1 m H_2O. The specific average net balance $\bar{b}_n$ should be decreased somewhat say from +0.37 to +0.32 m H_2O.

Figure 6-4 illustrates the change in surface elevation over the course of the 1971 ablation season. Stakes B2 to B5 not only lost the winter accumulation from 1970-1971 but also lost the 1969-1970 surplus mass. From stake B6 and above (that is above 960 m a.s.l.) the glacier retained part of the 1969-1970 mass but all the 1970-1971 accumulation was lost. Net balance over the glacier for the 1970-1971 balance year was clearly negative (Table 6-7) with an average specific net balance $\bar{b}_n$ = -0.2 m H_2O. The average annual exchange was, therefore, 0.66 m H_2O in decided contrast to the preceding year's figure. The average summer net specific ablation was close to 0.45 H_2O.

Ablation and Mass Balance Gradients:

The data in Table 6-7 indicate that a gradient in both ablation and mass balance (b_n) occurs between the elevation of the glacier snout (800 m a.s.l.) but that above 1050 m there is no discernible relationship. The steady state equilibrium line for the Boas Glacier, calculated on the basis of an AAR of 0.65, lies at about 975 m a.s.l. The ablation gradient is about 1.7 mm m^{-1} and the mass balance (activity index) gradient below 1050 m is close to 0.5 mm m^{-1}. Note that on the Boas Glacier the highest accumulations also occur in the area of the greatest ablation which tends to reduce the mass balance gradient.

Ice/Snow Temperatures:

Temperatures in the snowpack in early June, 1971 were close to -8°C immediately above the 1970 summer surface. This temperature was recorded at the three snowpits dug on the glacier regardless of their elevation. By contrast temperatures in the previous year had shown a marked decrease with elevation such that near the Glacier Camp temperatures at the base of the 1969-1970 pack were close to -10.5°C Figure 6-5 compares the readings in the 2 m holes between 1970 and 1971. It is evident that considerably more heat was conducted into the ice during the 1971 ablation season. The thermistor at 5 m depth recorded -9°C in both June and August whereas in 1970 the readings had fallen initially (due to penetration of the winter cold wave) from -9°C to -11°C and then fluctuated about -11°C. The major difference noted on Figure 6-6 is the great addition of heat into the higher levels of the glacier when compared with the previous year.

DISCUSSION

By a fortuitous chance our observations on the Boas Glacier occurred during the two contrasting years (Jacobs, et al., 1972). The very positive 1969-1970 balance year was followed by a year of average winter accumulation and relatively high summer ablation. However, the $<\bar{b}_n>$ for 1969-1971 was still +0.12 m H_2O suggesting that a year such as 1969-1970 has a considerable long-term effect on the glacier mass balance. The lack of firn on the Boas Glacier at an elevation of 1200 m a.s.l. meant that we could not investigate the long-term mass balance of the glacier and region by pit stratigraphy, such as Koerner (1970) successfully did for the Devon Island Ice Cap. His pit was excavated at an elevation of 1787 m a.s.l. Boyer (pers. comm.) found that even at an elevation of ca. 1500 m a.s.l. on an ice cap above the head of Maktak Fiord that the 1970-1971 snow rested on ice. Long-term mass balance fluctuations could be studied by firn stratigraphy near the top of the Penny Ice Cap at a height of 2000 m a.s.l. A 21 m combined pit and core was in fact taken during the AINA 1953 expedition at an elevation of 2050 m a.s.l. (Ward and Baird, 1954). However, these authors comment that '...below 13 m there is almost entirely solid ice...' (Ward and Baird, 1954, p.348). In

1953 the firn line was about 1500 m a.s.l. on the Penny Ice Cap whereas the equilibirum line altitude was close to 1380 m a.s.l. (Ward and Baird, 1954, p.354). This latter value is quite close to equilibrium line altitudes described on the basis of map interpretation by Andrews and Miller (1972, and see Chapter Ten). Further discussion of these points are given in Chapters Ten and Twelve. The presence of ice at a depth of 13 m near the top of the Penny Ice Cap indicates that investigation of the changes in mass balance as a function of time will have to be pursued by such methods as δO^{18} variations such as have been reported from the Greenland Ice Cap (Dansgaard, et al., 1971).

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Table 6-1

Average snow densities from pit studies, Boas

Glacier 1970 (g cm^{-3})

	Number of observations	Mean	± standard error of $\bar{x}$
All measurements	50	0.321	.0072
750-900 m	13	0.315	.0085
900-1050	11	0.337	.0159
1050-1200	12	0.327	.0107
1200-1350	5	0.284	.0164
1200-1350 east bowl	9	0.323	.028

Table 6-2

Probed snowdepths, Boas Glacier 1970,

corrected up to July 5

Altitude Interval	Number	$\bar{x}$ (m)	standard error
750-900	78	1.48	.030
900-1050	54	1.21	.044
1050-1200	50	1.30	.037
1200-1350	43	1.04	.032
1200-1350 east bowl	67	0.98	.026
all measurements	292	1.22	.0188

Table 6-3

Net specific winter balance (+), Boas Glacier 1970

Altitude Intervals	Average m H_2O	Weighted m H_2O
750-900	.466	.470
900-1050	.408	.402
1050-1200	.427	.416
1200-1350	.256	.242
1200-1350 east bowl	.343	.286
average	.379	.363
all glacier (average of Table 1 x average Table 2)	.393	.363

Table 6-4

Net winter balance B_w, Boas Glacier 1970, x 10^6 m^3

Altitude Intervals	Area (x 10^6 m^2)	B_w(z)average	B_w(z)weighted
750-900	.13	.060	.061
900-1050	.194	.079	.078
1050-1200	.405	.173	.168
1200-1350	.307	.078	.075
1200-1350 east bowl	.414	.1418	.118
Totals	1.45	.5329	.5004

Table 6-5

Net balance Boas Glacier 1970 (mid-August), m H_2O

Altitude Intervals	Average density	Average depth of snow (m)	Average superimposed ice (m H_2O)	Balance
750-900	.417	.618	.06	.307
900-1050	.488	.678	.088	.404
1050-1200	.467	.83	.160	.52
1200-1350	.385	.735	.126	.386
1200-1350 east bowl	.391	.584	.128	.330
average net specific balance $\overline{b}_n$=0.389 m H_2O				

Table 6-6

Net balance for Boas Glacier, 1970 (x 10^6 m H_2O)

Altitude Intervals	$\overline{b}_n(z)$	Area (x 10^6 m^2)	$B_n(z)$
750-900	.307	.13	.0399
900-1050	.404	.194	.0783
1050-1200	.52	.405	.2106
1200-1350	.386	.307	.1185
1200-1350 east bowl	.330	.414	.137
B_n glacier		1.450	.54230

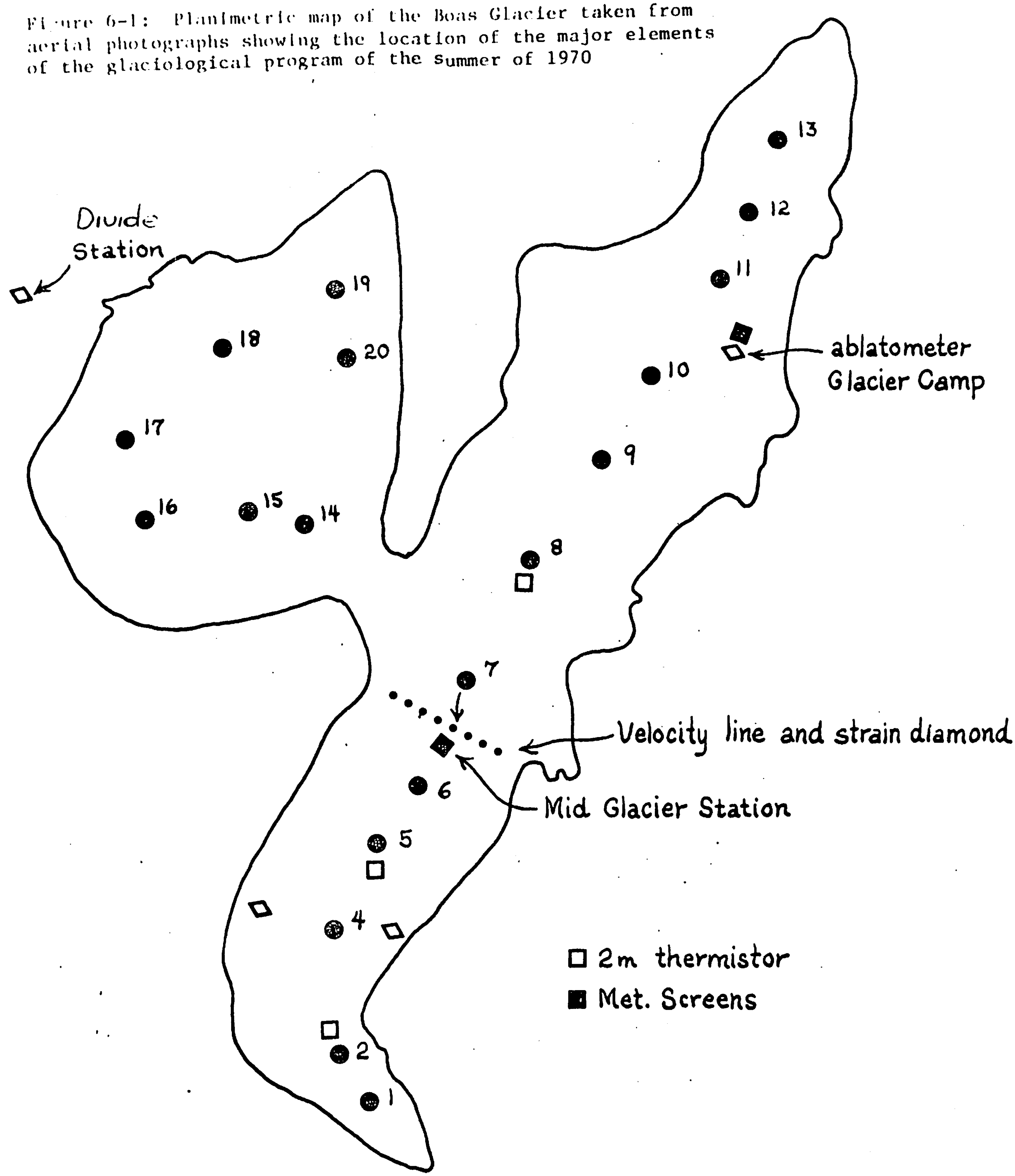
Table 6-7

Mass balance estimates, Boas Glacier, 1970-1971

(b values in m H_2O and B values in x 10^6 m^3 H_2O)

Altitude Intervals	$\bar{b}_w(z)$	$B_w(z)$	$\bar{b}_s(z)$	$B_s(z)$	$\bar{b}_n(z)$
750-900	.298	.038	-.64	-.0832	-.342
900-1050	.184	.036	-.418	-.081	-.234
1050-1200	.263	.1063	-.419	-.1694	-.156
1200-1350	.235	.0721	-.419	-.1286	-.184
1200-1350 east bowl	.235	.0972	-.419	-.1734	-.184
NB:	$\bar{b}_w$=.244		$\bar{b}_s$=-.444		$\bar{b}_n$=-.200
	B_W=.3492		B_s=-.6256		B_n=-.29

Figure 6-1: Planimetric map of the Boas Glacier taken from aerial photographs showing the location of the major elements of the glaciological program of the summer of 1970



BOAS GLACIER - 1970
(traced from air photograph)

Figure 6-2: Amount of superimposed ice as a function of elevation at the end of the 1970 summer season and the amount of surface lowering

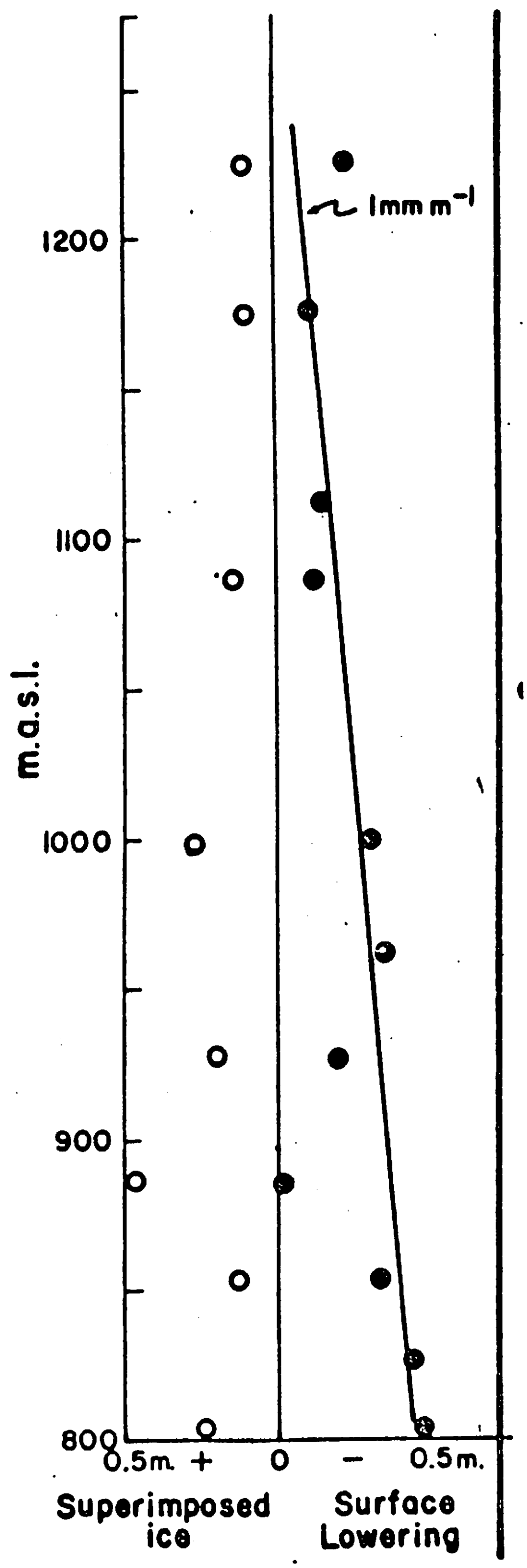


Figure 6-3: Englacial temperatures in the Boas Glacier as a function of elevation

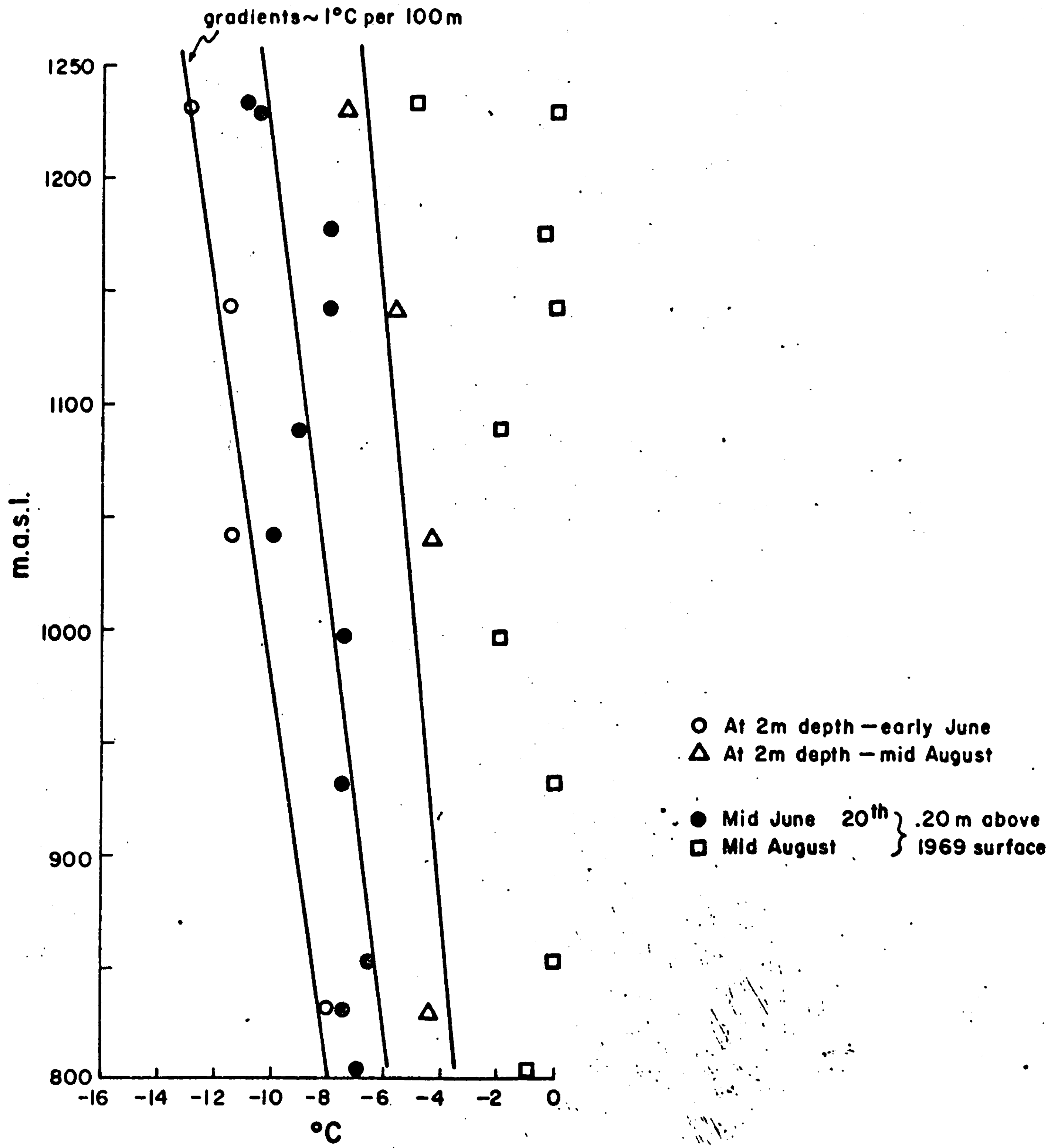


Figure 6-4: Stratigraphy in the vicinity of the main ablation stakes (Figure 6-1) in early June and August, 1971.

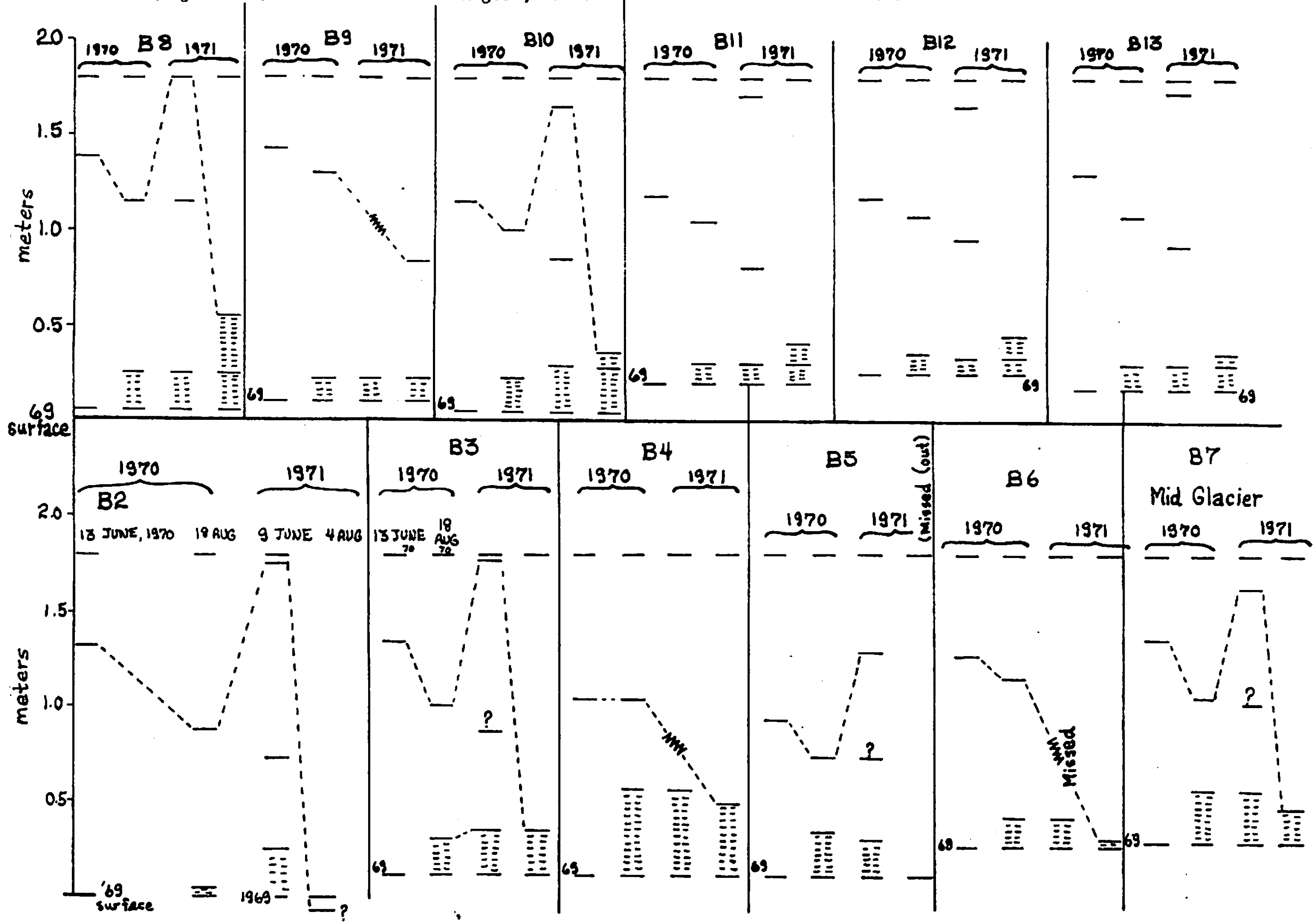
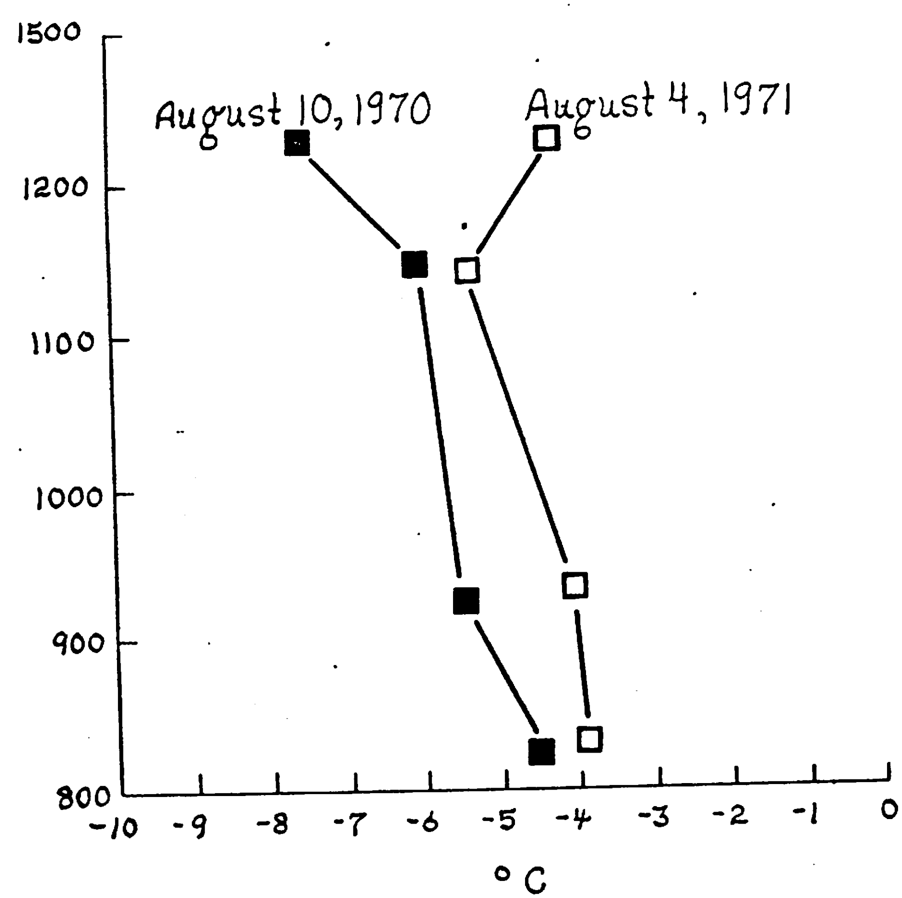


Figure 6-5: Comparison of the -2m (in 1970) temperatures up the Boas Glacier between 1970 and 1971 (June readings).



CHAPTER SEVEN

VELOCITY AND STRAIN MEASUREMENTS ON THE BOAS GLACIER (1970-1971)

R. S. Bradley and J. T. Andrews

INTRODUCTION

Paterson (1969) notes that the average surface velocity across a glacier is very close to the average velocity flowing through the cross-section at the point of measurement. Hence velocity measurements can be used to ascertain the long term mass balance characteristics of a glacier. Our approach was conditioned by a variety of logistic and other considerations. Accordingly we restricted this aspect of the research to the setting up of a velocity profile across the Boas Glacier at the inferred elevation of the steady state equilibrium line (Figure 6- 1). The actual elevation of the line is about 980 m a.s.l. and the glacier at this point is 340 m broad.

Eight stakes were set in normal to the long axis (in late June, 1970) of the glacier and these were surveyed in by resection to fixed cairns on both sides of the glacier. Resurvey was carried out on August 11, 1970, and in the August of 1971.

A strain diamond was also established in June 1970 and resurveyed in August of 1970 and 1971. The diamond was centered on stake 7 (Figure

6- 1) and had principal axes of about 58 m. Experience now indicates that the diamond could have been made larger.

VELOCITY RESULTS

The resurvey of the stake line in August 1970 indicated that problems existed in the survey procedures and no firm data could be interpreted from the short June-August 1970 record. Accordingly the August 1971 resurvey was set-up from permanent cairns on the valley side. The results of the June 1970-August 1971 surveys were then reduced on the CØGØ surveying program at the University of Colorado. The results indicate that the velocity field is quite small with a range from 1.27 to 4.05 m for the 13 month period. The average yearly velocity is estimated to be $\overline{U} = 2.04$.

In the vicinity of the velocity profile the glacier is sloping at $dy/dx = 0.25$ or about 13° with the steepest gradient being located on the west side of the glacier. The glacier also slopes west in cross-section. Figure 7-1 shows the surface velocity vectors based on our survey. The general westward component of the velocity vectors is striking and the fact that two stakes show easterly components and relatively large velocities may indicate that there may be errors in these two determinations.

We do not know the thickness of the glacier nor its bed profile in the vicinity of the stake line but a rough estimate can be derived via the relationship:

$$h = \frac{\gamma_o}{\rho g \sin aF} \qquad (7\text{-}1)$$

where a is the surface slope, ρ is the density of ice, g is the acceleration due to gravity, γ_o is the yield shear stress and F is the form ratio (Paterson, 1969). The form ratio is introduced because of the frictional effects of the valley sides on glacier flow. If the shear stress is set at 1 bar and the form ratio F = 0.65 then the thickness is estimated to be about 70 m and the cross-sectional area at the velocity line = 18,750 m^2. The accuracy of this estimate is unknown but it probably lies in the range ±30%.

As the average annual velocity is about 2 m yr^{-1} the mass of ice transfered through the steady state equilibrium line is 33,750 m^3 H_2O. Now if the glacier is in steady state equilibrium the following is true:

$$\overline{U}C = |X_c\overline{b}_n(c)| = |X_b\overline{b}_n(b)| \qquad (7\text{-}2)$$

where $\overline{U}$ is the average velocity through the cross-section C, X_c and X_b are the areas of the accumulation and ablation zones respectively and $\overline{b}_n$ is the net balance. If X_c and X_b are placed in this equation with $\overline{U}C$ = 33,750 m^3 H_2O then for X_c 324,000 m^2 and X_b 628,000 m^2 the average net balances are $\overline{b}_n(c)$ +0.05 m H_2O and $b_n(b)$ -0.1 m H_2O respectively. These values indicate a very slow mass turnover and, moreover, indicate that the Boas Glacier will be very sensitive to changes in the amounts of winter accumulation and summer melt.

Given that the measurement of velocity is reasonably correct it is possible to use the estimates of surface areas within specific elevation ranges to construct a tentative model for the steady state equilibrium condition of the Boas Glacier. There is one immediate problem, however, and that is how dynamically linked into the glacier is the contribution of the East Bowl of the glacier. This area above 1200 m a.s.l. comprises 0.414 km^2 and hence the decision is a major one. Field observations during 1970 and 1971 suggest that the channel connecting the East Bowl to the main Boas Glacier is very shallow and restricted and hence the area might really consist of a thin, semi-stagnant ice patch with little importance to the mass state of the main glacier. However, we will consider two cases - one with and one without the East Bowl. Considerations of the premise of steady state equilibrium indicate (Eq. 7-2) that we have three knowns and three unknowns. In order to derive preliminary results we have of course to estimate the cross-sectional area of the glacier in the vicinity of the ELA (Eq. 7-1).

Model 1: Boas Glacier and East Bowl

Figure 7-2 illustrates the computed mass balance diagram. To produce a net balance over the ablation zone of on average -0.1 m H_2O within the realm of our present knowledge suggests: 1) the need to have an accumulation gradient of the order of 0.4 mm m^{-1} and an ablation gradient of 1 mm m^{-1}. The latter compares with the value of 1.6 mm m^{-1} for the south-facing Penny Ice Cap traverse (Ward and Baird, 1954) and thus is reasonable for the north-facing Boas Glacier. The net balance in the accumulation area has to have a positive mass gain of +0.07 m H_2O at the top of the glacier and a net balance gradient (Figure 7-2) of ca. 0.28 mm m^{-1}.

Model 2: Boas Glacier without the East Bowl

The ablation area and the requirement for a net mass loss of on average -0.1 m H_2O remains the same. However, the accumulation area is now reduced by 0.414 km^2. If a steady state AAR of 0.6 to 0.7 pertains to this glacier then this latitude is sufficient to include the ELA in the range 950 to 1050 m a.s.l. To obtain a specific net balance $\overline{b}_n(c)$ = +0.05 m H O the net balance gradient has to be increased slightly to about 0.4 mm m^{-1}.

Discussion:

Both models fit reasonably well with our limited knowledge of the present glaciological regime of the Boas Glacier (Chapter 6). It is interesting that the inclusion or not of the East Bowl makes only a limited change in the net balance gradient in the accumulation zone, assuming that the ELA lies in the vicinity of 1050 m a.s.l. From the point of view of climatic change and paleoglaciology (Chapter 11) the decrease of accumulation with elevation (Chapter 6) indicates that changes in snow accumulation within the area of the accumulation zone would be most significant for changing the mass balance of the Boas Glacier. Lower than normal snowfalls would easily result in a translation of the net balance lines of Figure 7-2 to the left and the glacier would lie wholly within the ablation zone (Figure 7-2B) if snowfall led to an accumulation of 0.1 m H_2O whereas an increase in accumulation of 0.1 m H_2O (Figure 7-2B) would lower the ELA for the year to 925 m a.s.l.

The Boas Glacier would also be very sensitive to changes in the pattern of snow accumulation. For example, if the present trend of a decrease with elevation were reversed the effect would be to increase the activity of the glacier because this would require greater mass gains and losses for a glacier of the present size. Discussion on related matters will be taken up in Chapter 11.

STRAIN RATE MEASUREMENT 1970-1971

The strain network was measured in June and August 1970 and again in August 1971. The results were not as internally consistent as were hoped for but the net will be measured again in 1972 as an additional check. The measured strains were computed after the method outlined by Nye (1959).

At the present time we believe that the measurement in June 1970 and August 1971 are most reliable and these are used. The main problem is that the strains are so low that measurement error becomes very important. Hooke (1970, p.311) states, for example, in his study in Greenland that an accuracy of $\pm 0.005 \text{ yr}^{-1}$ is suggested; this is the same order as our computed strain rates!

We adopt the normal convention for the co-ordinate axes on a glacier, viz. the x axis is directed along the main axis of the glacier, the z axis is normal to the x in the horizontal plane and the y axis is vertical. The direction of the principal strain is directed toward the northwest in agreement with the movement of the velocity stakes (Figure 7-1). The strain rates are such that glacier is slightly compressive at the point of measurement as $\dot{\Sigma}_x$ is negative and $\dot{\Sigma}_x$ (vertical) is positive. If the glacier is 70 m thick the upward movement at the site is about 0.01 m yr^{-1}. This very low rate of upward movement is in substantial agreement with the mass balance estimates on Figure 7-2 and indicate that indeed the ELA is located close to, but upglacier from the stake line.

CALCULATED VELOCITY FROM THEORY

The velocity of a glacier can be estimated if the slope, depth and englacial temperature profile is known. Conversely, if the slope, depth and surface velocity are known it is possible to say something about the englacial temperature profile. The equation of interest is:

$$U_s - U = 2\underline{A}\ (\rho g)^n\ (\sin^n)\ (h-y)^{n+1}/(n+1)$$

where U_s is the surface velocity and U is the velocity at some depth y, h is glacier thickness, γ is the glacier surface slope, ρ is the density of ice and g the acceleration due to gravity, and n is a power that experimentally varies but a value of 4.2 is frequently used (Paterson, 1969) and $\underline{A}$ is a constant that is temperature dependent. According to Paterson (1969, p.83), the value for $\underline{A}$ differs by an order of magnitude over a temperature range from 0°C to -22°C. The value for $\underline{A}$ at 0° is 0.148 yr^{-1} $bar^{-4.2}$ (Paterson, 1969, p.88). In a small glacier such as the Boas Glacier the shear stress calculation has to be modified by the shape factor, F, to account for frictional effects of the valley sides. We have used a F value of 0.65. Table 7-2 lists the calculated velocity for the glacier with englacial temperatures of 0°C and -13°C. As the longitudinal strain rate is small this contribution can largely be ignored.

CONCLUSIONS

The Boas Glacier is a slow-moving subpolar glacier with a low rate of mass exchange. The temperatures at the -10 m level are probably close to -12°C and with the suggested depth of the glacier and a temperature gradient of 1C/20 m the basal temperatures must be below the pressure melting point. Measured velocities near the supposed ELA are close to 2 m yr^{-1} on average and strain rates are of the order of 10^{-3} for the x and z axes and 10^{-4} for the y axis. Greater accuracy is required and this will be forthcoming by repeating the measurements over a period of one or more years.

The results suggest that, and as expected, the Boas Glacier is not isothermal at its base nor is it as cold as -13°C at a depth of 70 m. We have not attempted to model the velocity any further although a linear temperature increase below -10 m of about 1°C/20m would appear to give a reasonable surface velocity.

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Table 7-1

Strain rates of the principal axes yr^{-1} for the period June 1970 to August 1971, direction of maximum strain and shear strain rates.

$\dot{\Sigma}_x = -0.00575$ $\dot{\Sigma}_x = +0.00552$ $\dot{\Sigma}_x = +0.00022$

Angle of principal strain rate from x axis is =12.7°

Shear strain rate $\dot{\Sigma}_x = 0.00128\ yr^{-1}$

Table 7-2

Calculated velocities (m yr^{-1}) on the Boas Glacier in the vicinity of the velocity profile for different temperatures (0 = 4.2)

	Temperature	
Depth (m)	0°C	-13°C
0	5.3	1.35
10	-	-
20	-	-
30	-	-
40	5.0	1.2
50	4.4	1.1
60	3.0	0.74
70	0	0

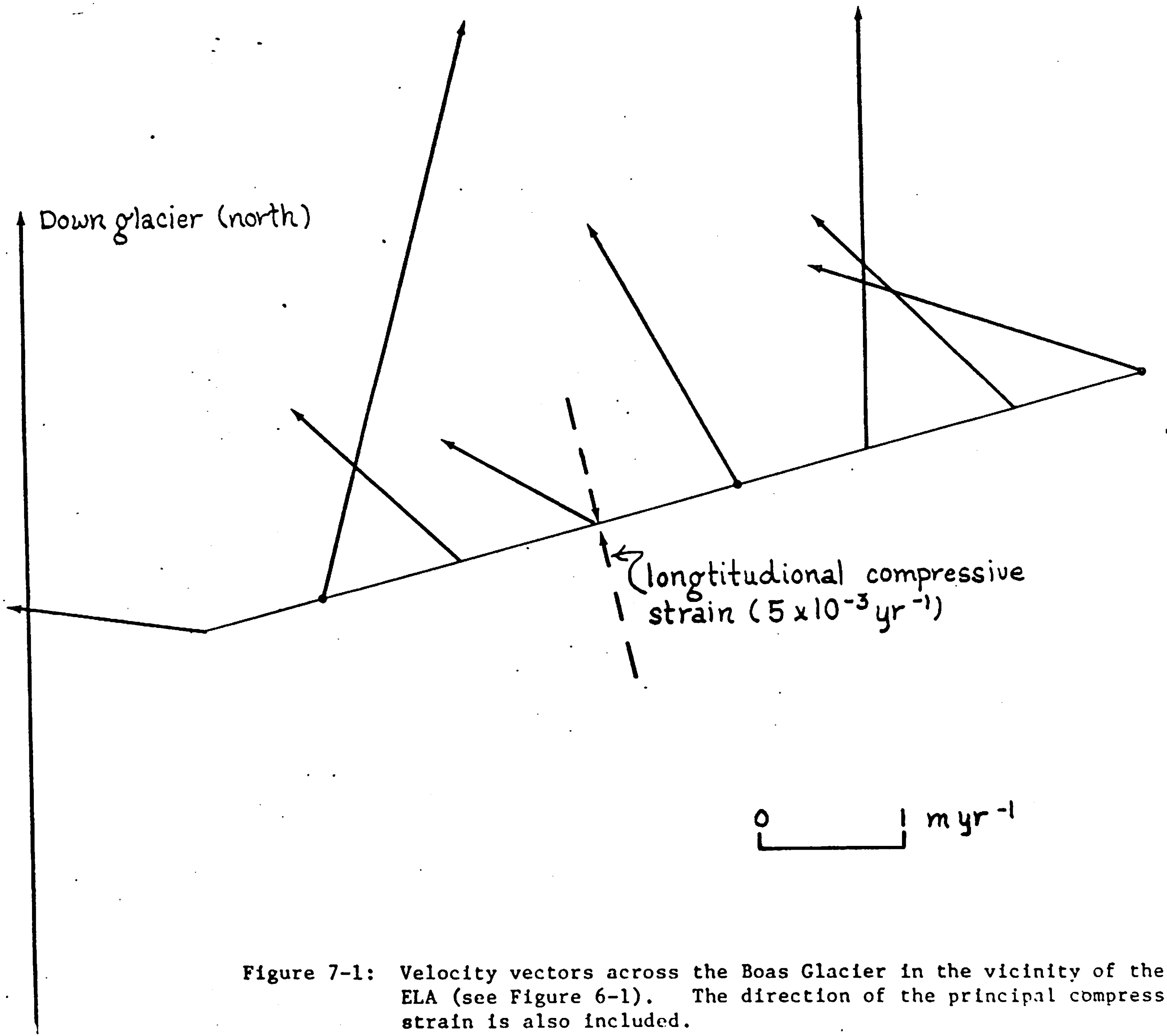


Figure 7-1: Velocity vectors across the Boas Glacier in the vicinity of the ELA (see Figure 6-1). The direction of the principal compressive strain is also included.

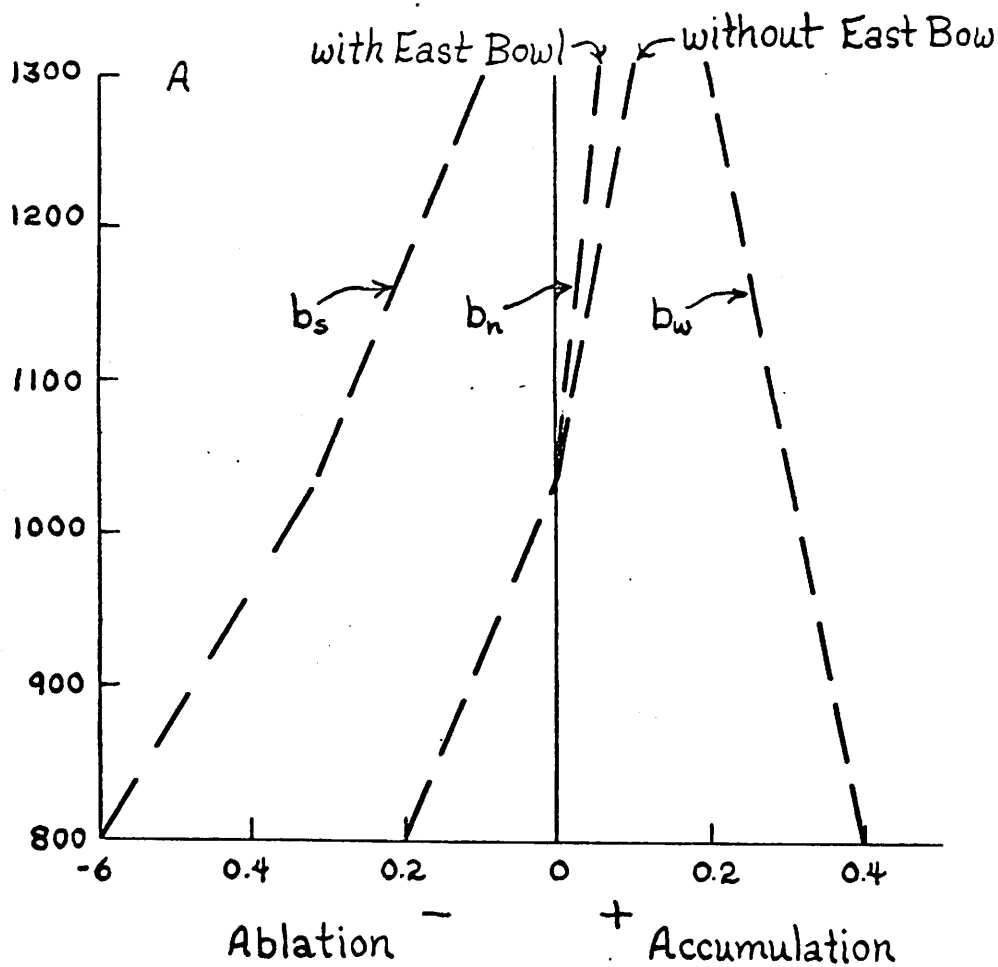


Figure 7-2A: Tentative steady state mass balance diagram for the Boas Glacier based on velocity profile in the vicinity of the ELA

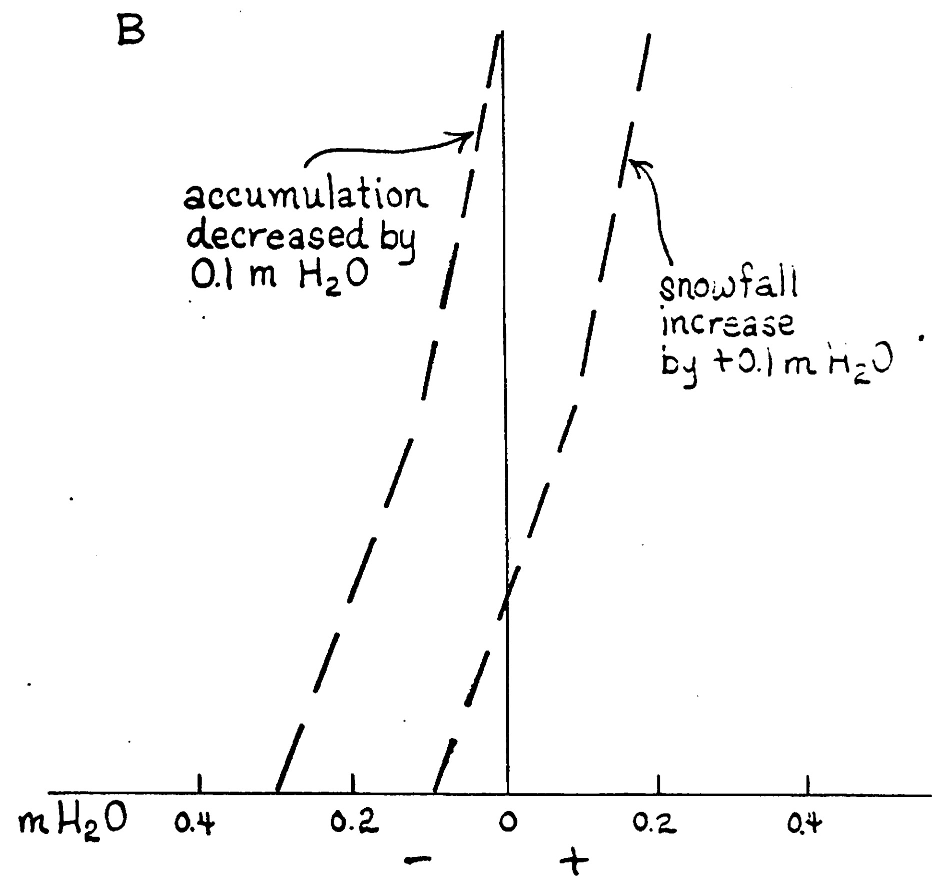


Figure 7-2B: Effect of increasing and decreasing snowfall on the Boas Glacier showing how sensitive the glacier is to snowfall amounts.

8. RUNOFF FROM BOAS GLACIER DURING THE 1970 ABLATION SEASON

L.D. Williams

Measurement of meltwater runoff from the Boas Glacier during the summer of 1970 was handicapped by late-lying snow in the stream channel. Running water in this stream was first observed on July 13, when it emerged from a snowbank to descend a steep falls and disappear again beneath a snowbank. Streamflow during the next two weeks appeared to be heavy, but it was not until July 26 that the channel was exposed to allow measurement of water level and discharge. At that time, a Stevens Type F stage recorder was installed on a bedrock platform in the channel about 300 m from the glacier, and water level was recorded continuously, except for times of icing of the stilling well, from July 26 to August 19. An identical stage recorder was intended for a second stream channel from Boas Glacier; however, that channel was found to have negligible flow at present, and so the recorder was set up on July 29 below the junction of the main streams from Boas and Akudlermuit glaciers. Correlation between the two records is excellent ($r = 0.93$ at peaks and lows), a circumstance which was used to improve the rating curve for the Boas stream, as well as to estimate Boas stream flow at times when the float of the recorder there was jammed by ice in the stilling well.

To obtain rating curves of discharge versus water level for the two recorders, eight measurements of discharge were made, four at each site, by means of a dye dilution method (Adams, 1962) whereby discharge is equal to dye dilution (determined by colorimeter) times rate of injection of the dye. The small number of discharge measurements (imposed by failure of the colorimeter power supply necessitating return of the samples to the laboratory for analysis) was felt to be inadequate to define a rating curve for the Boas stream. Measurements at the lower recorder, although also few in number, were well-spaced and quite linear, and so in view of the high correlation between flows at the two sites, the rating curve for the lower recorder was used to obtain estimates of peak and low flows for the upper recorder. This was accomplished as follows: the rating curve for the lower recorder was established as

$$Q_1 = 3.662\ d_1 - 13.094 \qquad (8\text{-}1)$$

(98.5% variance explained, significant at 99% level) where Q_1 is discharge and d_1 is water level. The discharge measurements at the upper recorder were then fit to predicted values of discharge at the lower recorder:

$$\log_{10} Q_u = 0.2428\ Q_1 - 0.3863$$

(99.6% variance explained, significant at 95% level). With these equations values of discharge at the upper recorder were predicted for all peaks and lows. The 19 points thus obtained, including the actual measurements, were found to fit very well the regression equation

$$\log_{10} Q_1 = 0.08757\ d_1 - 0.8175 \qquad (8\text{-}2)$$

where Q is discharge and d is water level at the upper recorder, explaining 87.4% of the variance and significant above the 99.9% level. With this rating curve, estimates of daily runoff from Boas Glacier during the period of measurement were computed, shown on Figure 8-1 Runoff was computed by summation over two hour intervals from noon to noon, as maximum discharge ordinarily occurred around midnight. The abscissa of Figure 8-1 gives the day ending each 24-hour summation.

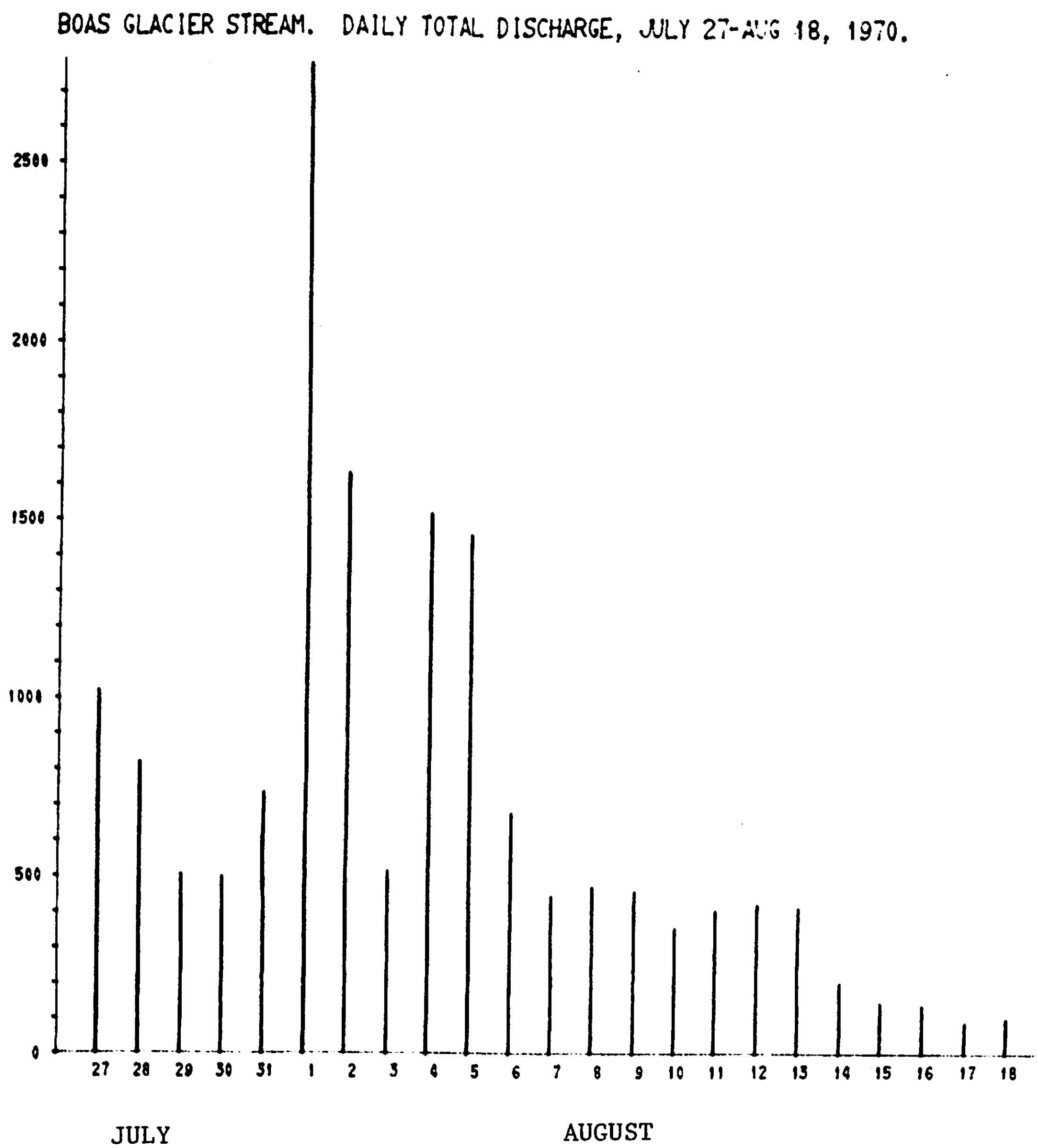
Total runoff from the Boas Glacier during the period of measurement July 26 -

August 19 was only 16,500 m^3. That the major part of the runoff occurred beneath the snow prior to installation of the stage recorder is evident from ablation measurements on the glacier. Net lowering was estimated as 0.06±0.03 m H_2O over the glacier surface area of 1.45 x $10^6 m^2$. If this were all by melting, and neglecting superimposed ice, possible runoff would be in the range 43,500 m^3 to 130,500 m^3. It is possible to demonstrate this in another way. During the period July 13-July 25, between the time water flow was first observed in the Boas stream and the time measurement was begun, discharge in the vicinity of the eventual site of the lower recorder had been estimated at 5 p.m. on all but two days, by means of a formula relating discharge to width, depth, and surface velocity of the stream. This formula is $Q = 4wd/t$ in liters/sec, where w = width in m, d = depth in cm, and t = time in seconds for a float to travel 150 cm in the center of the stream. When discharge measurements were made later, estimates by the formula were found to be in good agreement, and furthermore the 5 p.m. estimates for July 29-August 18 were well correlated with daily runoff from the Boas Glacier ($r = 0.91$). Thus, the 5 p.m. discharge estimates for the days July 13-July 25 can be used to approximate the daily amounts of Boas runoff during that period. Using a linear regression equation, the total runoff for July 13-July 25, excepting July 22-23, is estimated at 51,000 m^3, so the total runoff for July 13-August 19 would have been in excess of 67,500 m^3.

REFERENCES

Adams, W.P. 1962: A study of some methods of measuring ablation and run-off on an arctic glacier. M.Sc. Thesis, McGill University, 84 pp.

Figure 8-1: Noon to noon total daily discharge from the Boas Glacier computed from log regression on a linear model (see text).



9. QUATERNARY GEOLOGY AND GLACIAL CHRONOLOGY

J. T. Andrews, S. Boyer, P. Carrara, R. E. Dugdale and G. H. Miller

INTRODUCTION

The research reported in this chapter has been supported by the ARO under the terms of the grant to Andrews and Barry but considerable information was also obtained through research supported by the National Science Foundation grants GA-10992 (to Andrews) and GA-20883 (to Andrews and Harrison) as part of a study on recent crustal movements along the eastern Baffin Island coast. Published papers and thesis emanating from INSTAAR on the topic of this chapter include, under ARO support: Carrara (1972), Carrara and Andrews (1972), Dugdale (1972), and Miller and Andrews (1972), and under NSF support: Pheasant (1971), Boyer (1972), Mears (1972), Pheasant and Andrews (1972), and England and Andrews (subm.). Previous work in the region, as noted in Chapter One, was very limited.

In terms of the initial concept of this grant we are particularly interested in the glacial chronology of Narpaing and Quajon fiords and the cross-valley linking Narpaing to Maktak Fiord. Table 9-1 is a schematic presentation of the overall chronology of northern Cumberland Peninsula based on existing evidence. We will briefly review these data and then discuss in more detail the local glacier chronologies for the Boas, Akudnirnmuit,Sulung and Itidlirn valleys and of the Maktak-Narpaing through trough.

ZONE I

Zone I is the oldest surficial unit in the area. It is identified by, and separated from the younger Zones II and III by various stratigraphic criteria, namely composition (local or erratic), weathering and size. The concept of weathering "zones" embodies something of the concepts of both rock-stratigraphic and soil stratigraphic units. In a vertical sense Zone I is the highest recognizable unit. There is no clear unequivocal evidence that Zone I was ever glaciated by actively moving ice moving toward Baffin Bay from interior Laurentide source areas, such as Foxe Basin. This statement appears slightly paradoxical because several thin summit ice caps lie over extremely weathered bedrock that shows no evidence of active glaciation. A typical example is in fact the summit between the Boas and Sulung valleys. The extreme weathering is primarily pinpointed by:. 1) large weathering pits on boulder surfaces; 2) the projection from boulder surfaces of feldspar and quartz crystals; and 3) the extreme rounding of the edges of boulders and the formation of grussified boulders. This zone is considered to represent weathering throughout the Quaternary and probably in Pre-Quaternary.

ZONE II

The contact between Zones I and II slopes seaward (Figure 9-1) along the axes of the fiords (Pheasant, 1971; Boyer, 1972). The boundary is mapped largely on the basis of the highest exposure of till. The till is, however, weathered to a considerable degree, especially when compared with the fresh till of Zone III. Boulders in Zone II are not as badly weathered as those in Zone I - inclusions are present to cause considerable micro-relief on boulder surfaces but large weathering pits are absent. A considerable percentage of surface erratics show oxidized weathering rinds when broken with a hammer. Low weathered moraines and glacial drainage channels exist within Zone II. Discriminant analysis on weathering variables (Boyer, 1972) indicates that Zone II is distinguishable from Zone I but is closer to this litho-stratigraphic unit in weathering criteria (= age) than it is to Zone III. The actual age difference between Zone I and II depends primarily on how we believe weathering rates change as a function of age. If the relationship is some power function, as is most likely, then Zones I and II might be similar in their weathering characteristics but be widely separated in time.

ZONE III

The Zone II/III boundary is frequently marked by a major lateral moraine. Within Zone III, the till and bedrock are fresh and although morpho-stratigraphic units are recognized within Zone III (Table 9-1) these cannot be differentiated on the basis of our existing weathering parameters (Boyer, 1972). They are, however, easily delimited on the basis of the position of morphostratagraphic units. In certain areas these units descend to near sea level and become part of lateral moraine/glaciomarine morpho-stratigraphic units. Fossil marine shells in the raised marine deposits have been dated by ^{230}Th and ^{14}C and provide the preliminary calibration for Zone III statials and interstades (Table 9-1) (Pheasant and Andrews, 1972). The ^{230}Th date of 137,000 ± 10,000 comes from marine shells immediately underlying a till. The till is, in turn, associated with a lateral moraine that forms the boundary between Zone II/III on the outer coast between Narpaing and Quajon fiords. The shells come from about 19 m a.s.l. and indicate that glacio-isostatic depression was already affecting the crust during the early stages of glaciation.

TABLE 9-1

Quaternary glacial chronology, Okoa Bay, Baffin Island, N. W. T.

Age		
0		
	Neoglaciation (several phases)	
5,000	"Warm" interstade	
6,000		
8,000	Cockburn readvance - maximum of "late Wisconsin glaciation" in the area	
10,000	"Cold" interstade	
20,000		
24,000		
32,000±	advance advance Broughton Island	
40,000±		ZONE III
68,000 ± 5,000		
68,000 ± 5,000	'Warm' interstade	
	earliest last glaciation - possibly second stades	
137,000 ± 10,000		
500,000±		
Early and Mid Quaternary	Zone II - possibly several glaciations	
Pre-Quaternary	Zone I	

Figure 9-1 shows the relationship between the three weathering zones and the elevation of corries. As noted by Williams (1972) the coincidence between corrie floor elevation and Zones II and III is probably not coincidental. Pheasant (1971), Boyer (1972) and Mears (1972) have all shown that the gradients of the Zone I/II and Zone II/III boundaries are compatable with glacier physics in that the boundaries can be used to model former fiord glaciers with basal shear stresses of about 1.0 bar.

Figure 9-2 is a schematic time/distance diagram from the fiord heads to the outer coast. During Zone III there has been a progressive decrease in the amount of ice in the area. The ice was most extensive during the earliest stade and is now at a minimum - although it is important to note that the amount of present day ice is not greatly different from that during the Cockburn readvance of about 8,000 BP. This is easily seen on Figure 9-2.

GLACIAL CHRONOLOGY OF THE BOAS/SULUNG AREA

Let us recapitulate the significant aspects of the present and past glacier extent in the Boas/Sulung area (see Chapter One also). The catalyst for this study was the observation on aerial photographs that empty, south-facing corries had massive moraines some 6 km from their headwalls whereas tentatively identified correlative moraines in front of north-facing corries lie very close both present glaciers and younger ice-cored moraines. The present distribution of corrie glaciers is notably asymmetric (Andrews et al., 1970; Andrews and Dugdale, 1972; and Williams, 1972) with virtually all features located in the northern quadrats (see Chapter Ten). One of the primary aims of our research was to determine the relative, and preferably absolute ages of the moraine systems, and to derive estimates of former equilibrium line altitudes.

Our dating of the moraines in the research area has been based on:

1) ^{14}C dates and related glacial chronology of the major fiords (Maktak, Narpaing and Quajon);

2) general position of a moraine within a moraine complex;

3) lichenometry - used with success for features up to about 8,000 years old;

4) weathering studies on moraines and statistical analysis of the data.

The following chronological framework is suggested:

1) During the earliest stades of the last glaciation (134,000 and 768,000 BP) the Maktak/Narpaing trough was filled with ice flowing into it from both the present area of the Penny Ice Cap and from the high broad mountains east of the trough. Large outlet glaciers flowed to the outer coast along the main fiords. In the vicinity of the trough the ice limit is marked by lateral moraines at elevations of 850 m or so. This indicates that the equilibrium line on these glaciers lay above 850 m a.s.l. Williams' (1972) analysis of the relationship between corrie floor elevations and the weathering zones suggest that during this stade the corries fed ice into the main outlet glaciers. Ice from the through trough overflowed into Quajon Fiord through a col at 600 m a.s.l.

2) Marine fossils in Quajon Fiord record a warm interstade ^{230}Th dated at 68,000 ± 5,000 BP. Ice advanced sometime after that date (Table 9-1) but in Quajon Fiord was considerably less extensive than during the earliest stades and was wholly from local sources. The same statement holds true

in the through trough and over much of the area (Figure 9-2). Figure 9-3 is a map showing the extent of ice in our field area during 'mid' Wisconsin time (from: Boyer, 1972). The correlation between moraine units was achieved by field mapping and statistical grouping of weathering criteria using multiple discriminant analysis. In the ice free area between the two main piedmont glaciers (Figure 9-3) a lake was dammed up some 85 m above the present lake level. It is clear from this map that the present day asymmetrical glacier distribution did not hold so dramatically during this stade although a difference in extent still existed. During the middle stade(s) ice from north facing glaciers flowed into Quajon Fiord and terminated in the fiord. Similarly, ice flowed to Broughton Island via Kingnelling Fiord from the main mountain ice cap on the east side of the through trough whereas glaciers that flowed south into the trough did not have sufficient momentum and size to join the main Maktak outlet glacier. Judging from the shape and area/elevation relationships the ELA's in the through trough were of the order of 750-850 m a.s.l.

No absolute ages are available from the area of the through trough, although on Broughton Island, England and Andrews (subm.) have ^{14}C dates on marine shells which indicate a readvance between 24,000 and 32,000 years ago. Further north in Baffin Island a date of 39,500 BP (GSC-209) is reported on willow stems from the crest of an 'old' corrie moraine similar to those discussed here.

Dugdale (1972) and Carrara and Andrews (1972) tried to gain an estimate of the age of the Sulung moraines by dating the Boas and Akudnirmuit moraines on the basis of lichenometry and other evidence and then using these dates to establish a curve of weathering against age. Their evidence indicated that the outer Sulung moraine (Dugdale's Phase 1) was ca. 35,000 years old and that the lateral moraine of a glacier that had once lain in the through trough ca. 41,000 BP. These estimates are based on the premise that the Phase 3 (youngest) Sulung moraine was 12,500 years old. The middle moraine unit (Phase 2) has an intermediate age of 23,000 BP.

Boyer's analysis of the data does not support a three-fold age separation - at least, although most glaciers in the through valley have three old end moraines, Boyer's data indicates that the outer one is distinct in age from the two younger ones that appear to have similar weathering criteria. Boyer (1972) also takes the stance that Dugdale's Narpaing Phase (the main lateral moraine tracing ice in the through trough) and the Phase 1 moraine in Sulung valley are of comparable age.

Table 9-2 (from Boyer, 1972, p. 62) lists the means for his weathering age criteria for a two-fold division of the moraines into 'mid' and 'late' glacial.

Table 9-2

Means of age dependent parameters from 19 stations in the through trough (from Boyer, 1972, p. 62)

Variable	Means: Late	Means: Mid	Entry Disc. Analysis
Edge rounding (Cailleux)	1021	1099	3
Height (m)	57	44	
Distal slope	18	7.6	1
Crest width (m)	79	152	2
Boulder surface count	62	85	

The main difference between the two age units are in the angle of the distal slope and width of the moraines and to a lesser extent the height of the moraines. Slopes on Neoglacial ice-cored moraines are nearly 30° and the decrease in slope and increase in width with age of the moraines in the cross-valley indicate the reduction of moraine relief through time. The ratio of late:mid on these two criteria are close to 1:2. We will return to the question of the age of these moraines in the next section.

3) The work of INSTAAR in northern Cumberland Peninsula has led to the realization that the late glacial maxima in the area occurred not 24,000 or 15,000 years ago but sometime between 7,800 and 8,500 BP during the Cockburn Stade (Table 9-1). This late glacial expansion refers not only to the major outlet glaciers but also to the local mountain ice caps and corrie glaciers (Pheasant, 1971; Mears, 1972; Miller, 1972; Carrara and Andrews, 1972). The general evidence for this late glacial expansion and a discussion of the lack of evidence of extensive glaciation between 10,000 and 20,000 BP is given by Andrews and Ives (1972) and Andrews, Mears, Miller and Pheasant (in press).

On the basis of lichenometry, weathering and relationship of the expanded Boas Glacier to a gully and alluvial fan at the head of Narpaing Fiord, Carrara and Andrews (1972) date the outermost moraines fronting the Akudnirmuit and Boas glaciers at 7,000 BP. This provides a date on substrate stablization and hence is not necessarily at variance with a maximum advance at about 8,000 BP. A problem arises, however, when it comes to dating the innermost Sulung moraine and the inner moraine in the adjoining Itidlirn Valley (Dugdale, 1972). On the basis of a comparison of the 'appearance' and edge weathering of pebbles Carrara and Andrews (1972) infer that the innermost Sulung moraine is older than the outer Boas moraine by approximately 4,000 - 5,000 years. Lichens under the headwall of the Sulung corrie reached 150 mm indicating that the area had been ice free for at least 6,000 years. During the formation of the Boas outer moraine steady state ELA was about 800 m a.s.l. compared with 850 m a.s.l. in the Sulung valley (see Chapter 10).

Many of the late glacial corrie moraines in the region are double crested and the outer one appears somewhat more weathered and subdued than the inner feature (G. H. Miller, pers. comm., 1972). No clear example of this moraine has been found fronting the Boas and Akudnirmuit glaciers although in the through trough most of the late glacial moraines have two distinct units (Boyer, 1972). It is suggested that the outermost one might date from early in the beginning phases of the last glacial stade (late Wisconsin in the south) and that the inner one dates from the late glacial expansion correlative with the Cockburn Stade (Figure 9-4). In this scheme the tentative date of Dugdale's (1972) of 23,000 BP might be reasonable, or at least a date within the range of 15,000 to 23,000 BP. It is reasonable to ask at this juncture why have no ^{14}C dates been obtained on marine deposits related to this early late glacial (Figure 9-4)? This question has also been raised by Andrews and Ives (1972) and Andrews _et al_.(subm.). The answer seems to be that relative sea level during this period was _below_ present sea level.

Table 9-2 suggests that our best estimate for the age of the outermost (mid glacial) moraines is that they are approximately twice as old as the inner moraine loopes. This estimate is based simply on a linear extrapolation of the reduction in slope angle and the increase in crest width.

Rampton (1970) has used the reduction in slope as an age discriminant in his study of moraines in the Yukon, Canada. If the overall best estimate for the inner moraines is 8,000 ± BP then our best estimate for the outer mid glacial event is of the order of 40,000 BP.

4) In the period between about 6,500 and 4,500 BP the local glaciers and ice caps retreated to positions less extensive than they are today. In this same period the moraines formed about 8,000 BP lost their ice cores and became subdued, low features. Sometime prior to 3,200 BP a climatic deterioration led to the regeneration of some glaciers and the expansion of others during the period now known as the Neoglaciation (Table 9-1).

The glacial history during the Neoglaciation has been studied in particular by Miller (1972) and Boyer (1972) and in detail in the specific research area by Carrara and Andrews (1972). Miller's findings (1972) are shown as Figure 9-5. The main point of this figure is to demonstrate that the great majority of glaciers in the area and particularly the outlet glaciers from the Penny Ice Cap reached their maximum Neoglacial extent within the last 70 years although there is unequivocal evidence for a number of earlier advances (Figure 9-5). During the Neoglaciation the present asymmetric distribution of glaciers developed. A discussion of present and past Glaciation Limits and snowlines is deferred until Chapter Ten. Figure 9-5 could be taken to read that this section of the Arctic has been undergoing periodic climatic deteriorations during the Neoglacial but that these are superimposed on an overall non periodic (in the time scale of 4,000 years) deterioration of climate that is leading the area toward full glacial conditions. Andrews, Barry, Bradley, Miller and Williams (1972) have indeed argued that eastern Baffin Island will provide the lead into the next full glaciation of the northern hemisphere. The current climatic deterioration is considered in Chapter 11.

DISCUSSION

Two questions of different scale require answers: first, were the south-facing corrie glaciers coexistent in general with the north-facing counterparts and what was the relative size of glaciers with these two aspects; and second, what is the relative chronology and size of the Boas/Akudnirmuit glaciers and the Sulung/Itidlirn glaciers and how representative are they when compared to the larger sample?

The answer to the first part can be partly seen in Figure 9-3 and 9-5 where it is apparent that during the mid and late glacial times considerably greater areas and volumes of ice were maintained on slopes and mountains with northerly aspects compared with southern slopes. From our data it seems that ELA's were approximately of similar elevation for north- and south-facing orientations - a situation radically different from that which prevails under current climatological conditions. It is also clear from Figure 9-3 and 9-5 and from the occurrence of similar moraine units on either side of the through trough that the south- and north-facing glaciers co-existed during the same glacial stades.

In detail, the comparison and contrast between the Sulung/Boas glaciers is not typical. This is so on several counts: 1) the Sulung and Itidlirn glaciers were larger than the two north-facing glaciers; 2) only one moraine fronts the present termini of the Boas and Akudnirmuit glaciers compared with three distinct moraines fronting other glaciers in the through trough; 3) the late glacial moraine, date ca. 7,000 BP by Carrara and

Andrews (1972) lies immediately in front of the oldest Neoglacial moraine whereas in most other cases the late and mid glacial moraines lie some kilometers in front of the Neoglacial moraines. The reason for the difference may be in part related to a difference in general topographic setting for in most examples in the through trough the north slope is longer and gentler than the south (southwest) slope but the reverse is true in the Sulung/Boas area.

Despite these findings, the paleoclimatic problem of maintaining glaciers in the open south-facing valleys under current climatic conditions of higher global radiation and less solid precipitation is still apparent. The findings of this chapter are extended in the next chapter using different methods of analysis.

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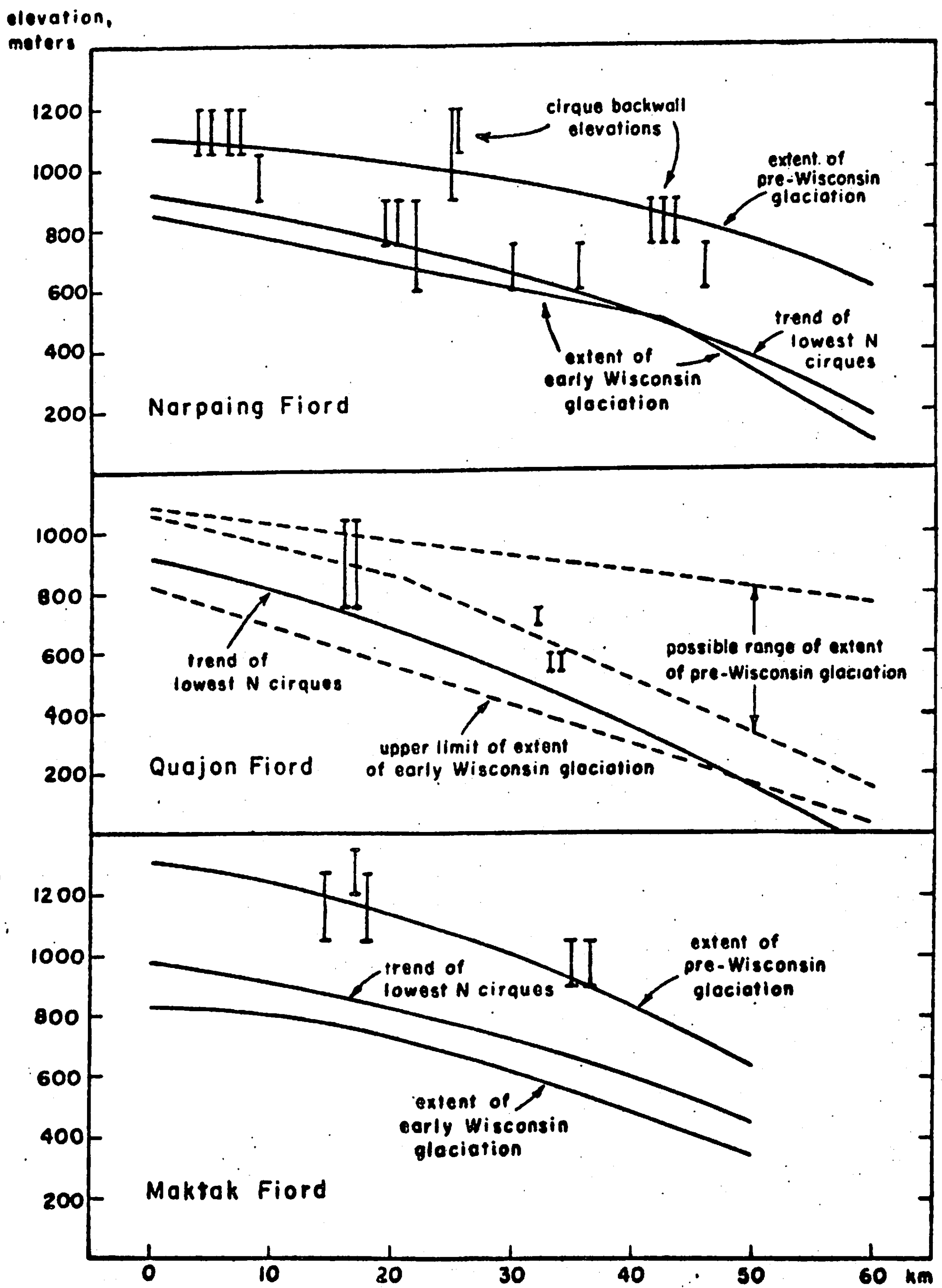
Figure 9-1: The relationship of the major weathering zones and corrie floor elevations. Note the weathering zones slope seaward (i.e. to the right of the diagram).



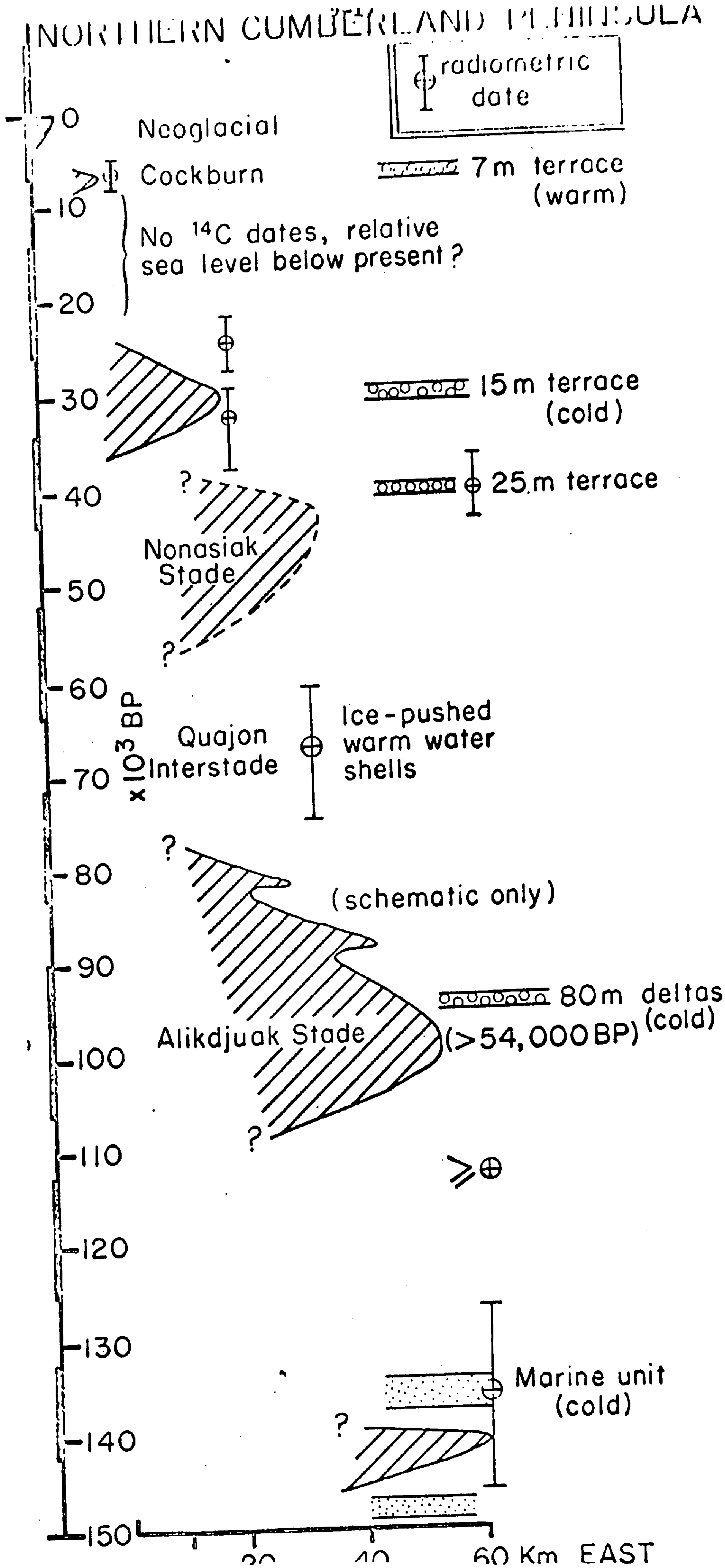


Figure 9-2: Time/distance diagram showing the relationship between ice margin position during the last glaciation and distance.

Figure 9-3: Extent of Middle Wisconsin ice in the cross valley

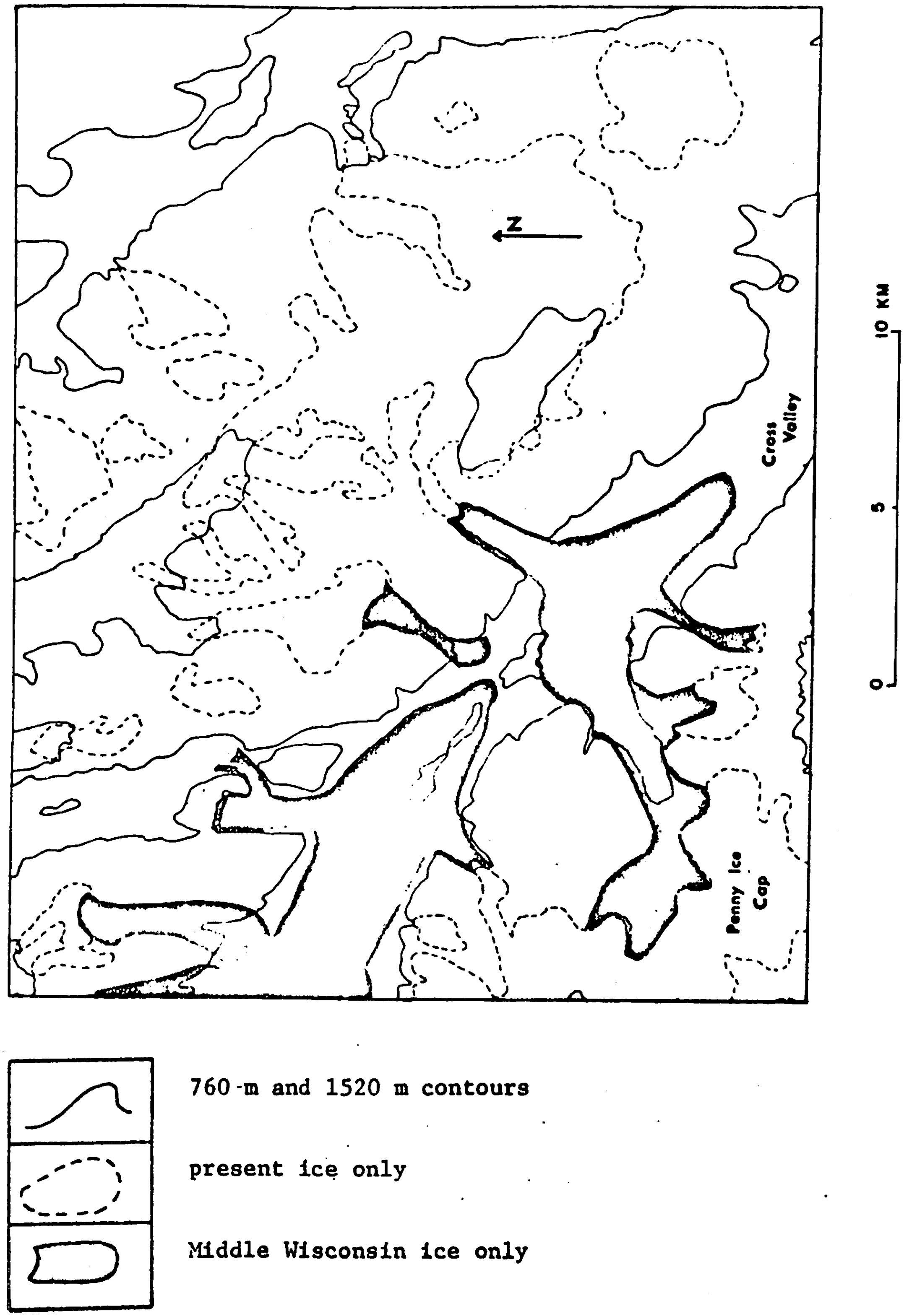


Figure 9-4: Extent of Late Wisconsin ice in Maktak Fiord and the cross valley

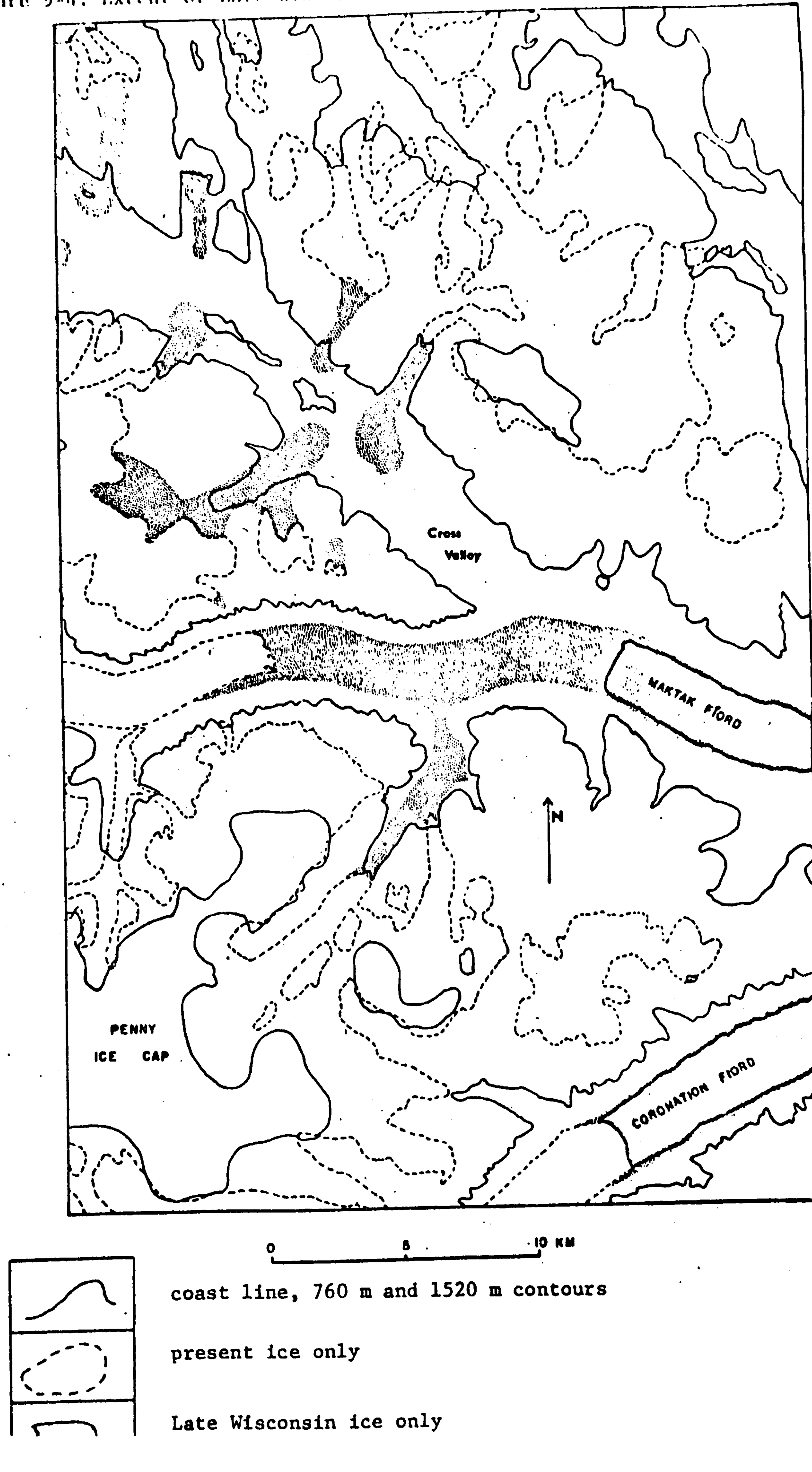
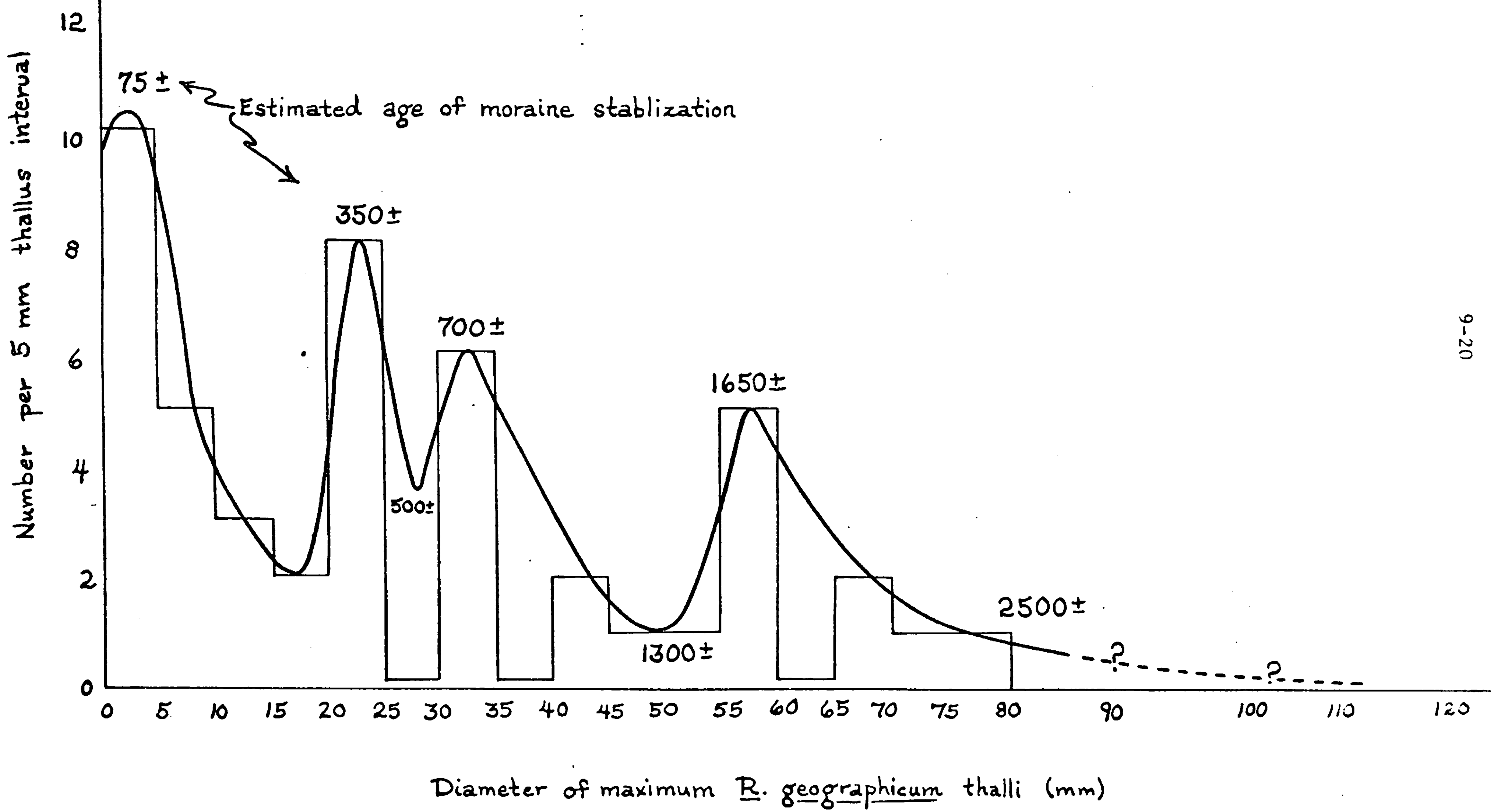


Figure 9-5: Histogram of the number of moraines with specific sizes of R. geographicum of their surfaces



10. PRESENT AND PAST GLACIATION LIMITS, SNOWLINES, AND ICE DISTRIBUTION

J. T. Andrews, R. E. Dugdale, G. H. Miller and L. D. Williams

INTRODUCTION

The mapping of present and past glaciation limits and snowlines for the Okoa Bay area and the wider region of Baffin Island has been a major research interest of the INSTAAR group. Indeed, the study by Andrews, Barry and Drapier (1970) in part contributed to the framing and design of the present work. Investigation of the past and present glacierization of the area provides a necessary link between the Quaternary geology, present climate and paleoclimatology. Studies that originated out of the ARO grant include those of Andrews and Dugdale (1972), Andrews and Miller (1972), Andrews, Barry, Bradley, Miller and Williams (1972), Williams (1972), and Bradley (1972). Aspects of the statistical evaluation of data have been discussed by Andrews, Fahey and Alford (1971) with particular reference to the correlation of cumulative data. Finally, under the terms of the ARO grant, Dugdale (1972) examined the statistical relationship between changes in the ELA, ablation and net balance gradients for Norwegian glaciers both in terms of spatial associations and changes of the relationship over a number of years.

The purpose of this chapter is to synthesise these papers and to attempt to model the mass balance of the Sulung "Glacier". In addition, an assessment is made of the relative importance of global and direct short-wave radiation on the past and present distribution of corrie glaciers. To accomplish this a computer program was written by L. D. Williams to evaluate solar radiation receipt on slopes, taking into account the effect of topographic shadowing. The relationship between glacier distribution and radiation has been discussed by Williams, Barry and Andrews (1972), Williams (1972), and Andrews (in press). The effects of Milankovitch orbital variations on the radiation totals and their probable effect on glacierization has also been examined (Andrews, Barry, Bradley, Miller, and Williams, 1972).

The results represent a new approach to the question of broad regional controls on glacierization. A major conclusion of all these investigations is the finding that the northern area of Cumberland Peninsula (and probably the Peninsula in general) is extremely sensitive to the effects of climatic change. In Chapter 9 the chronology of the last glaciation indicated that the area experienced very early glacierization. Therefore, we consider that the area should be designated a key monitoring region for climatic change which may well prove to be of global consequence.

PRESENT GLACIATION LIMIT, ELA AND CORRIE FLOOR ELEVATIONS

Figures 10.1 and 10.2 are maps of the glaciation limit and lowest glacier ELA's for Baffin Island (Andrews and Miller, 1972). The glaciation limit isa fundamental measure of regional 'glacial' climate because it separates mountain masses that harbor ice caps and glaciers from those that are currently ice free. Figure 10.1 indicates that along the outer coast of Cumberland Peninsula the glaciation limit ranges between 700 and 1,000 , and increases to 1,300 m a.s.l. near the Penny Ice Cap. The lowest ELA's of distinct corrie glaciers lie on average only 200 m below the glaciation limit and have a similar inland gradient (insert B , Figure 10.1).

In order to evaluate changes of snowline, and by inference climate, during the Quaternary some method of estimating former snowlines is required. Traditionally (e.g. Flint, 1971) reliance has been placed on the mapping of the regional corrie (cirque) floor surface as an index of Quaternary snowline shifts. A number of our studies (Andrews et al., 1970; Andrews and Dugdale, 1972 and Williams, 1972) have been concerned with delimiting the corries of Okoa Bay region and describing them by variables which could have had some influence on a broad three-fold categorization of: 1) corries with glaciers, 2) corries with ice patches, and 3) empty corries. In the studies by Andrews et al. (1970) and Andrews and Dugdale (1972) 169 corries were delimited, whereas Williams (1972) mapped 196 corries for a larger area. Trend surfaces were fitted to various sets of the data in order to derive statistically acceptable surfaces. Figure 10-3 is a cross-section through the Okoa Bay area and the Cape Dyer area (to the east and south) showing various 'snowlines' superimposed on the topography (from: Williams, 1972). Note that the trend of lowest glacierized corries occurs 200 to 400 m below the glaciation limit and is, therefore, very similar in position to the lowest glacier ELA's (Figure 10-1). The equivalence between steady state equilibrium ELA's and corrie floor elevation at present adds confidence to our attempts to analyse Quaternary snowline changes by an examination of the floor elevation of empty corries.

The height of the glaciation limit has been considered to be correlated with the height of the average July $0^{o}C$ isotherm. A preliminary test of this assumption (Andrews and Miller, 1972; Figures 5 and 6) indicated that some association does exist but several problems were outlined. This question has been studied in more detail by Bradley (1972), by Bradley and Barry (in prep)and in Chapter 12 of this report. A major problem is the presence of a regional summer inversion up to heights of 300 m. The inversion will clearly affect weather stations at or near sea level, which are the majority.

PAST CORRIE FLOOR ELEVATIONS AND SNOWLINE CHANGES

Approximately half (ie. 96) of the Okoa Bay corries are currently unoccupied. Analysis of their elevations indicates that they occur at lower elevations and with a greater variety of orientations than do glacierized corries. A significant number, (ca. 40) of the empty corries faced south. However, an analysis of corrie floor elevations between north and south-facing empty corries indicated no statistical difference in their heights and that, on average, empty corries occur only 200 m below glacierized corries today. This small elevational difference indicates that Okoa Bay under present climatological conditions is remarkably close to the degree of glacierization that marks a full glaciation! Figure 10.3 illustrates the trend surface of the lowest north-facing empty corries from the two areas. In the case of the Cape Dyer area the similarity in elevation of the two corrie floor surfaces is even more pronounced.

Our data indicate that if the glaciation limit and corrie floor surfaces maintain a parallel and equal relationship to each other, a lowering of the present glaciation limit by about 200 m would promote full-scale glaciation of the region. Assuming that the glaciation limit is correlated with the altitude of the July $0^{o}C$ average isotherm, such a lowering is equivalent to a July temperature decline of $1.4^{o}C$. As noted in the next chapter the summer temperatures have fallen by nearly this amount in the decade of the 1960's. However, because of the inertia in the response of the glaciation limit to climatic change, it is very necessary to ascertain to what climatic period the present glaciation limit refers. If it refers to the temperature minima of the last 300 years or so, then extensive glacierization of Cumberland Peninsula will occur with a climatic deterioration slightly more severe than experienced within that period. In this sense it is very interesting to examine the trend for increased glacial severity from the beginning of the Neoglaciation to the present (Figure 9-5). If this trend continues, the next major deterioration could be sufficient to promote glacierization of the empty corrie basins.

PRESENT AND PAST CORRIE GLACIER DISTRIBUTIONS AND CONTROLS

In any region the reasons why a specific corrie contains a glacier is controlled by regional climatological and topographical considerations and by local site factors. In a few cases local site factors may assume sufficient dominance to cause deviations from a deductive model but such deviations are not of particular interest; what is of great interest is the definition of the regional boundary conditions for the controls on glacier distribution. For the Okoa Bay map sheet we have used existing maps or air photographs to determine variables (selected by Andrews and Dugdale, 1972)(Table 10-1) for each of the 196 corries in the 3 categories - with glaciers, with ice patches, or empty referred to above. The analysis indicates that the major factor is elevation and orientation. Figure 10.4 illustrates a discriminant analysis between corries with and without glaciers based on 5 variables. Note that the corries with glaciers form a compact, unimodal distribution of D values whereas the empty corries appear to be composed of two separate distributions.

The importance of orientation on present corrie glacier distribution (Figure 10.5) suggests that the spatial variability of direct beam radiation in areas of high relief is one of the most important limiting conditions as to whether a glacier can successfully maintain itself at some specific site. Even so it is only an indirect measure. The critical term is net radiation , but the precise relationship between direct solar and net radiation in this area requires further study. Figure 10.6 indicates the spatial variability in computed global radiation on June 21st over the detailed field area under clear skies. Figure 10.7 shows a marked discrimination between empty and ice-filled corries solely on the basis of elevation and global radiation (Williams et al., 1972).

Williams (1972) undertook an intensive study of the importance of radiation and elevation controls on corrie glaciation for the Okoa Bay map sheet and the combined Padloping and Cape Dyer map sheets (these last two with 297 corries). The elevation of all corries and the amount of direct radiation falling at the base of the backwall on August 15th was determined and discriminant analysis carried out to test the thesis implicit in Figure 10.7. The results confirmed the initial hypothesis that well defined boundary conditions exist between empty and ice-filled corries. Figures 10-8 and 10-9 are graphs of the corries plotted against elevation and direct solar radiation for the two regions (from Williams, 1972). The two group discriminant analysis (ice versus no ice) resulted in 138/196 corries correctly classified in Okoa Bay and 196/297 correctly classified in the Cape Dyer region. Separation of the groups in both areas is significant at the 0.01% level by a F-test (Williams, 1972, p.41). The discriminant equations for the two areas are given on Figure 10-8 and 10-9. Williams (1972) pointed out that the slopes of the two are quite different indicating that in Okoa Bay radiation is the primary control on glacierization whereas the steeper gradient of the Cape Dyer discriminant line (representing a value D_o of the discriminant function) indicates that elevation is of greater importance. "Elevation" is, however, some indirect measure of climatological factors. Increasing altitude implies cooler temperatures and a change in the ratio of the two major summer heat sinks: sublimation/melting. However in many mountain areas it also implies an increase of precipitation. At the present state of understanding perhaps we should talk simply of a "complex elevation gradient". In the same way, but possibly to a lesser extent, the major control on insolation is orientation of the corrie but this might also affect such things as snow drift, etc. Because of the pronounced decrease in winter precipitation between Cape Dyer and Broughton Island the steep gradient of Figure 10-9 is interpreted as indicating that in the Cape Dyer area glacierization is controlled mainly by snowfall.

Analysis of the Cape Dyer and Okoa Bay regions thus indicates considerable differences in controls on glacierization, paralleling the conclusions of Andrews, Barry and Drapier (1970) that Okoa Bay and Home Bay (ca. 150 km north of Broughton Island) have different controls on their glacierization. In general our work indicates 1) regional controls on glacierization at a scale of $1 \times 10^4 km^2$ but 2) significant departures at larger scales ($5 \times 10^4 km^2$). Table 10-2 lists the means and standard deviations of three variables for the three categories of corrie in the Okoa Bay and Cape Dyer areas.

The difference in direct radiation between empty and ice-filled corries is in the ratio of 2:1 and amounts to between 100 and 120 ly on average (Table 10-2). This is a very large difference and one that is too big to be accounted for by Milankovitch orbital changes. Calculations indicate that the extreme "glacial" condition results in a reduction of ~50ly on August 15th. Such a change only leads to a small percentage increase in the number of empty corries that would become glacierized (Andrews, et al. 1972).

Williams (1972, p.65) used the discriminant functions shown in Figures 10-8 and 10-9 to develop preliminary models relating to the effect of changes in radiation and snowfall on corrie glacierization. These models are shown in Figure 10-10; their rationale is summarized in Table 10-3. The implications of these models are that high south-facing glaciers are sensitive to changes in snowfall whereas low, north-facing corries are sensitive to temperature (radiation). The amount of change required to promote extensive glacierization can be judged from Figure 10-11 which shows the density of corries in elevation/radiation space at a 2% contour interval (from 2 to 10%). A surprising result is the strong modality of the density distributions with the number of corries receiving large amounts of radiation about equal to those receiving rather small smounts, and a distinct lack of corries in the intermediate category (Williams, 1972, p.68). Comparison of Figures 10-10 and 10-11 suggests that increased snowfall is most likely to change the present discriminant line so as to result in maximum corrie glaciation.

We suggest that the early glaciation of the region commences with the onset of increased winter accumulation probably associated with a decrease in summer ablation. In other words a climatological pattern very similar to that which prevailed during the last decade (see Chapter 11) and typified in its extreme conditions by the 1969-1970 mass balance year (see Chapter 6). During the mid-part of the last glaciation the region probably received less precipitation than today, the climate would be more continental and radiation controls on glacierization would become critical. Sauberger and Dirmhirn (1951) note that with high albedos and clear skies that the radiation balance in summer can become negative. Such a situation might have prevailed over eastern Baffin Island during the maximum of the last glaciation when Lamb and Woodroffe (1970) suggest the region was dominated by a high pressure call.

THE SULUNG GLACIER - MODELS OF ITS MASS BALANCE

Mapping of the Sulung Valley enabled us to reconstruct the outline of the former Sulung Glacier and to calculate its area and volume relationships with elevation. These calculations yield estimates of past "glaciologies" of the Sulung Glacier under the assumptions of steady state equilibrium. This assumption is considered correct for the period during which the massive end moraines were deposited at 600 m a.s.l. at the mouth of the valley (see Chapter 9). Plan and cross-sections of the glacier are illustrated in Figures 10-12 and 10-13. The glacier was about 90 m thick, 6 km in length with a total area of 6.5 km^2. The area under the ELA was ca. 8,900 m^2. Areas under other contours were measured using the 1:125,000 maps. The information on the cross-sectional area and the area of the glacier lying above that particular contour can be used to estimate the average velocity of ice moving through the cross-section.

Iterative methods were used to develop a number of steady state models. There is, of course, an infinity of possible models but we selected only those considered "realistic" in terms of our knowledge and concept of past climates in the area. In all four models were generated and for each one we obtained: 1) the ablation gradient, 2) velocities along the length of the glacier, and 3) the net balance. The prime inputs were the amount of winter accumulation and the accumulation gradient. The plausibility of each model can be assessed from the calculated velocities based on the mass balance compared with estimated velocities based on ice depth, glacier slope and englacial temperatures. Tables 10-4, 5, 6 and 7 list the main data for the 4 models (the work was performed by Mr. Thomas Geist, University of Colorado, as a class project). All the 4 models were based on our earlier suggestion that increased winter accumulation is the prime cause of extensive glacierization in the area, particularly during the initial states of a glacial "cycle". Increased accumulation is required to counter the additional radiative energy incident upon south-facing slopes (nearly 100 ly/day on June 21st). The velocities of the 4 models indicate a rapid increase in ice movement near the ELA because of the change in cross-sectional area of the glacier. However, models 1 and 2 produce very high velocities and for this reason are not as favoured as #3 and 4 which are also closer to current conditions. Model 3 is based on winter accumulation rates similar to those experienced today on the north-facing Boas Glacier which may be about twice the present snowfall, whereas model 4 is based on higher total accumulation but the accumulation decreases from the snout of the Sulung Glacier to the divide (in other words a reverse pattern to that discussed in Chapter 6 where current accumulation decreases upslope on the Boas Glacier and is less on the south side of the divide, see Chapter 3). This model would require considerable southwesterly flow to bring moisture into the region and would also probably require that the Penny Ice Cap be removed as its 2,000 m elevation might act as a barrier. During the last interglacial the Barnes Ice Cap disappeared (Terasmae, Webber and Andrews, 1966) and the Penny Ice Cap may have undergone the same history - in which case southwest flow during the early stages of glacierization would be feasible (cf. Andrews, et al., 1970).

One problem with the steady state mass balance solutions is that they do not allow for adaptation of the glacier to changing regimes. Although model 4 might apply to early glacial conditions our work clearly demonstrates that the Sulung Glacier was present and active during mid and late Wisconsin times when a large continental ice sheet lay to the west. If Baffin Bay was largely ice covered then precipitation might have been less than experienced today. However, once developed the Sulung Glacier, and other south-facing glaciers that occupied presently empty basins, might have survived a reduction in precipitation if this was accompanied by a general tendency to increased continentality so that the south-facing exposure was partly counteracted by low summer temperatures and an increase in sublimation over melting (as in the 1970 summer on the Boas Glacier).

Our finding that the elevations of north- and south-facing empty corries are on average the same indicates that during full glacial conditions we have to be able to explain the lack of asymmetry caused by direct radiation, which is apparently important under current climatic conditions, in order to explain the pattern of corrie glacierization (Williams, 1972). The lack of asymmetry might be explained by a rather low, even accumulation and by low summer temperatures to make the ratio: sublimation/melting very large. In this way the difference between north- and south-facing slopes would be relatively insignificant and yet rapid deglaciation of south-facing glaciers would occur with an increase in temperature.

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TABLE 10-1

CHARACTERS (i.e., VARIABLES) USED TO DESCRIBE THE CORRIES OF THE OKOA BAY AREA, EAST BAFFIN ISLAND, N.W.T. AND NUMBER AND NUMERICAL CLASSES OF CHARACTER STATES[a]

Character 1:	Maximum elevation of mountain into which corrie is excavated. Eight equal classes commencing at 590 m a.s.l. with class intervals 125 m. (Protection from insolation; inducing snow drifting)
Character 2:	Maximum elevation of the corrie. Eight equal classes commencing at 220 m with class intervals 172 m. (Summer temperature, ratio: sublimation/melting)
Character 3:	Character 1 minus Character 2. Eight equal classes commencing at 0 m with class intervals 70 m. (Snow drifting, protection from insolation)
Character 4:	Elevation of the corrie lip. Eight equal classes commencing at 150 m a.s.l. with class intervals 144 m. (Summer temperature; ratio: sublimation/melting)
Character 5:	Maximum vertical development of the corrie. Eight equal classes commencing at 40 m with class intervals 70 m. (Protection from insolation)
Character 6:	Length of the corrie long axis. Eight equal classes commencing at 220 m with class intervals 300 m. (Size of accumulation basin; insolation)
Character 7:	Maximum breadth of the corrie at normal to the long axis. Eight equal classes commencing at 290 m with class intervals 246 m. (Size of accumulation basin; insolation)
Character 8:	Ratio of corrie length/breadth. Eight equal classes starting at 0.258:1 with class intervals 0.581. (Effect on insolation)
Character 9:	Ratio of corrie length/height. Eight equal classes commencing at 0.629:1 with class intervals 4.171. (Effect on insolation)
Character 10:	Direction of long axis. Eight equal classes beginning at 0° with class intervals of 45°. (Relation to insolation, direction of prevailing winds, snowdrift)
Character 11:	Type of ice body. 1 = empty, no ice; 2 = snow patch; 3 = ice patch; 4 = corrie glacier.
Character 12:	Height of the corrie backwall. Eight equal classes commencing at 60 m with class intervals of 36 m. (Protection from insolation)
Character 13:	Ratio of height of backwall to corrie length. Eight equal class intervals commencing at 1:0.629 with class intervals of 5.0. (Protection from insolation; snow drifting?)
Character 14:	Aspect of corrie with regards to the mountain (see Fig. 1B). Eight equal class intervals starting with 0–44 with class intervals of 45°. Effect on insolation; relationship to predominant and storm winds)
Character 15:	Distance from the outer coast (arbitrary line running NW–SE). Eight equal classes commencing at 3.4 km with class intervals of 4.9 km. (Effect of continentality; effect of outer fog and stratus banks; effect of pack ice on local temperatures)
Character 16:	Nearest neighbor densities (see text for further definitions). Eight equal classes commencing at 0 with class intervals of 4. (Effect of nearby corries on accumulation)
Character 17:	Estimated volume of glacier ([length × breadth × height]/2). Eight equal classes commencing at 0 with class intervals of $53 \times 10^7 m^3$. (Size of accumulation basin; protection from insolation)

[a] The suggested climatological/glaciological impact of each character is included in brackets (see Fig. 1 and Table 2).

TABLE 10-2

Population parameters for glacierized corries (G), corries with ice patches (P), and empty corries (E). (from Williams, 1972, Table 3).

		Okoa Bay		Cape Dyer	
		$\bar{X}$	σ	$\bar{X}$	σ
Elevation(m)	G	1014	170	964	142
	P	839	194	870	172
	E	773	274	652	282
Vector Orientation	G	18	36	-10	65
	P	32	64	22	91
(+ or - from north)	E	105	83	128	87
Direct Radiation	G	122	69	144	93
(ly) on Aug.15th	P	171	73	190	107
†	E	240	89	240	96

Numbers in group: G = 64, P = 36, E = 96   G = 177, P = 45, E = 75

TABLE 10-3

RESPONSES OF DISCRIMINANT LINE TO CLIMATIC CHANGE

Climatic Change	Influence on Glaciers	Effect on Discriminant
Summer Temperature Increase	1. More isotropic ablation	1. Higher slope
	2. Less radiation permitted to maintain glac.	2. Downward translation
	3. Ablation greater at given elevation	3. Translation right
Summer Temperature Decrease	1. Less isotropic Ablation	1. Lower slope
	2. Glacierization possible with more radiation	2. Upward translation
	3. Ablation less at given elevation	3. Translation left
Decreased Regional Snowfall	1. Ablation more critical accumulation less	1. Lower slope
	2. Less radiation permitted to maintain glac.	2. Downward translation
	3. Less accumulation at given elevation	3. Translation right
Increased Regional Snowfall	1. Ablation less critical accumulation more	1. Higher slope
	2. Glacierization possible with more radiation	2. Upward translation
	3. More accumulation at given elevation	3. Translation left

TABLE 10-4

Model 1, Sulung Glacier

Winter mass balance ($\bar{b}_w$) = 0.7 m H_2O; Accumulation gradient = 0;

Ablation gradient in ablation zone = 5.5 mm m^{-1}, in accumulation zone = 2.75 mm m^{-1}

Area (km^2)	Elevation(m.a.s.l.)	$\bar{b}_n$	Velocity (m yr^{-1})	Cross-section area (m^2)
1.87	600-750	-1.00	35.6	5300
0.5	750-850	-0.29	22.8	8900
0.5	850-900	+0.07	20.6	9660
1.41	900-1050	+0.36	8.0	18,900
0.904	1050-1200	+0.64	5.5	16,700
1.3	>1200	+0.70	--	

TABLE 10-5

Model 2, Sulung Glacier

Winter mass balance ($\bar{b}_w$) = 0.5 m H_2O; Accumulation gradient = 0; Ablation gradient in ablation zone = 3.95 mm m^{-1}, in accumulation zone = 1.98 mm m^{-1}

Area (km^2)	Elevation (m a.s.l.)	$\bar{b}_n$	Velocity (m yr^{-1})	Cross-section area (m^2)
1.87	600-750	-0.74	27	5300
0.5	750-850	-0.25	17	8900
0.5	850-900	+0.05	16	9660
1.41	900-1050	+0.299	6	18,900
0.904	1050-1200	+0.45	4	16,700
1.3	>1200	+0.500	-	--

TABLE 10-6

Model 3, Sulung Glacier

Winter mass balance ($\bar{b}_w$) = 0.35 m H_2O; accumulation gradient = 0; Ablation gradient in ablation zone = 2.9 mm m^{-1}, in accumulation zone = 1.45 mm m^{-1}.

Area (km^2)	Elevation (m.a.s.l.)	$\bar{b}_n$	Velocity (m yr^{-1})	Cross-section Area (m^2)
1.87	600–750	-0.51	17.8	5300
0.5	750–850	-0.15	11.4	8900
0.5	850–900	+0.03	10.3	9660
1.41	900–1050	+0.18	3.0	18,900
0.904	1050–1200	+0.32	2.7	16,700
1.3	>1200	+0.35	-	-

TABLE 10-7

Model 4, Sulung Glacier

Winter mass balance ($\bar{b}_w$) = 0.7 m H_2O; Accumulation gradient = -0.44 mm m^{-1};

Ablation gradient in ablation zone = 1.7 mm m^{-1}, in accumulation zone = 0.85 mm m^{-1}

Area (km^2)	Elevation (m.a.s.l.)	$\bar{b}_n$	Velocity (m yr^{-1})	Cross-section Area (m^2)
1.87	600-750	-0.22	7.9	5300
0.5	750-850	-0.06	5.0	8900
0.5	850-900	+0.02	4.5	9660
1.41	900-1050	+0.06	1.9	18,900
0.904	1050-1200	+0.12	1.5	16,700
1.3	>1200	+0.19	-	-

FIGURE 10-1

Figure 10-1: Glaciation limits and lowest equilibrium line altitudes for southern Baffin Isłand

Figure 10-2

Figure 10-2: Glaciation limits and lowest equilibrium line altitudes for northern Baffin Island.

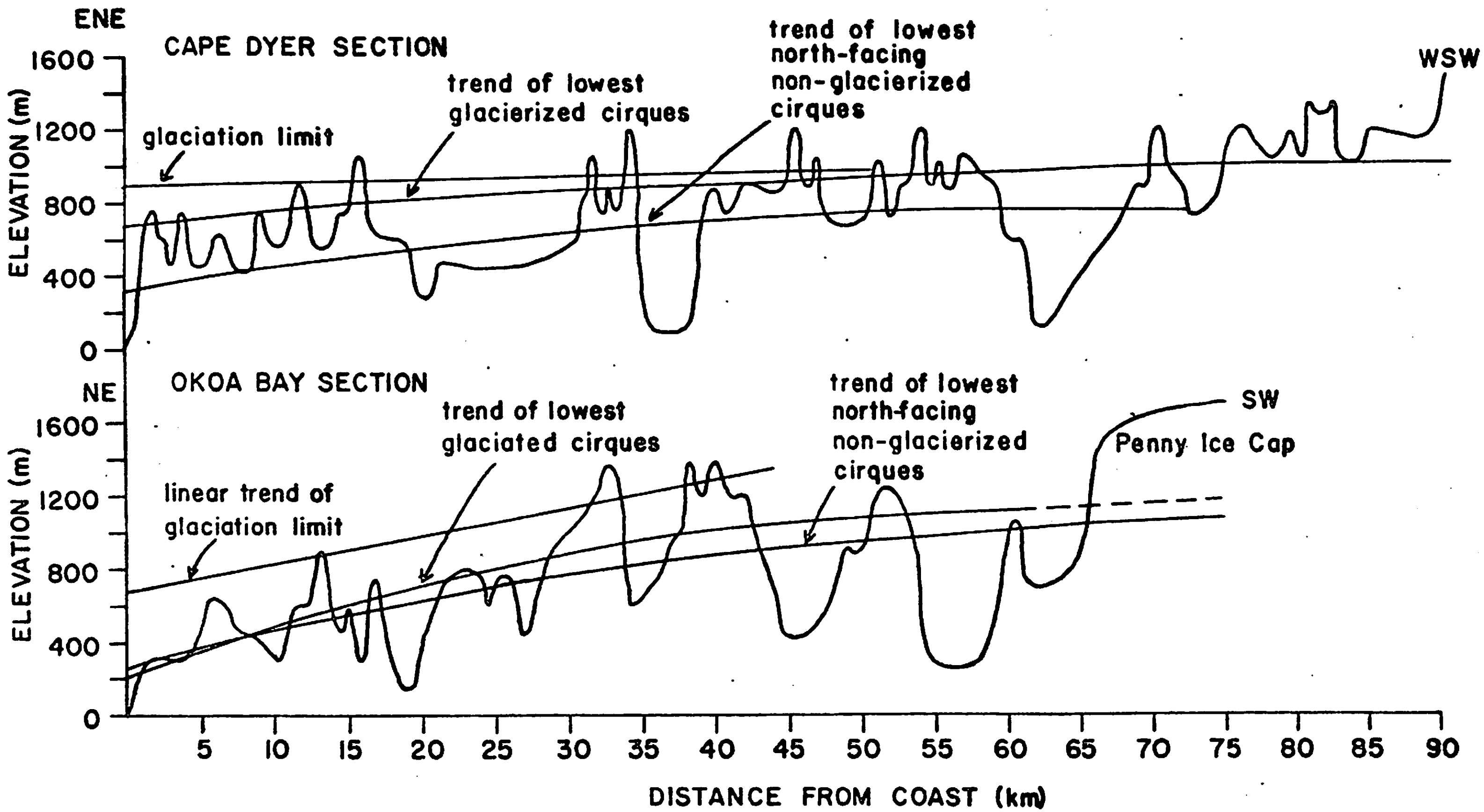


Figure 10-3: Cross-section through th4 Cape Dyer area and Okoa Bay area showing the generalized topography and trend surfaces of various 'snowlines'.

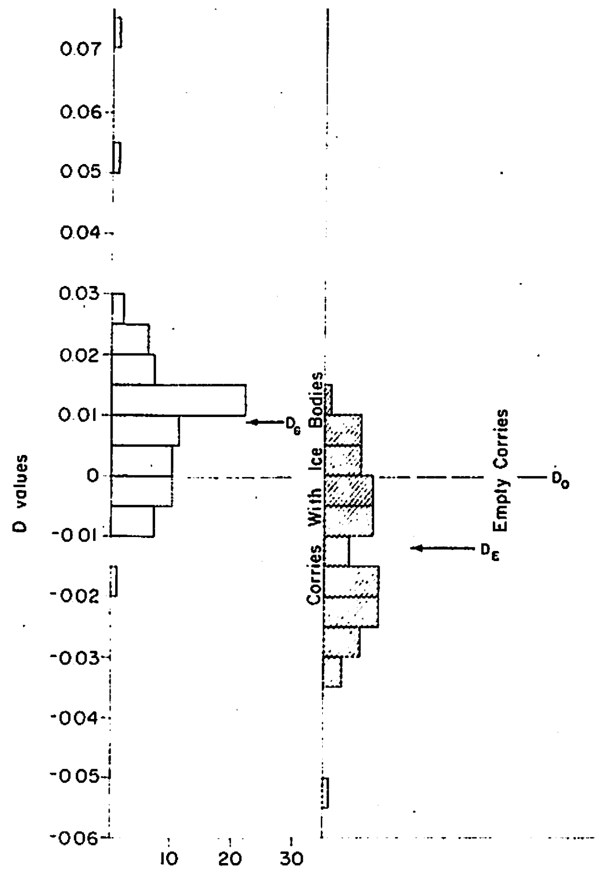


Fig 10-4 Results of discriminant analysis on the two corrie groups, empty and ice-filled. D_G and D_E are the means of the respective groups and D_o the mean discriminant value. Plot shows frequency histograms of the two groups against the discriminant function value.

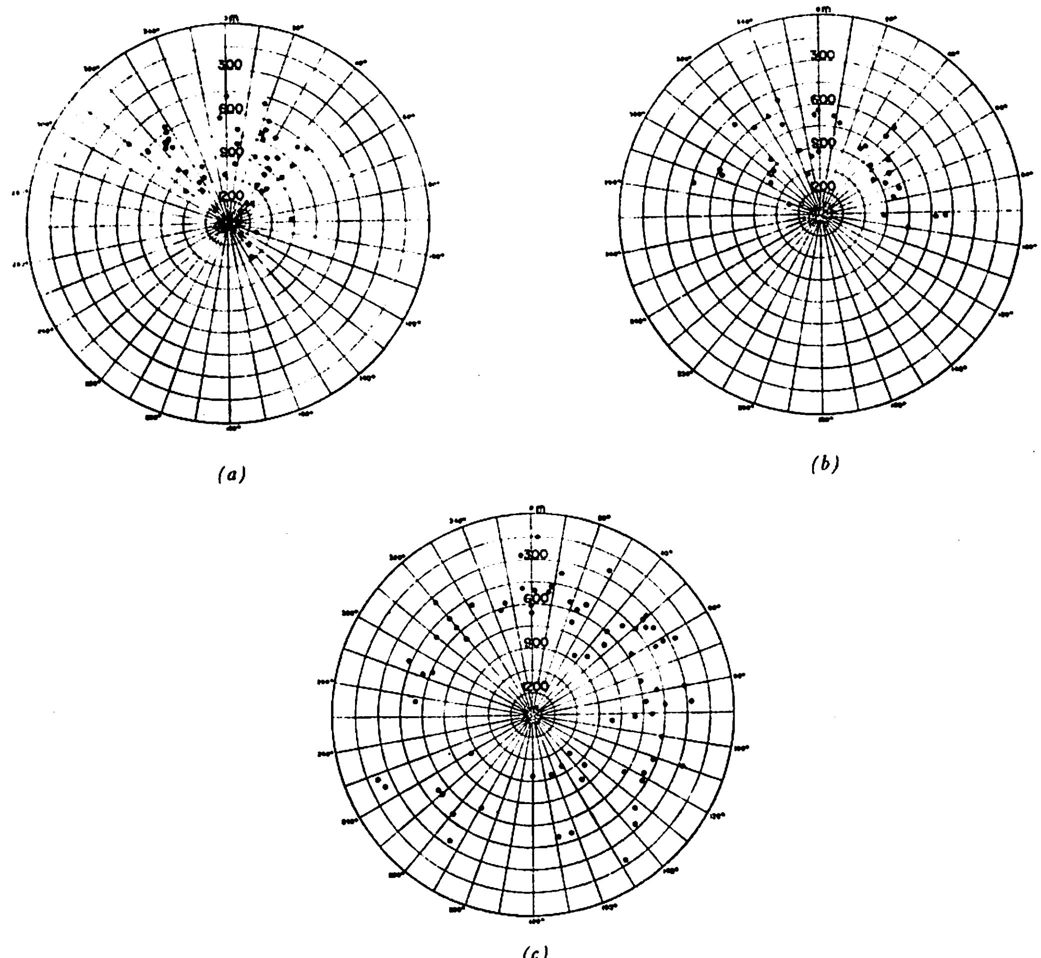
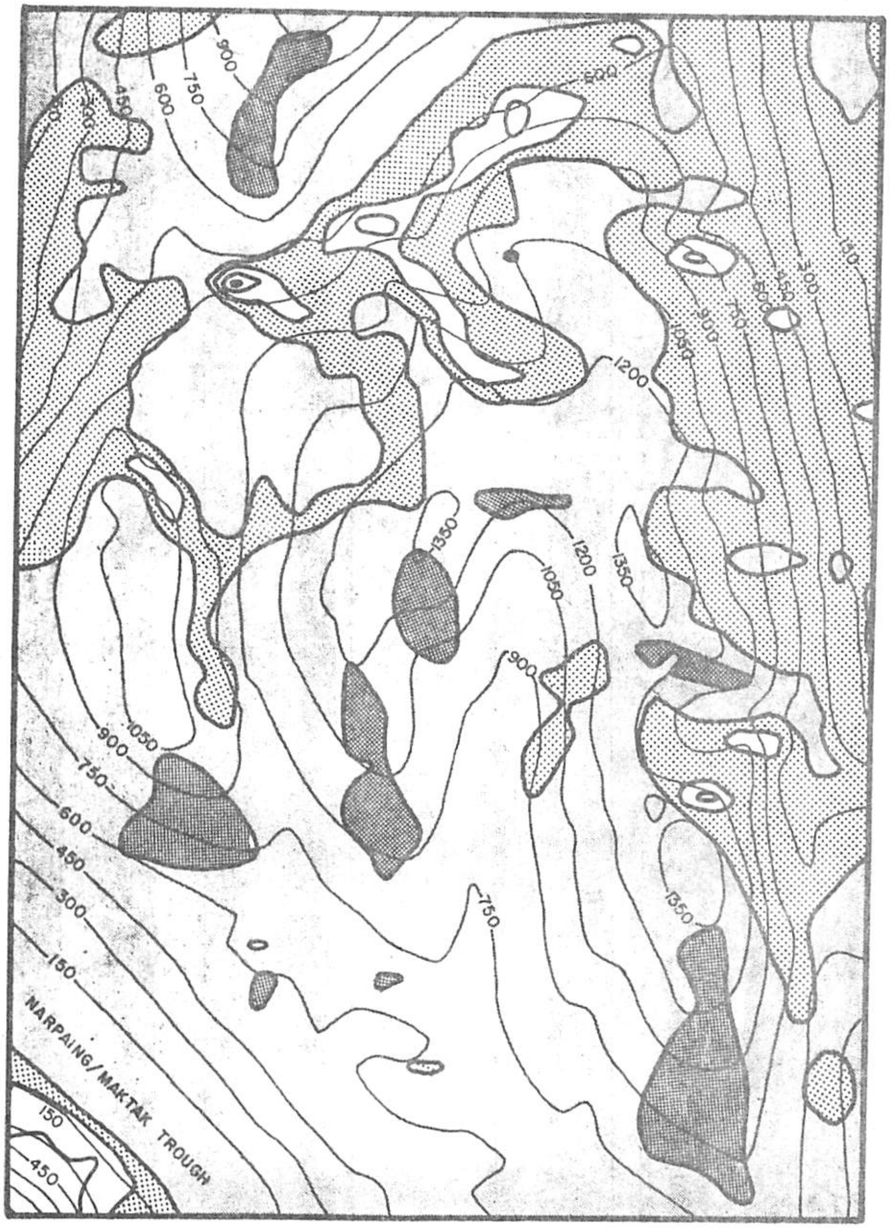


Fig 10-5 *Cirque data for Okoa Bay: A: orientation and elevation of cirque floors containing glaciers. B: orientation and elevation of cirque floors containing ice patches. C: orientation and elevation of all cirque floors in the Okoa Bay map area*



— 900 — Elevation in meters
——— Global radiation in cal cm^{-2} day^{-1}, June 21
(Contour interval = 100 cal cm^{-2} day^{-1})

< 500
600 - 700
500 - 600
> 700

Fig 10-6 Computed global radiation for 21 June under clear skies for the rectangular area in Fig. 1-2 superimposed on the topographic map of the area. Comparison should be made with Fig. for the relationship of global radiation and distribution of ice bodies.

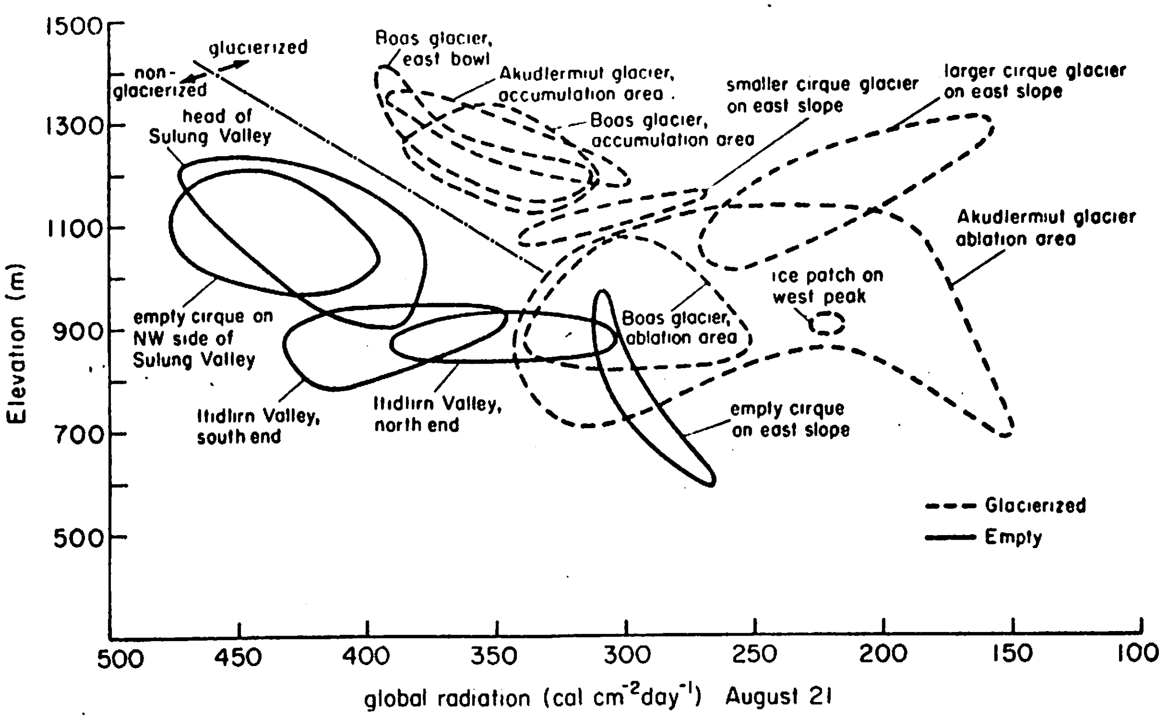


Fig 10-7. Comparison of glacierized and non-glacierized valleys and cirque basins of Fig.12 with respect to elevation and computed global radiation for 21 August. Each curve encloses the spread of values of elevation and global radiation for the individual basins.

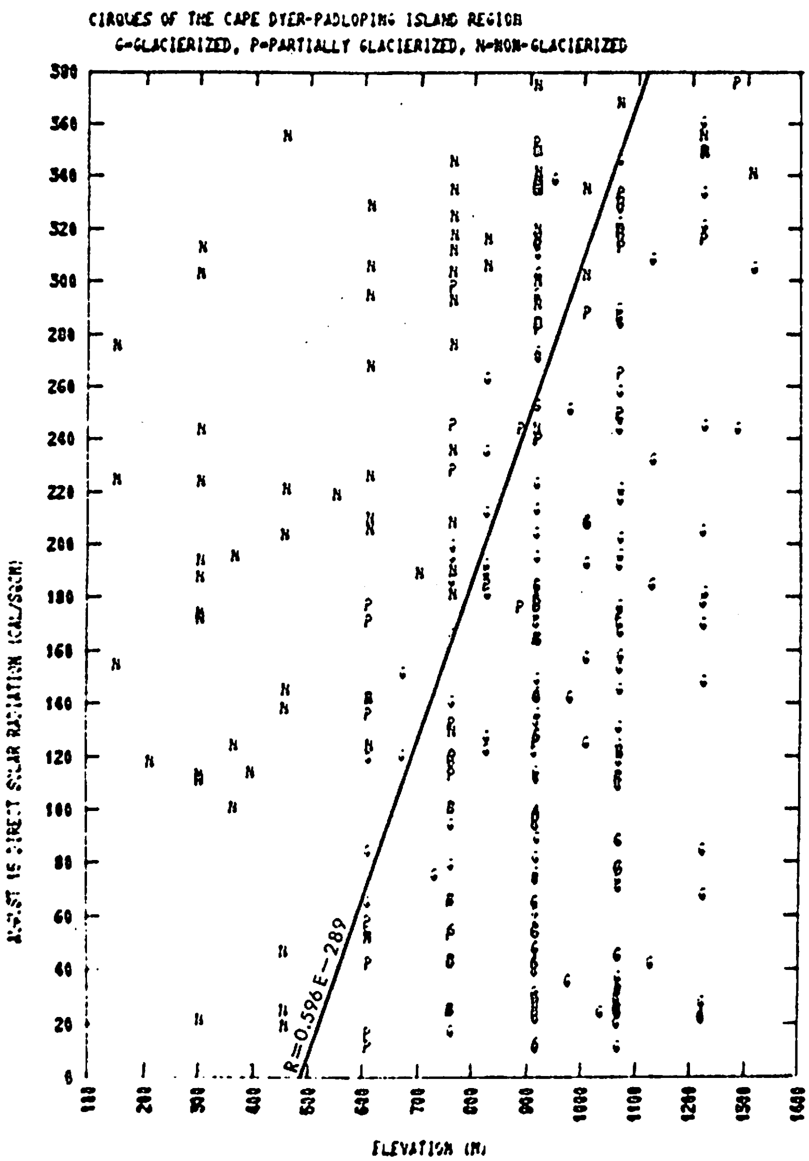


Figure 10-8: Corries of the Cape Dyer area plotted against elevation and computed global radiation showing the discriminant line dividing empty and ice filled.

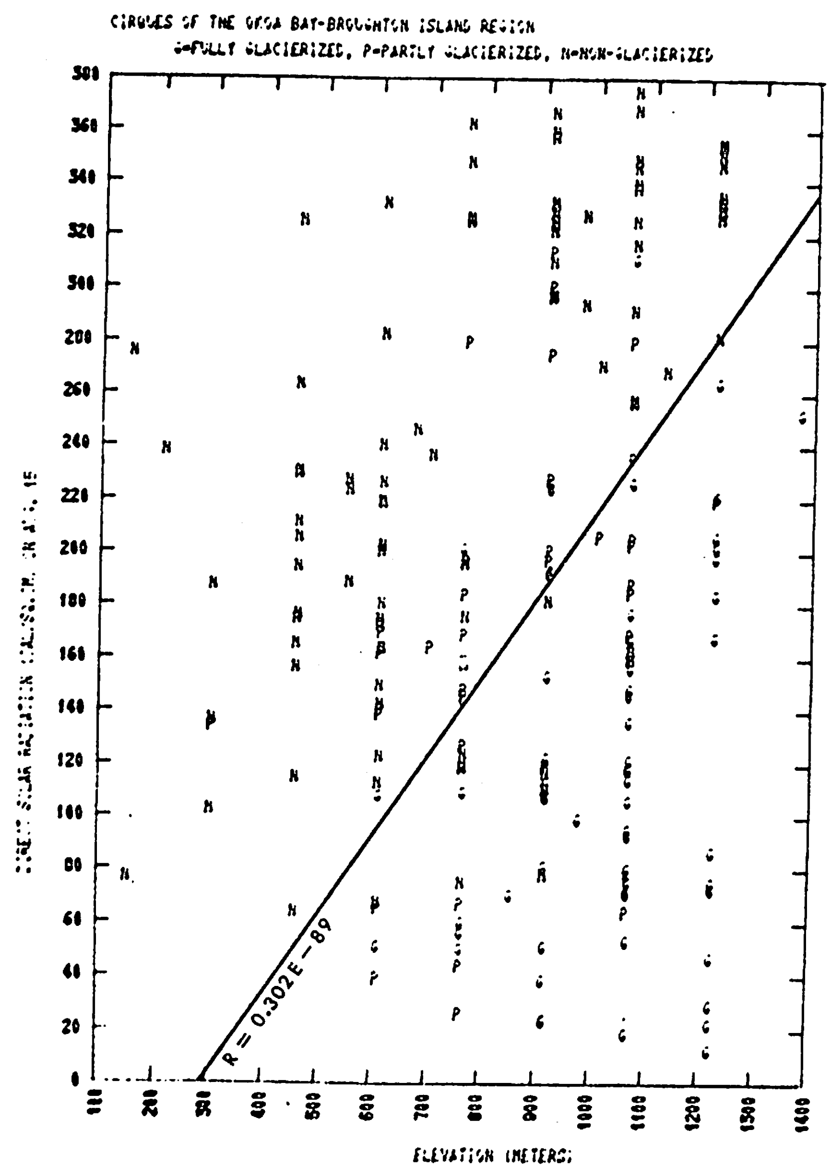


Figure 10-9: Corries of the Okoa Bay area plotted against elevation and computed global radiation showing the discriminant line dividing empty and ice filled - note the difference in slope of the discriminant lines.

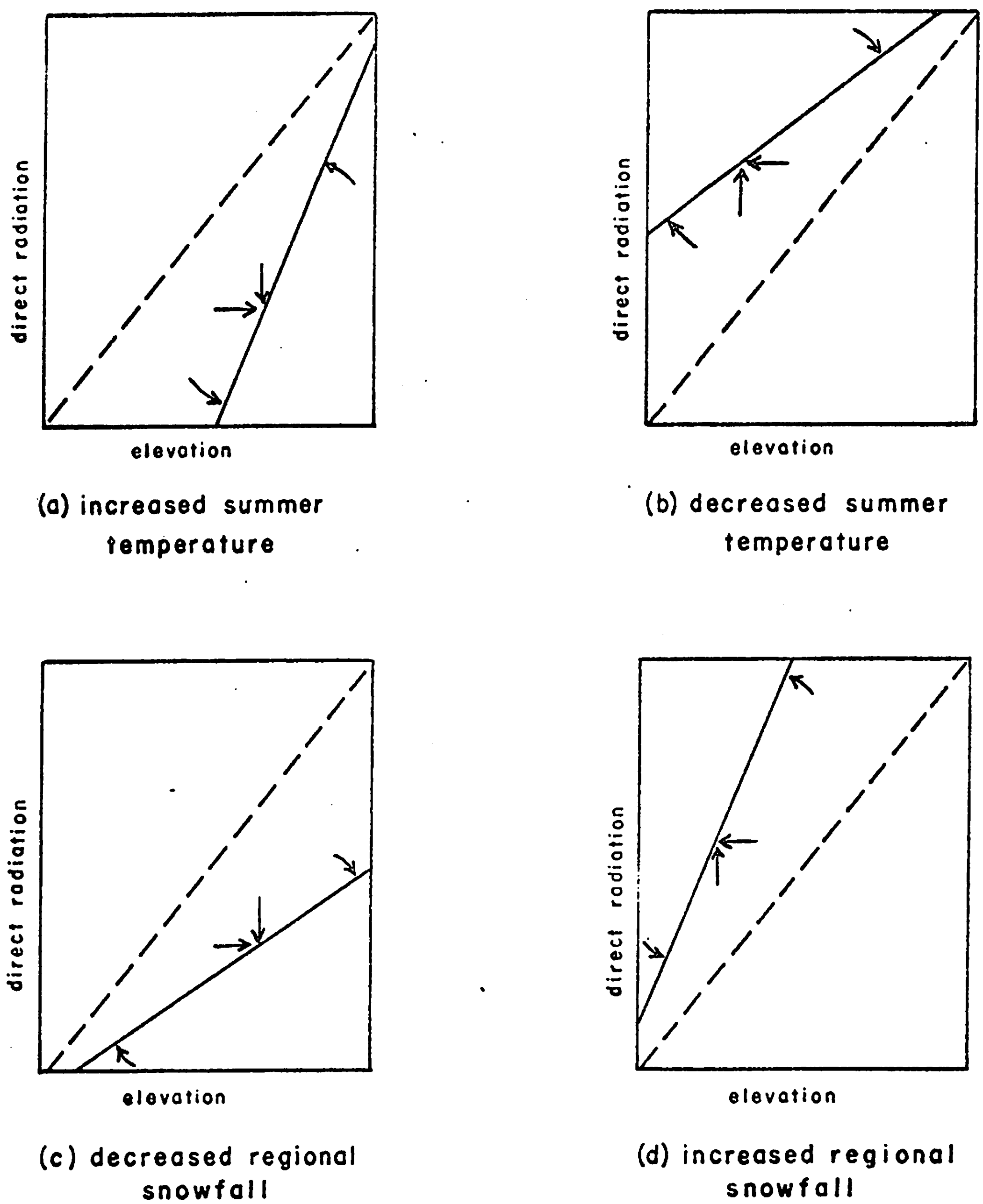
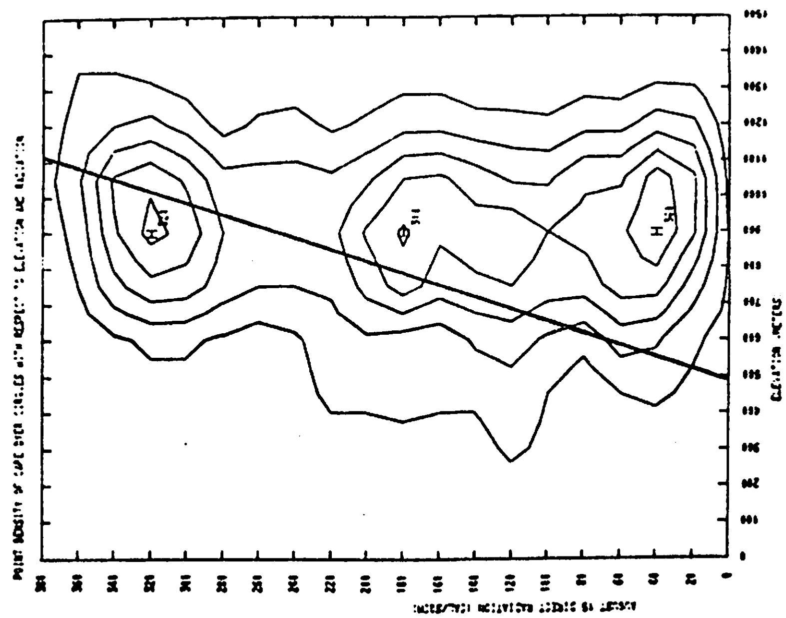


Figure 10-10: Response of discriminant to climatic change



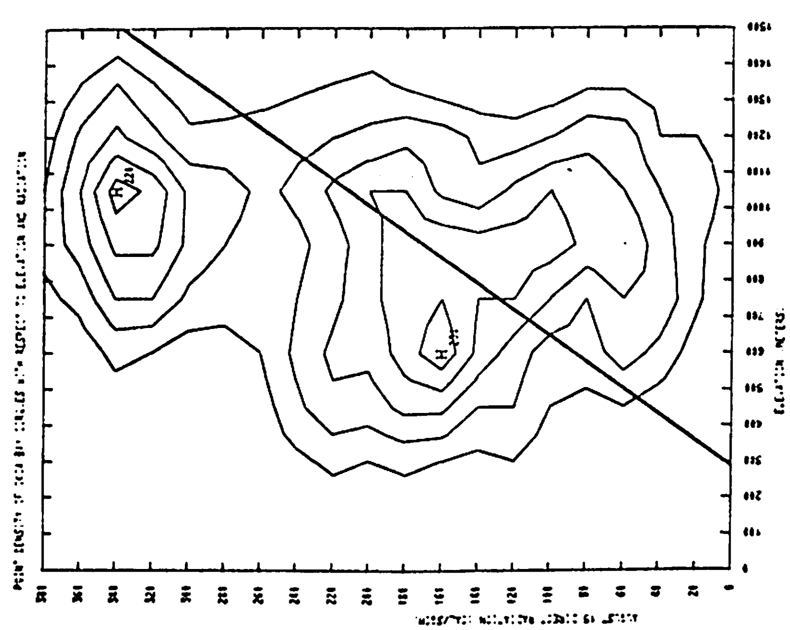


Figure 10-11: Point density diagrams (against elevation and August 15th global radiation) for (top) Cape Dyer and (bottom) Okoa Bay and the trend of the discriminant.

Figure 10-12

LONG PROFILE OF SULUNG GLACIER (Phase 2)

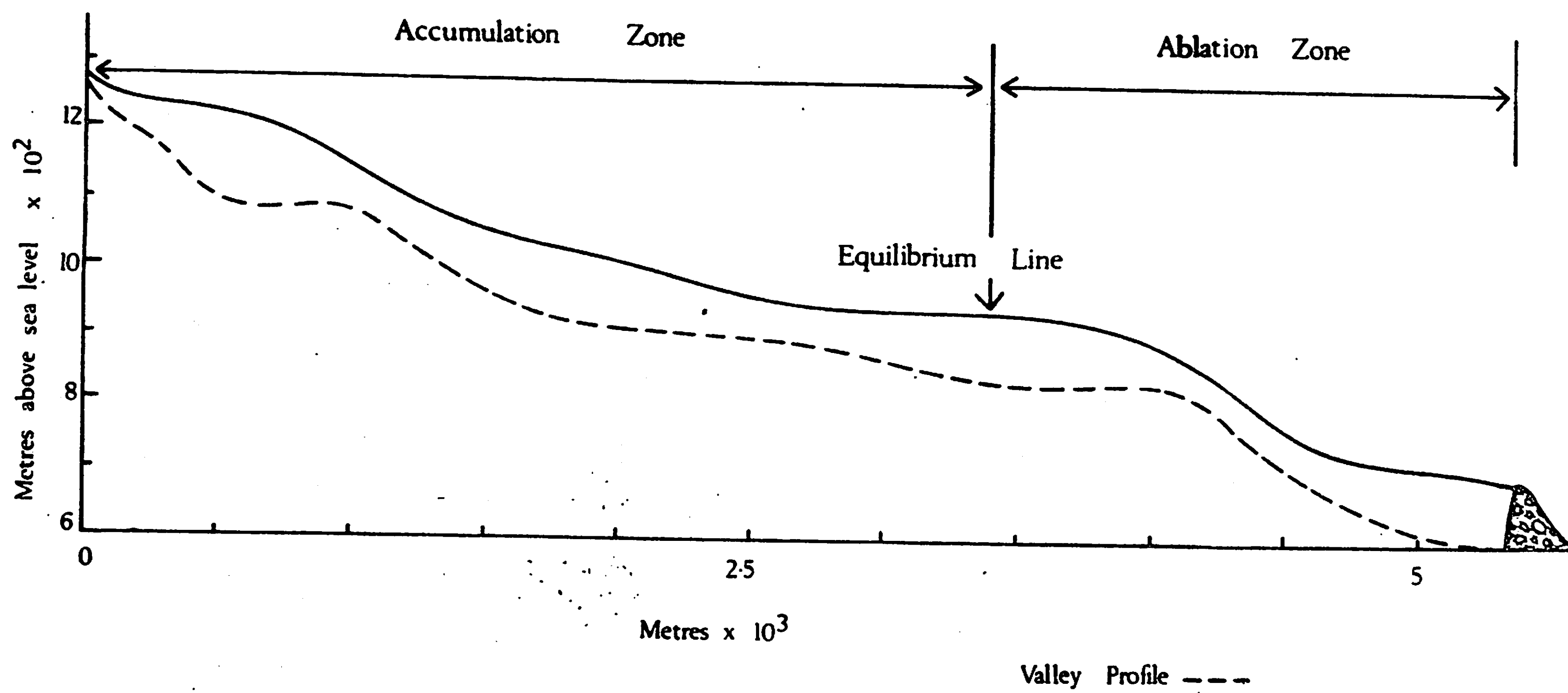
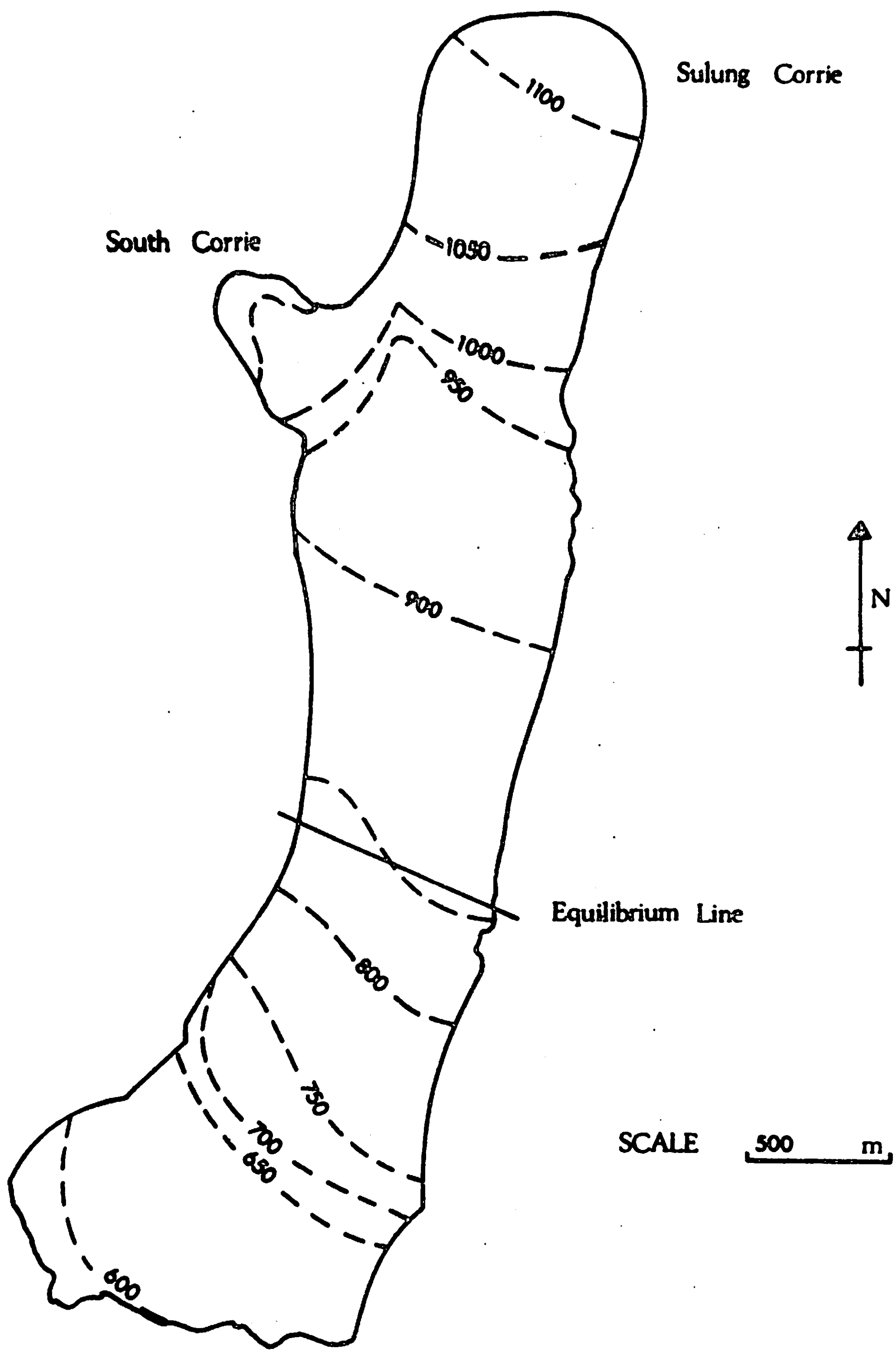


Figure 10-13: SULUNG GLACIER (PHASE 2)



Contours : metres above sea level

11. SEASONAL CLIMATIC FLUCTUATIONS DURING THE PERIOD OF INSTRUMENTED RECORDS, AND FIELD EVIDENCE FOR INCREASED GLACIERIZATION DURING THE LAST DECADE

R.S. Bradley and G.H. Miller

INTRODUCTION

A number of studies in recent years have been concerned with climatic fluctuations on a global or hemispheric basis (Putz, 1971; Treshnikov and Borisenkov, 1970; Mitchell, 1961, 1963). A notable feature of these studies is that in the northern Hemisphere the regions of greatest warming from the 1880's to the 1930's or later have been in higher latitude zones. Similarly areas of greatest cooling over the last 30 years have also been in these regions. These indicate that the largest negative changes occurred in higher latitudes (>60°N), suggesting that a return to conditions of the late 19th century is underway. Both the early 20th century warming trend and the subsequent cooling has been most marked in "winter" months (December, January, and February).

Such studies indicate that the Arctic and Sub-arctic regions are extremely sensitive to climatic fluctuations, and may be considered indicators of hemispheric trends.

The use of annual values to evaluate climatic fluctuations (e.g. Longley, 1953), can be misleading when seasonal trends are antithetical. This study examines on a seasonal basis instrumentally recorded climatic data for Baffin and adjacent islands and illustrates that significant fluctuations have occurred over the past 60 years. A further analysis of data for other parts of the Canadian Arctic Archipelago is presently underway.

Our glacio-geomorphological and climatological studies of eastern Baffin Island indicate that Baffin Island is extremely sensitive to small climatic shifts (Andrews, et al., 1972), a fact noted first by Tarr in 1897. Ice-filled and ice-free cirques in the Cape Dyer area (eastern Cumberland Peninsula) shows an elevation difference of only 200-300 m (equivalent to approximately 1.2°C mean summer temperature) between these two groups (Chapter Ten). Thus, any persistent changes in ablation season temperatures could result in significantly increasing the degree of glacierization of the area (Bradley and Miller, 1972).

No use has been made of Baffin Island data for studies of climatic change. Longley (1953) combined data from Resolution and Nottingham islands into a "Hudson Strait" record and found a rise in mean annual temperature during the 1930's and 1940's of approximately 0.5°C. For stations on western Greenland, Putnins (1971) found a general tendency for July temperatures to increase from approximately 1880 to 1930-40, followed by a cooling trend to 1955 when the analysis ends. January temperatures show a similar pattern but differences at some stations are apparent.

The meteorological station inventory for the Northwest Territories indicates that over 30 stations have been operative on or around Baffin Island since November 1903 when observations were first made on Blacklead Island (65°N, 66°12'W) in the Cumberland Sound. However, the earliest record of any length is that of Lake Harbour (Figure 11-1). Twelve stations have been chosen for detailed analysis; of these, eight are currently operative (see Figures 11-1 and Table 11-1). For the purposes of this study, two seasons were recognised - a summer or ablation season (June, July and August) and a winter or accumulation season (September to May). For glaciological purposes, this division is adequate; any melting of snow or ice on Baffin Island is almost entirely restricted to the three summer months selected (Chapter 6).

BAFFIN AS A CLIMATIC REGION

A primary interest was to determine to what extent climatic fluctuations over Baffin Island could be considered synchronous both in time and space. Product moment correlation coefficients between all stations operative between 1959 and 1969 were computed for seasonal mean temperatures and seasonal precipitation totals. The seasonal temperature coefficients are shown in Tables 11-2(a) and 11-2(b), for accumulation and ablation seasons respectively. Clearly in both seasons, correlation coefficients are extremely high (over half the correlations are significant at the 1% level or greater and most are significant at the 5%level or greater), indicating that temperature fluctuations were felt throughout the area. Only the summer record for Frobisher Bay appears to be entirely anomalous and this is presumably due to local factors influencing the record. It is suggested that the presence of open water early in the summer may lead to more frequent conditions of fog and thus result in the obvious anomaly. Dewar Lakes is the only 'interior' station on Baffin Island and data indicate that its temperature record is synchronous with the coastal sites particularly those on the east coast. Thus, it would appear that the coastal stations are not as unrepresentative of inland temperature conditons as has been suggested (Andrews, Barry, and Drapier, 1970).

Correlation coefficients between all stations (1959-69) for seasonal precipitation totals show clearly the effect of local topographic factors in influencing the amount of precipitation received at a site. Only occasionally do two station records correlate at the 5% level or greater indicating a generally random pattern of precipitation. Accumulation season correlations are slightly higher than in the ablation seasons but do not show a clear regionally coherent pattern to the extent that the temperature records illustrate.

SEASONAL FLUCTUATIONS OF TEMPERATURE AND PRECIPITATION

Seasonal fluctuations of temperature and precipitation were examined by the use of weighted binomial running means. Three or five year periods were weighted depending on the length of record. Weighting was carried out according to the formulae:

$$T_1 = \frac{(T-1) + 2T + (T+1)}{4} \qquad (11\text{-}1)$$

for 3 year weighted periods, and

$$T_1 = \frac{(T-2) + 2(T-1) + 4T + 2(T+1) + (T+2)}{10} \qquad (11\text{-}2)$$

for 5 year weighted periods, where T_1 is the value plotted at time T. The effect of such weighting is simply to smooth out the higher frequency (short-term) variations while rendering the lower frequency (longer-term) variations more clearly.

Ablation Season Temperatures (Figures 11-2 and 11-3):

The 1930's, when a number of stations began operations, were characterized by a cooling trend through to approximately 1943 (1947 at Arctic Bay). This fall in temperatures amounted to approximately 1.4°C at Nottingham Island, 1.1°C at Resolution Island, 1.8°C at Lake Harbour and 1.2°C at Pangnirtung.
This was followed by an increase of mean summer temperatures to approximately 1949-50 (1.7°C at Nottingham Island, 1.0°C at Arctic Bay), followed by a downward trend through to the present. The 1950's are not completely consistent for all stations. some showing rather indeterminate variations (e.g. Nottingham and Resolution islands). Others, e.g. Frobisher Bay and Clyde, show strong downward trends amounting to 0.8°C and 0.6°C respectively. All stations show a very consistent trend towards cooler summers throughout the 1960's amounting to 2.1°C at Clyde and Longstaff Bluff based on linear regressions for the period (see Bradley and Miller, 1972 for discussion). On the whole, mean summer temperatures at the close of the 1960's were cooler than they have been for 30-40 years.

Although most of the data begin in the early 1930's, a few ablation season records are available prior to this for Lake Harbour (Figure 11-3). Summer temperatures are available for the middle and late twenties and one complete record for the summer of 1914. This suggests a period of falling temperatures from the mid-twenties or earlier to the end of the twenties amounting to 0.7°C or more, followed by increasing temperatures until the early 1930's (1.4°C or more) when the cooling period already discussed set in. The 1914 summer was 2.2°C cooler than average temperatures at the station 1930-44. This information alone could be interpreted as simply an anomalously cold summer (although the departure below the mean temperature for 1930-44 amounts to 2.5 standard deviations). However, it is interesting to note that the subsequent Julys of 1915 and 1916 (the only summer month at this time for which data are available) experienced cooler temperatures (5.4°C and 6.1°C respectively) than the whole of the rest of the record with the exception of July 1926 (6.4°C) and July 1927 (6.1°C). This suggests that temperatures around 1914 were 1.5°C to 2.5°C cooler than in the late 1930's which are in turn approximately the same as today or slightly cooler. This is supported of course by the well-documented evidence of a global warming since the 1880's (Callendar, 1950; Willett, 1950; Longley, 1953; Putnins & Schallert, 1959; Mitchell, 1961, 1963). If this cooler period prior to 1920 was widespread throughout Baffin as one might expect then the actual ablation season or period of melting would have been considerably shorter than it is at present with elevations over 600 m possibly experiencing mean summer temperatures of < 0°C.

<u>Ablation Season Precipitation (Total Amounts in Water Equivalent)</u>:

Precipitation data are far less easy to interpret on a regional scale than temperature data. This is evidenced by the mean annual precipitation amounts for Cape Dyer (52.3 cm) and for Broughton Island (24.4 cm)(Table 11-1), only 150 km to the north (values for 1959-69). Thus a spatially coherent pattern is very difficult to perceive. Lake Harbour, for example, shows decreasing precipitation in the 1930's while Pangnirtung shows an upward trend. The records for Clyde and Frobisher Bay in the 50's and 60's show an upward trend in precipitation which would seem to fit with a falling temperature trend at these stations for the same period. In the 1960's some stations show increasing precipitation; however, this is not true in all areas. Patterns of temperature and precipitation fluctuations during the ablation season are summarized in Table 11-3.

Accumulation Season Temperatures: (See figures 11-4 and 11-5).

Accumulation season temperatures are characterized by very large fluctuations over the last 40-50 years at a number of stations. From the one record at Lake Harbour the mid to late twenties appear to have experienced a cooling trend followed by a rapid rise to the start of the 1930's. Marked cooling occurred in the first few years of the 1930's followed by equally marked warming to 1940-41; mean winter temperatures at Lake Harbour fell 1.8°C in the early 1930's and rose 3.6°C by 1940-41. Mean temperatures also fell by 1.0°C at Pangnirtung in the early 1930's and at Nottingham Island by at least 0.8°C. Mean temperatures 1934-40 increased 2.2°C at Nottingham Island, 1.3°C at Resolution Island. For the next few years the overall pattern is indeterminate (i.e. regionally incoherent) while in the 1950's most stations show a cooling trend (1.7°C at Arctic Bay, 0.8°C at Clyde). All stations except Cape Dyer then show marked warming in the 1960's which have generally resulted in mean winter temperatures at the end of the last decade somewhat warmer than for the last 30 or 40 years.

Accumulation Season Precipitation (Total Amounts in Water Equivalent):

As in the case of ablation season precipitation accumulation seasons show wide variations over time from station to station. There is some coherence in the record post-1954 which suggests that an upward trend began and has continued to the present. Resolution Island, however, shows a very marked downward trend through the record (11.4 cm mean decrease per year). For the 1930's, Lake Harbour shows a downward trend while Pangnirtung shows increasing precipitation, similar to the ablation season records at each station. During the 1960's, all stations except Frobisher Bay show increasing precipitation, amounting to over 100% of the mean 1960-69 at Cape Hooper. However, a mean figure of 34% for the regional increase during this period is probably more accurate (Table 11-4). Fluctuations of precipitation and temperature for accumulation seasons are summarized in Table 11-5. The spatial coherence of the precipitation trend over the last decade or two is particularly remarkable in view of the problems associated with recording snowfall accurately, amounts generally being underestimated.

Thus it is clear that in the last decade an extremely significant climatic fluctuation has occurred. Winter temperatures are now warmer than for the past 30-40 years and recent years have seen very large increases in precipitation. Summer temperatures on the other hand have fallen to a level probably cooler than for the previous 30 to 40 years. The net effect is hence for more snow to be deposited on the island and for less to be removed, resulting in notably increased glacierization of the region (Bradley and Miller, 1972).

Field Observations:

Studies of certain glaciological features in the northern Cumberland Peninsula, Baffin Island, and comparisons with aerial photographs taken of the area in 1949 and 1961 support the climatological arguments for a significant climatic change over the last decade. Permanent snowbeds and incipient glaciers respond essentially concurrently with climate, and these responses are readily observable in the field. The duration of ice-cover on small lakes also contains climatic information, but may be affected by several meteorological factors.

Several dozen snowbanks were observed late in the 1969-1971 field seasons and compared with their respective sizes on aerial photographs taken late in the ablation seasons of 1949 and 1961. Snowbanks generally decreased in area between 1949 and 1961, whereas all observed snowbanks increased markedly during the past decade. The actual increases are variable; permanent snowbanks now exist in many areas that were snow-free in 1961 and many of the smaller (<10 m diameter) snowbanks have nearly doubled in size whereas larger (>50 m diameter) snowbanks have changed less noticeably. In one instance, snowbanks now occupying an area snow-free in 1961 overlie 25 mm diameter thalli of the lichen species *Allectoria minuscula*. A growth curve developed for this species relating thallus diameter to substrate age indicates a relatively snow-free substrate for the previous 40 ± 10 years, and implies an amplitude of climatic fluctuation not experienced for at least that length of time. In contrast to snowbank growth, glacier termini had receded between 1949 and 1961 and continued active recession through 1971 due to their much longer response time to climate. Most detailed snow and ice studies in the Arctic have been carried out on large glaciers or ice-caps whose response lags at least 50-100 years behind any climatic shift. Current climatic conditions might be better documented by more detailed studies of snowbanks or related features.

Another, albeit less frequently observed result of climatic change is the formation of incipient glaciers. In 1970, two cirques which were snow-free in 1961, were noted as being presently occupied by glacierets or large snowbanks.

Supporting evidence for a recent climatic deterioration is provided by data on the duration of ice on small lakes (<1 km^2) in the area. A number of lakes ice-free in 1961 were observed in recent years to remain ice covered throughout the the ablation season; others, also ice-free in 1961, have only partially melted during recent ablation seasons retaining some ice throughout the summer. On these lakes, ice accumulation from several years was noted. On all lakes observed, the percentage of ice-cover at the end of the ablation season was equal to or greater than ice-cover in 1961. Such evidence alone does not imply a general shift towards more glacial conditions but supports other field evidence previously discussed.

If, as the climatic records imply, the present climatic deterioration is being felt throughout Baffin island, a brief comment on the present distribution of ice-bodies is worthwhile. Much of north-central Baffin Island is a rolling upland plateau with elevations of 400 - 600 m a.s.l., 2% of which is glacierized at present. In the recent past (330 ± 75 years BP), however, 70% of the area was glacierized (Ives, 1962). The glaciation limit for this area is estimated by Andrews and Miller (1972) to be 700 - 800 m a.s.l., as inferred from maps based on aerial photographs taken in 1958 and 1961. Summer insolation, the elevation of the July 0°C isotherm and winter precipitation are the major controls on the elevation of the limit. With the exception of Frobisher Bay, all weather station sites show increases in winter precipitation during the last decade and a mean summer cooling of 1.3°C. Assuming a lapse rate of 0.75°C/100 m (based on various field data), and taking into account the observed increases in winter precipitation, the present glaciation limit is theoretically at least 200 m below its 1958-1961 value, which puts it at or below a significant proportion of the land surface. If present trends continue, therefore, this will lead to more extensive glacierization of the region. An increase in the number and size of thin ice-caps on the plateau region, and accompanying increases in the mean albedo of the island during the ablation season would give positive feedback and further reinforce the climatic trends already in progress

ANALYSIS OF SYNOPTIC TYPES FOR BAFFIN ISLAND

In the light of the previous discussion, changes in the frequency of certain synoptic types over Baffin Island were examined for the last decade. A catalogue of synoptic types for the eastern part of the Canadian Arctic Archipelago has been developed by Barry (1972) for a number of months during the year. The classification is based on principal 'static' features of upper air circulation but incorporates the dominant air flow over the region where appropriate. Classifying the types into two groups - those situations where the dominant circulation control was cyclonic and those situations where it was anticyclonic - the frequency of occurrences in each group was calculated for the period 1961-65 and 1966-70 (Table 11-6). The frequency of anticyclonically controlled situations increased between the periods at the expense of the cyclonically controlled situations. In view of the downward trends of temperature in these months, noted above, it is at first rather surprising to find that anticyclonic situations have increased. However, the answer lies in airflow associated with each type. The classification scheme of Barry covers a much larger area than Baffin Island alone and thus an anticyclonically controlled synoptic type may in fact bring cold northerly airflow over most of Baffin Island (if, for example, a ridge is present to the west of Baffin Island). Thus a further breakdown of the types on an airflow basis was made (Table 11-7) dividing those types with a northerly and southerly component, and those types with an easterly and westerly component. The analysis is crude but makes the best use of available data. Airflow was defined as "gross airflow over central and eastern Baffin"and taken from the ideal charts for each type according to Barry (1972) (see Figure 3-2).

Table 11-7 shows that the frequency of days with airflow having a westerly component, particularly a southwesterly component, has decreased by approximately 29% between the two periods. A concurrent increase in the frequency of easterly particularly northeasterly flow is also apparent. An analysis of synoptic patterns typical of cool summer days using temperature data from eastern Baffin (Andrews, Barry and Drapier, 1970) indicated that all days were associated with an easterly flow component over the area. A similar analysis of warm summer conditions revealed southwesterly flow was dominant. Thus it is clear that the cooling trend over the last decade is related to a higher frequency of cold air being advected into the area from the east and northeast and a lower frequency of west and southwesterly flow advecting relatively warm air into the region. Further analysis of climatic parameters at a number of stations associated with each synoptic type is presently underway and will no doubt shed more light on their problem

CONCLUSIONS

It is clear from the foregoing discussion that Baffin Island has recently experienced some very large seasonal climatic fluctuations which on the whole do not support evidence elsewhere of an overall cooling trend post-1940. Changes in circulation are the cause of the most recent climatic fluctuatinns. Conditions in recent years have illustrated how sensitive the area is in terms of the build-up of snow and ice and the growth of glacieriets (Bradley and Miller, 1972). Geomorphological evidence shows that whereas only 2% of the area is presently glacierized, in the recent past (330 years ± 75 BP), 70% of the region was glacierized (Ives, 1962; Falconer, 1962). This latter pattern would be reestablished if the present climatic fluctuations persist.

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Table 11-1

Meteorological Stations Referred to in the Text

Station	Latitude	Longitude	Elevation (m)	°C Mean Temperatures (1959-69)		Total Precipitation (cm/we) (1959-69)	
				Winter	Summer	Winter	Summer
1. Arctic Bay	73°00'	85°18'	11	-20.0*	4.4*	7.7*	4.9*
2. Broughton Island	67°33'	64°03'	581	-15.5	2.6	24.4	6.6
3. Cape Dyer "A"	66°35'	61°37'	376	-14.3	3.6	52.3	13.7
4. Cape Hooper	68°26'	66°47'	401	-16.0	2.3	18.5	7.1
5. Clyde	70°27'	68°33'	3	-16.9	3.0	13.2	7.1
6. Dewar Lakes (mid-Baffin Is.)	68°39'	71°10'	518	-18.2	3.2	10.9	9.4
7. Frobisher Bay "A"	63°45'	68°33'	21	-14.1	5.9	26.9	15.5
8. Lake Harbour	62°50'	69°55'	16	-12.3+	6.8+	22.8+	11.5+
9. Longstaff Bluff (Foley)	58°57'	76°18'	162	-17.9	4.7	10.9	7.9
10. Nottingham Island	63°07'	77°56'	16	-13.2	4.5	17.0	7.9
11. Pangnirtung	66°09'	65°30'	113	-14.1+	6.1+	25.7+	13.0+
12. Resolution Island	61°18'	64°53'	39	- 8.8*	2.5*	18.5*	10.6*

* Values of mean temperature and total precipitation for 1951-60.

\+ Values of mean temperature and total precipitation for 1931-40.

Table 11-2

Correlation Coefficients Between Baffin Island Stations 1959-69

(a) Accumulation Season Temperatures

	1.	2.	3.	4.	5.	6.	7.	8.
1. Broughton Island	1.00							
2. Cape Dyer	0.83	1.00						
3. Cape Hooper	0.93	0.73	1.00					
4. Clyde	0.82	0.50*	0.81	1.00				
5. Dewar Lakes	0.95	0.82	0.85	0.73	1.00			
6. Longstaff Bluff	0.79	0.64	0.83	0.80	0.69	1.00		
7. Frobisher Bay	0.88	0.86	0.72	0.57*	0.90	0.69	1.00	
8. Nottingham Is.	0.78	0.69	0.53*	0.51*	0.78	0.51*	0.92	1.00

(b) Ablation Season Temperatures

	1.	2.	3.	4.	5.	6.	7.	8.
1. Broughton Island	1.00							
2. Cape Dyer	0.77	1.00						
3. Cape Hooper	0.76	0.67	1.00					
4. Clyde	0.72	0.71	0.71	1.00				
5. Dewar Lakes	0.88	0.83	0.81	0.84	1.00			
6. Longstaff Bluff	0.82	0.82	0.69	0.83	0.96	1.00		
7. Frobisher Bay	0.28*	0.33*	0.50*	0.34*	0.51*	0.43*	1.00	
8. Nottingham Is.	0.61	0.80	0.67	0.71	0.86	0.72	0.86	1.00

* Correlations not significant at the 5% level or greater. 1% significance level = 0.74. 5% significance level = 0.60.

Table 11-3

Summary of Ablation Seasons

Temperature:	1914 - early 1920's:	Marked warming
	1932 (or earlier)to approx. 1943:	Cooling
	1943 to ca. 1950:	Warming
	1950 to present:	General cooling particularly accentuated in the last 10 years
Precipitation:	1930's	Indeterminate
	1940's to early 1950's:	Downward (?)
	1960's to present:	Upward trend for some stations

Table 11-4

NET CHANGES IN SEASONAL MEAN TEMPERATURES AND SEASONAL PRECIPITATION TOTALS, 1960-1969, DERIVED FROM LINEAR REGRESSIONS

				Ablation Season (J,J,A)			Accumulation Season (Sept-May)		
Station	Latitude	Longitude	(meters) Elevation	Temp.(°C)	Precip.(cm/we)	%[1]	Temp.(°C)	Precip.(cm/we)	%[1]
1. Broughton Is.	67°33'	64°03'	581	-0.9	-0.2	- 3.4	+0.7	+12.6	+51.7
2. Cape Dyer	66°35'	61°37'	376	-2.0	-6.9	-49.7	-0.4	+11.8	+21.8
3. Cape Hooper	68°26'	66°47'	401	-0.5	+5.8	+82.5	+0.4	+23.4	+127.4
4. Clyde	70°27'	68°33'	3	-2.1	-0.2	- 2.2	+1.6	+ 2.2	+16.6
5. Dewar Lakes	68°39'	71°10'	518	-1.4	+0.2	+ 1.9	+0.9	+ 2.5	+23.3
6. Frobisher Bay	63°45'	68°33'	21	+0.03	-0.2	- 1.3	+1.2	- 2.6	-10.0
7. Longstaff Bluff	68°57'	76°18'	162	-2.1	-2.1	-24.9	+0.7	+ 0.7	+ 5.9
8. Nottingham Is.	63°07'	77°56'	16	-1.7	+3.7	+54.4	+2.0	+ 5.8	+32.9
Average				-1.3	+0.5	+ 7.2	+0.9	+ 7.1	+33.7

1 Values are for net precipitation changes, 1960-1969, as a percentage of the means 1960-1969

Table 11-5

Summary of Accumulation Seasons

Temperature:	Late 1920's	Cooling then warming to ca. 1930(?)
	1932-34	Rapid cooling
	1934-40	Marked warming
	1940-48	Warming (?)
	1948-60	Cooling
	1960-70	Marked warming
Precipitation:	1930's and 1940's	Incoherent and indeterminate
	Early 1950's to present	Upward trend accentuated in the last 10 years

Table 11-6

Frequency of Synoptic Types for Baffin Island - July and August (Barry, 1972)

"Cyclonic Control"

Types	Definition	Total of Occurrences	Total of Occurrences	Difference
A. 100,101,102,110,130	Central low/trough	59	38	-21
B. 500,510,542	Davis St. low	35	23	-12
C. 200,210,520,532,540,720	Baffin Bay low	14	15	+1
D. 600,610	low to SW	33	30	-3
E. 612,620,630,640	Low to SW or S with other lows	24	33	+9
	Total	165	139	-26

"Anticyclonic Control"

Types	Definition	Total of Occurrences	Total of Occurrences	Difference
F. 900,910,920,930	Anticyclone	19	43	+24
G. 120,800,810	Ridge Situations	24	28	+4
H. 300,310,320	Ridge, low to S	25	39	+14
J. 400,410,420,430	High in E, low to W	25	36	+11
K. 511,521	Ridge; Baffin Bay low (NE flow)	14	9	-5
L. 320,531,701	Ridge; Baffin Bay low (N,N,W flow)	38	16	-22
	Total	145	171	+26

Table 11-7

Frequency of "Airflow Types" (after Barry, 1972) for Central and Eastern Baffin, 1960-69

Southerly Airflow with Westerly Component

Type	1961-65	1966-70	Difference
100	19	13	
102	5	0	
130	16	6	
420	7	8	
430	9	6	
620	6	5	
640	1	3	
	63	41	-22

Northerly Airflow with Westerly Component

Type	1961-65	1966-70	Difference
110	11	7	
120	9	6	
200	1	1	
210	3	1	
521	7	5	
531	18	4	
532	1	3	
540	2	3	
701	5	5	
810	5	8	
930	2	7	
	64	50	-14

Southerly Airflow with Easterly Component

Type	1961-65	1966-70	Difference
320	11	18	
410	8	20	
500	22	10	
600	23	15	
610	10	15	
920	4	13	
	78	91	+13

Northerly Airflow with Easterly Component

Type	1961-65	1966-70	Difference
300	8	5	
310	6	16	
510	8	8	
542	5	5	
630	13	18	
702	5	7	
900	7	17	
	52	76	+24

Total Change Westerly Airflow -36 days (-29%)
Easterly Airflow +36 days (+29%)

FIGURE 11-1: Location of meteorological stations referred to in the text. Numbers refer to stations listed in Table

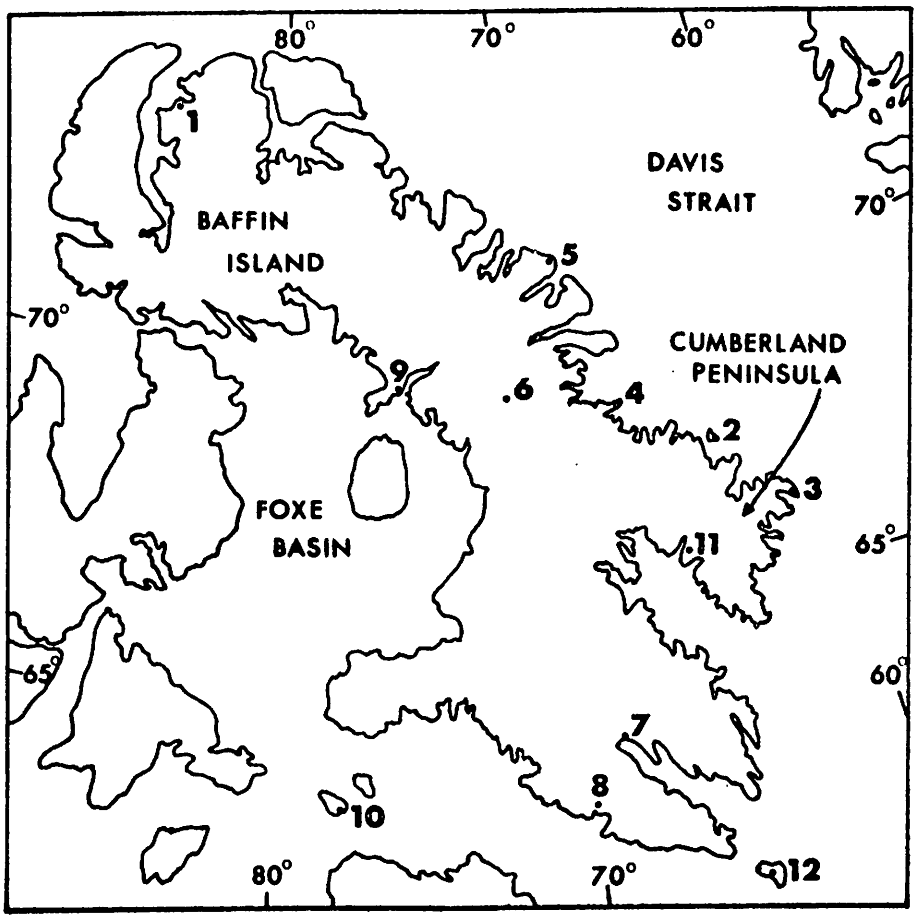


FIGURE 11-2: Ablation season mean temperatures: five year weighted binomial running means.

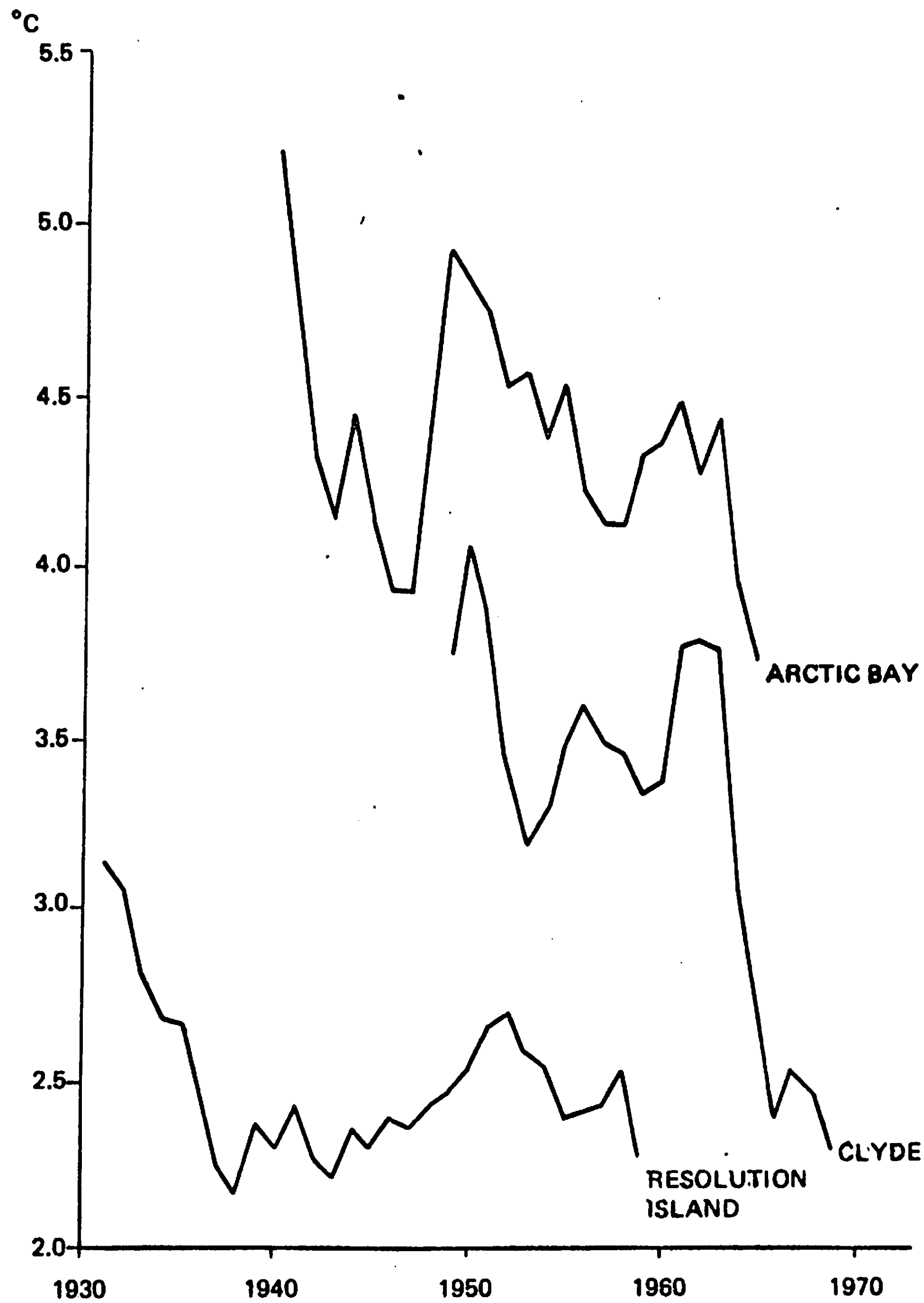


FIGURE 11-3: Ablation season mean temperatures: Three* and five+ year weighted binomial running means.

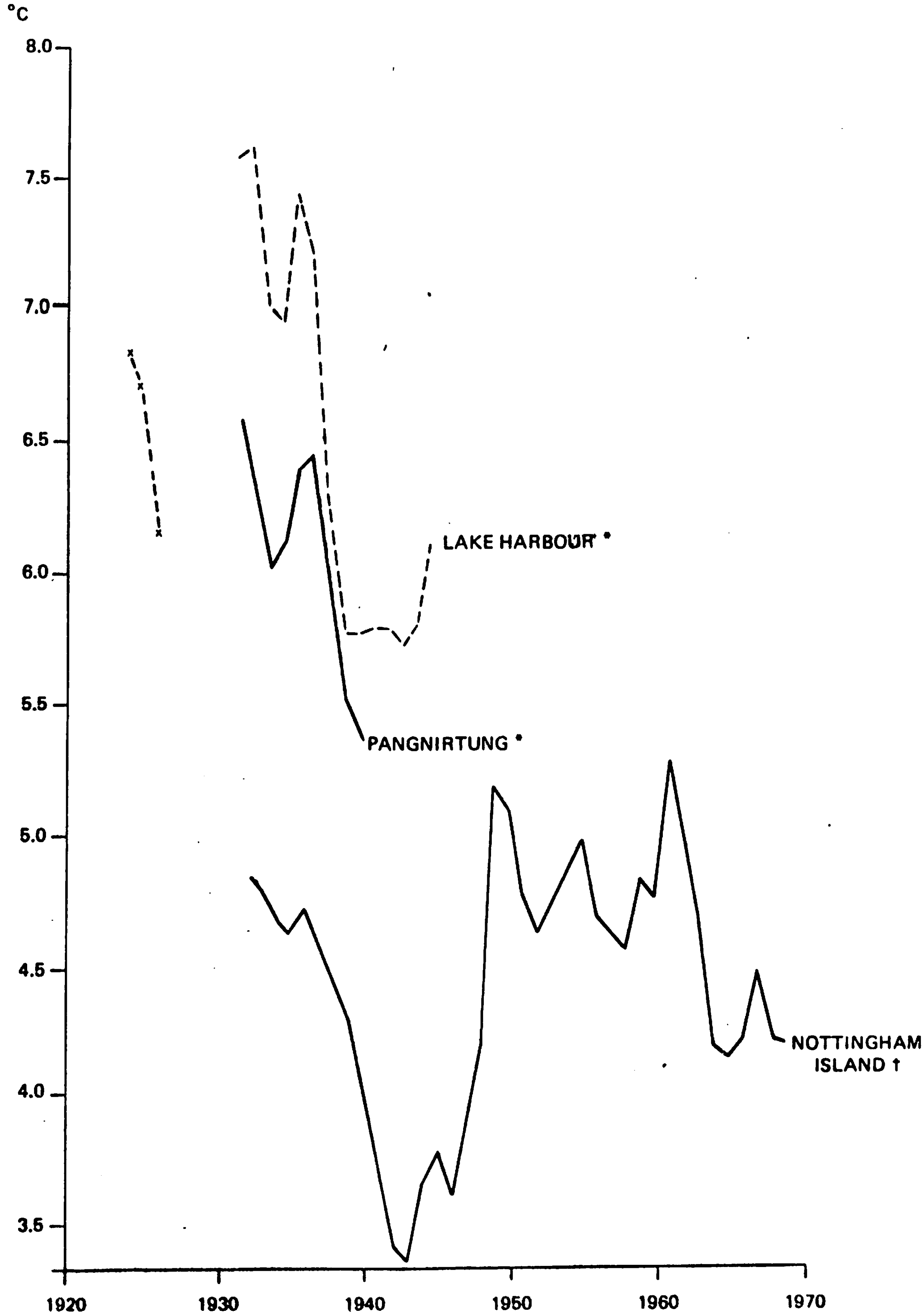


FIGURE 11-4: Accumulation season mean temperatures: three* and five+ year weighted binomial running means.

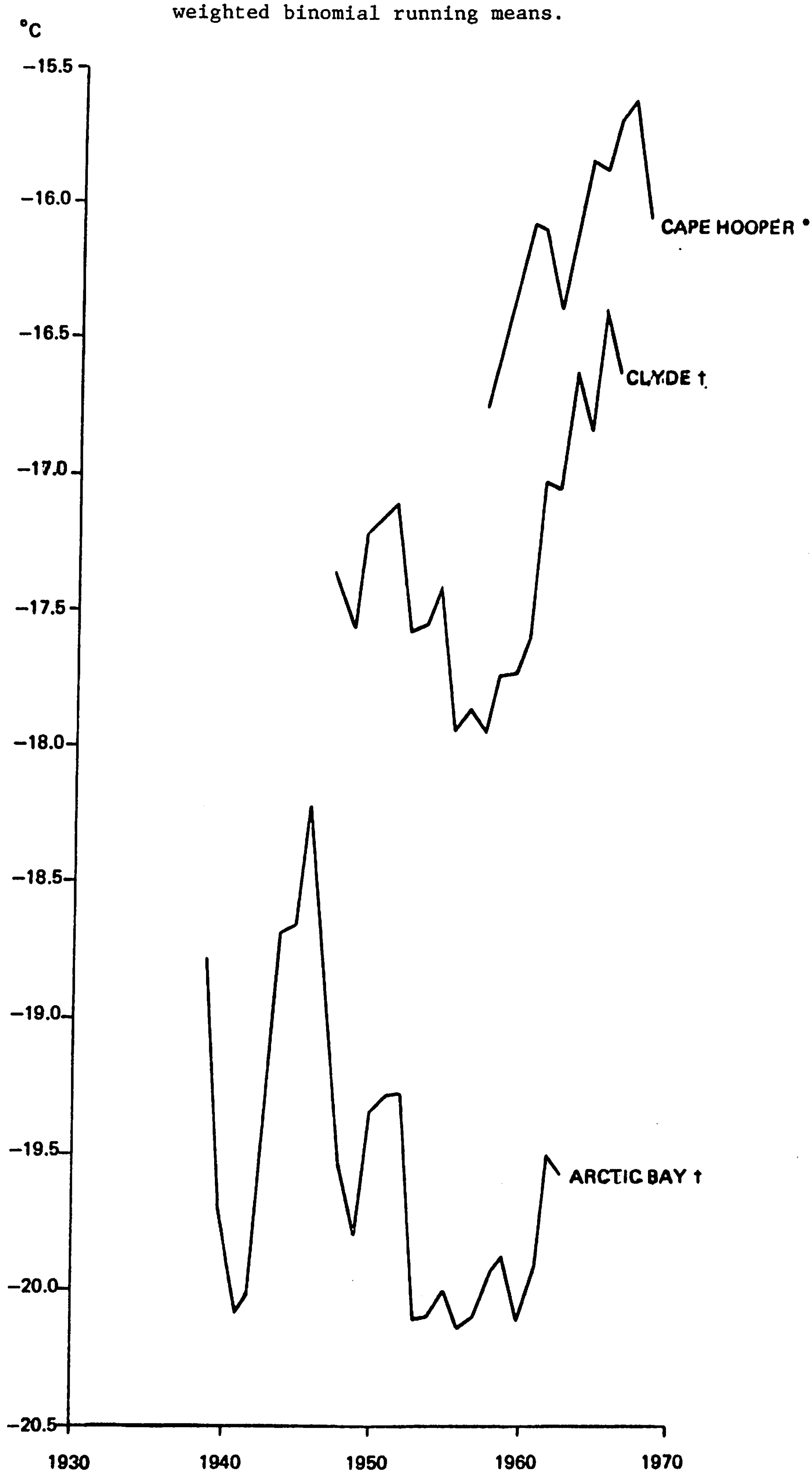
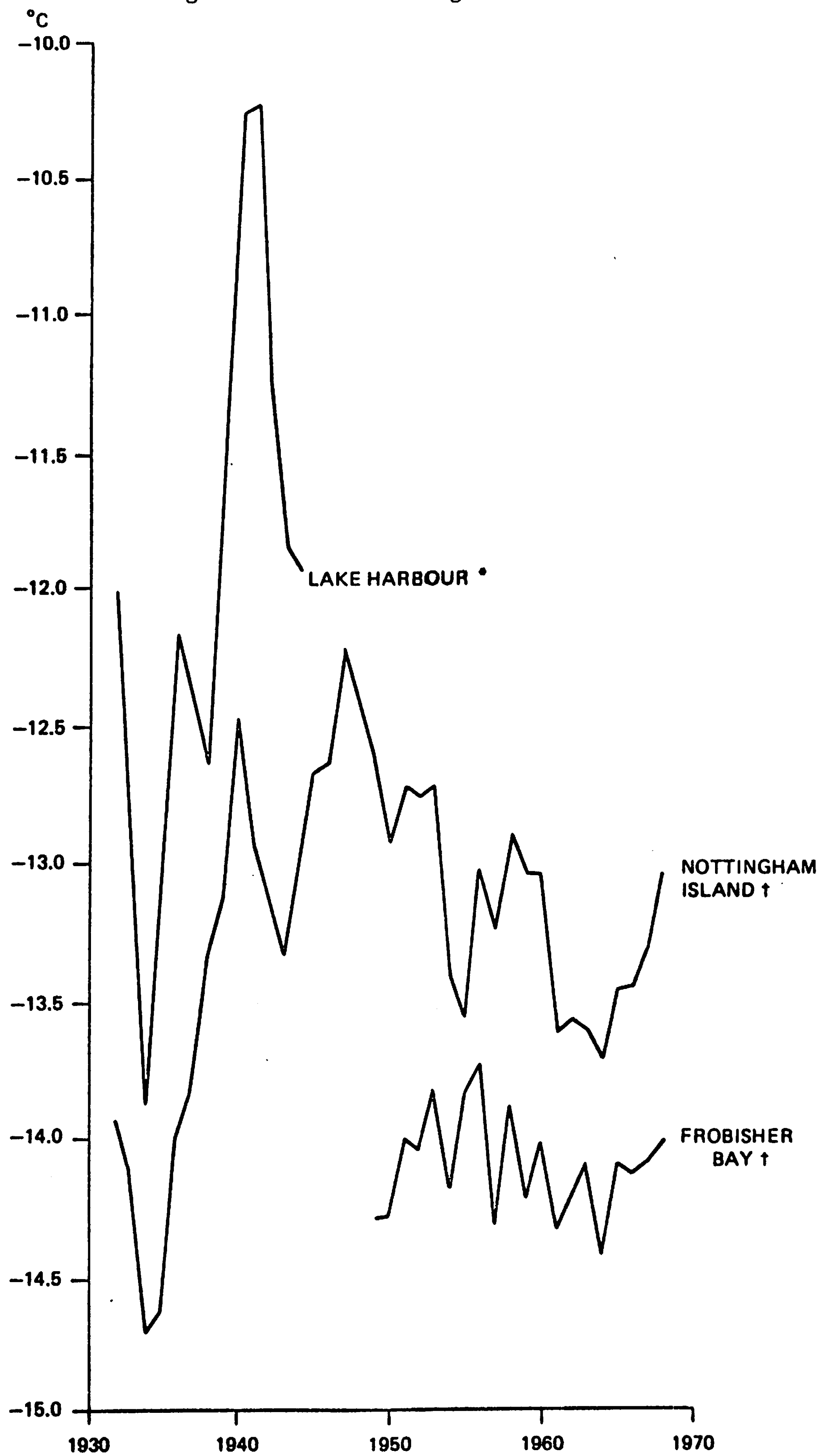


FIGURE 11-5: Accumulation season mean temperatures: three* and five+ year weighted binomial running means.



12. THE RECONSTRUCTION OF PAST CLIMATIC CONDITIONS IN BAFFIN ISLAND

J.T. Andrews and R.G. Barry

Introduction

This report has partly documented the various spatial and temporal scale data relating to climatic-glaciological interactions defined in Table 1-1. There remains the question of the links between the different scale processes - i.e., between the micro-, topo- and macroscale climatology - as well as that of the inferred climatic sequence that could have produced the reconstructed pattern of glacierization. It is also appropriate to outline problems worthy of future study.

This chapter provides: 1) a climatological analysis of conditions favoring greater or lesser glacierization, based on current synoptic weather patterns; 2) a study of the relationship between glacier mass balance and winter and summer weather over the last decade; and 3) a broad sketch of the glacial sequence and the inferred climatic changes that led to the glacier responses.

SYNOPTIC ASPECTS OF CONDITIONS FAVORING GLACIERIZATION

The climatological interpretation of the reconstructed glacial history of the area has several facets. One is the question of radiation and elevation relationships of ice-free and presently-occupied cirques discussed by Williams (Chapter 10). Here, we are concerned with the more speculative interpretation of possible conditions favoring greater or lesser glacierization in the area than at present. The findings of Bradley and Miller (1972) confirm the suspected sensitivity of this area to small climatic shifts and their results lend more confidence to hypothetical arguments based on various synoptic climatological analyses.

It is widely held that the onset of the Pleistocene glaciations followed a decrease in summer temperature of up to 5° - 6°C, producing a shorter ablation season and causing a greater proportion of the annual precipitation to fall as snow. (Precipitation totals will also decrease as a result of the lower moisture content of the colder air). Applying this concept to present conditions, we find that Broughton Island receives 79% of July-August precipitation as rain with a mean temperature of 4.4°C and Cape Dyer 73% of June-August precipitation as rain with a mean temperature of 3.9°C. Resolute (73°N) receives 64% of June-August precipitation as rain with a mean temperature of 2.8°C. We may estimate, therefore, that a decrease of 1.5°C in summer temperature would produce a decrease of about 15% in the precipitation falling as rain for Baffin Island. A decrease of 3°C (to a mean of 1°C) should result in about half the summer precipitation occurring as snow which accords with other findings (Lamb, 1955) indicating that the probability of snow (or rain) is about 50% at 1.2°C or alternatively when the freezing level is at 250 m (Barry, 1967 unpub.). For the northern Cumberland Peninsula this change amounts to about 3 cm H_2O and by comparison with winter accumulation is insignificant for glacier mass balances.

We can argue, therefore, that modest changes in summer conditions are mainly important in terms of the amount of ablation. This will be determined primarily by the degree of cloudiness and the cloud type. As far as cirque glaciation is concerned, large amounts of any medium or low cloud will reduce south slope/north slope contrasts. However, as Jacobs has shown, the relationship between net radiation (which accounted for 60% of the energy source for ablation in 1970) and cloud cover is complex.

Using Jacobs' data, we can determine that between 11 and 16 days are required to warm the snowpack (at 1200 m) to 0°C, assuming that net radiation is 60% of the energy source, although this warming process is proceeding at the same time as

	Number of days needed of each radiation-day type				
	Energy(ly)	A(10L)	B(0-2)	E(6-10H)	C-D(6-10M)
Warm Snowpack to 0°C	1100	16	14	13	11
Melt 0.15 m H_2O snowpack	1200	15	13	12	10
Evaporation/sublimation	3650	45	3	35	31
Total		76	65	60	52

sublimation. The time required to remove the 1969-70 winter snowpack (~0.4 m H_2O) can be estimated as between 121 days (for C-D type) and 174 days (for A type) if the melting and evaporation/sublimation terms are also increased proportionally. For the 1970-71 winter snowpack (~0.25 m H_2O) these estimates become 70 and 100 days, respectively. This accords with the observed removal of the snow in the 1971 summer, assuming a starting date of 1 June and frequent C-D type days (i.e. 6-10/10 medium cloud cover).

Unfortunately, the radiation-day types do not correspond in any simple manner with the synoptic circulation types (Barry, 1972), at least on the basis of the limited data collected in summer 1970. Anticyclonic and ridge situations have a mean net radiation in summer of 60 cal $cm^{-2}day^{-1}$ which is very close to that for types with lows in Davis Strait (56 cal $cm^{-2}day^{-1}$). For central lows, or lows in Foxe Basin, the figure is 71 cal $cm^{-2}day^{-1}$, but in each case the range of values is considerable and the differences are not statistically significant. This situation prevents the formulation of simple glacial/non-glacial circulation patterns from this standpoint at least.

A further vital consideration in terms of ablation is the synoptic character of the month of June. In this respect 1970 and 1971 are very dissimilar. The cyclonic/anticyclonic frequencies are respectively, 19 and 11 days for 1970, 12 and 18 days for 1971. Moreover, there were 10 days with a low pressure over Baffin

Island in June 1970 compared with only 1 such day in June 1971. The figures for a high pressure cell were 3 and 8 days, respectively. Clearly, the early part of the two melt seasons established a markedly different regime. Interestingly, June 1969 was similar to June 1970 with 19 cyclonic days although the lows were mainly in northern Hudson Bay and the northern Labrador Sea in 1969 so that precipitation was only average, whereas in 1970 they were frequently centered over Baffin Island and precipitation was well above average (see Table 3.2).

These results suggest that the early summer period is a critical one for the ablation conditons. If frequent depressions maintain a snow cover the high albedo and cloud cover greatly slow the warming of the snow pack. This occurred in 1970. In 1971 when the snowpack of 0.25 m was entirely removed, the early start with an anticyclonic June must have helped greatly to offset the near-average cyclonic activity in July-August. Conversely, in 1969 a rather unfavorable start was reversed by an unusually anticyclonic mid-summer so that glacier ice was exposed over most of the surface in mid-August.

Temperature levels must also be taken into account. For example, the cool July of 1970 meant that 22% of the month's total precipitation fell as snow on the Boas Glacier, 39% at Cape Dyer, whereas in July 1971 the precipitation at Cape Dyer (255% of normal) all fell as rain.

It has been argued that in higher latitudes an increase in winter accumulation is a prerequisite for glacierization (Tanner, 1965; Loewe, 1971). This is supported by the models of Williams (Chapter 10) for cirque glacierization in the Cumberland Peninsula. In southern Baffin Island significant autumn and winter precipitation occurs with lows in the Davis Strait-Baffin Bay area or over Baffin Island (Barry, 1972) as indicated in Table 12.1. Such lows commonly move northward from the Labrador-Ungava and Labrador Sea areas during periods of low hemispheric index (Barry, 1966). Positive precipitation anomalies (and positive temperature departures) occur over Baffin Island in winter in association with a 700 mb mean trough displaced west of its normal position (Brinkmann and Barry, 1972) which gives rise to such depression tracks. In spite of the discrepancy in the Table between precipitation amounts at Broughton Island and Cape Dyer with two of the groups of Ridge situations, which probably requires a longer period of analysis to resolve, the general conclusion regarding the pressure fields likely to encourage glacierization is clear. The moisture source for eastern Baffin Island is the North Atlantic, as suggested earlier (Andrews, Barry and Drapier, 1970) from a more limited study. The role of the inverted low type, with its warm sector in the north affecting most of Baffin Island, is especially striking in view of its infrequent occurrence. More investigation of this synoptic pattern is desirable.

Deglaciation conditions are more difficult to specify since they derive from a less certain combination of low winter accumulation and strong summer ablation. At least we can suggest from the earlier analysis that the occurrence of high insolation receipts and of little or no snowfall in June are likely to be highly significant. Low winter accumulation will occur during winters with a 700 mb trough near its normal location or east of it with a paucity of depressions from the Labrador Sea area.

Table 12.1 SYNOPTIC CLIMATOLOGY OF PRECIPITATION AT BROUGHTON ISLAND AND CAPE DYER. JAN.-FEB. AND SEP.-OCT. 1961-65.

	JAN.-FEB.		SEP.-OCT.	
Description	Broughton Is.	Cape Dyer	Broughton Is.	Cape Dyer
Inverted Davis St. low	31.4%	14.6%		
Central low/trough	18.6	34.5	7.7%	24.4%
Davis Strait low	13.8	6.1	12.5	31.0
Baffin Bay low	1.3	17.8	17.4	16.8
Low to SW	0	2.0	2.9	1.8
Low to SW or S with other lows	15.8	4.5	8.7	6.1
Cyclonic control	80.9	79.5	49.2	80.1
Anticyclone	5.6	5.5	8.3	5.7
Ridge situations	8.6	4.3	11.7	1.5
Ridge, low to S	0	1.8	6.3	1.2
High in E, low to W	0	3.4	3.8	2.4
Ridge, Baffin Bay low (NE flow)	1.5	0.1	2.3	1.5
Ridge, Baffin Bay low (N-NW flow)	3.5	4.4	18.1	7.6
Anticyclonic control	19.2	19.5	50.5	19.9
Total precipitation 1961-1965	14.0 cm	75.7 cm	55.5 cm	74.6 cm

PREDICTION OF BOAS GLACIER MASS BALANCE BASED ON BROUGHTON ISLAND CLIMATOLOGICAL RECORDS

There have been several attempts to relate glacier mass balance on some specific ice body to climatological records at nearby weather stations (e.g., Hattersley-Smith, 1963; Koerner, 1970) with conflicting opinions as the usefulness of such relationships. Part of the problem may lie in the elevation of most arctic weather stations which are located close to sea level and thus below the regional inversion (Bradley, 1972). In contrast, the DEW line weather stations are located on hilltops above the inversion and may, therefore, more accurately depict weather on nearby glaciers.

The Broughton Island DEW line site is at an elevation of 581 m a.s.l. and is 65 km ESE from the Boas Glacier. Correlations (J. Jacobs) between the DEW line site (x) and the Base Camp (y) below the glacier at 490 m a.s.l. for 18 days in August, 1970 are:

$$\begin{aligned} &T_{mx} \qquad y = 2.85 + 1.01x; \quad r = 0.94 \\ &T_{mn} \qquad y = 6.22 + 0.88x; \quad r = 0.89 \end{aligned} \tag{12-1}$$

The higher potential temperature at Base Camp was noted in Chapter 3. The relationship between Base Camp and Glacier Camp between June 8th and August 15th, 1970 was:

$$\begin{aligned} &T_{mx} \qquad y = 12.4 + 0.65x; \quad r = 0.78 \\ &T_{mn} \qquad y = -0.47 + 0.86x; \quad r = 0.80 \end{aligned} \tag{12-2}$$

The high intercept for T_{mx} is probably related to screen heating as discussed previously.

Bradley (pers. comm., 1972) investigated the relationship between winter temperature and precipitation and the succeeding temperatures of the summer (June - August) season and noted that there was a negative correlation between the amount of winter accumulation and the following summer temperatures. Insofar as a glacier's mass balance is a function of winter accumulation and summer ablation, this result was of considerable significance. Figure 12-1 graphs winter accumulation (x) as recorded at the Broughton Island DEW line site (winter = September - May) against summer (June - August) degree days, defined in this case as the number of degrees above a daily mean of 0°C (y). The regression relationship for the 11 year period is:

$$y = 438 - 7.1x; \quad r = 0.67 \tag{12-3}$$

where x is in cm H_2O.

The known specific mass balance data ($\bar{b}n$) for the Boas Glacier are also included on this diagram but are not sufficient in number to do more than suggest negative and positive mass balance years. Bradley (submitted) has shown, however, that on the basis of the climatological records of the last decade (1960-1969) that temperature and precipitation trends were largely in phase, and statistically associated, across most of Baffin Island (with the exception of Cape Dyer and Frobisher Bay). Accordingly, we have included on Figure 12-1 the specific mass balance results (+ or -) from the Decade Glacier (69°39'N, 69°55'W). This glacier is part of the Canadian I.H.D. program and has been studied since 1964. Mass balance data have been obtained through the courtesy of the Glaciological Subdivision, Inland Waters Directorate, Environment, Canada. On the basis of these data and our limited set from the Boas Glacier we estimate that the 1971-72 balance year will be strongly positive with an accumulation of 30 cm H_2O and degree days in the region of 100-200 (weather records not fully available).

The data are not of adequate length to construct a detailed predictive model but they are sufficient to illustrate two possible approaches. We can establish a predictive equation of the type:

$$\text{Specific mass balance} = k_o + k_1X + k_2Y$$

where X = winter precipitation and Y = summer degree days > 0°C. Alternatively, we can formulate a disciminant equation for the best separation of negative and positive mass balance years

$$D_o = k_1X + k_2Y \qquad (12\text{-}4)$$

with X and Y as above. Figure 12-1 shows 3 dividing lines - #1 based on the multiple correlation case with zero intercept; #2 based on the case with multiple regression intercept, and #3 based on the disciminant equation from the 6 years of mass balance records. This latter equation has the form:

$$1.71 = 0.0992X - 0.00338Y \qquad (12\text{-}5)$$

with X and Y as defined above. We have used the discriminant function to indicate the regions on the graph of positive, negative and approximate balance for $\bar{b}_n$. The latter case has been computed on the basis of the standard deviation about D_o (eq. 12-4).

The multiple regression approach to specific mass balance on the Decade Glacier, being a function of winter accumulation and summer warmth at Broughton Island, indicates excellent correlations; multiple r values are 0.825 and 0.852 for the regressions with and without the intercept, respectively. These results indicate that Bradley's conclusions (submitted) regarding the climatological homogeneity of Baffin Island as far as the gross similarity of winter and summer seasons is concerned, also appear to apply to the relationship between climate and glacier mass balance and ultimately glacier response to climatic change. Additional data from the Decade Glacier and the Boas Glacier should enable researchers to investigate further the relationships discussed in this preliminary survey. It should be noted that Figure 12-1 is concerned primarily with a relative ranking of the $\bar{b}_n$ data (into positive, zero and negative categories) rather than with the prediction of actual specific mass balance.[1]

GLACIAL SEQUENCE AND INFERRED CLIMATE AND CLIMATIC CHANGE

Figure 12-2 illustrates our concept of stratigraphic sequences for the course of the last glaciation (= Wisconsin in the broad glacial/climatic designation) in eastern Baffin Island and compares it with the time-transgressive events of the southern edge of the Laurentide Ice Sheet. The suggested trends of winter accumulation and summer temperature are shown on the right side of the diagram. As we have seen, evidence for the last decade indicates that there is a statistical relationship between the amount of winter precipitation and the succeeding summer degree days above 0°C.

Initial glacierization is assumed to be related to increased winter precipitation and a sympathetic decrease in summer temperature. Once eastern Baffin Island is glacierized, the locus of accumulation for the Laurentide Ice Sheet moves west and south to Labrador and Keewatin. Eventually the area of eastern Baffin Island becomes located on the northeastern margin of the Laurentide Ice Sheet under a climatological regime of low winter temperatures, little accumulation and mass loss primarily a function of incoming solar radiation. Under these conditions ice volumes are reduced from their early maximum extent. This phase occurs during the extensive mid- to late Wisconsin Glaciation recognized in southern areas of the ice sheet and would occur with sea level -80 to -130 m below present on a world-wide basis. On the basis of our observations on maximum glacio-isostatic depression of 70 m at the effective ice edge, it appears certain that large areas of the continental shelf off eastern Baffin Island were exposed during this long period. We suggest that it is during this period of low sea levels and radiation control on glacier distribution that low level corries on the outer coasts were glacierized. Many of these features are now drowned which adds

[1] Fig. 12-1 includes mean specific mass balances for the Devon Island ice cap and the Barnes Ice Cap and indicates that the Broughton Island discriminant is a reasonable approximation to the larger region. Sagar (1966) also estimated the mass balance of the Barnes Ice Cap based on the Clyde River meteorological data from 1950.

support to our concept that they were eroded during periods of world-wide, low sea level stands. Preliminary echo-sounding by G.H. Miller in 1972 has confirmed the existence of drowned deltas in fiords southwest of Padloping Island and further bathymetric work is planned as part of NSF grant GA-20883 to Andrews and Harrison.

The Cockburn stade, dated at 8,000 ± BP is an extremely well documented event along both sides of Baffin Bay/Davis Strait having been mapped extensively by Ten Brink (1971), although in West Greenland he distinguishes two major events at 8,800 and 8,300 BP. We propose that the Cockburn stade represents the response of the northeastern margin of the Laurentide Ice Sheet to an increase in winter accumulation as the atmosphere readjusted from its glacial to interglacial mode which included renewed cyclogenesis in Baffin Bay. There is some indication of such a shift on Lamb and Woodroffe's (1970) maps for July 11, 500 BP. It is worth noting that on the continental scale the deglacial history of the Laurentide Ice Sheet represents a shrinkage toward the northeast with the Barnes Ice Cap and possibly the Penny Ice Cap being relicts of the Laurentide Ice Sheet. The Cockburn stade was relatively short lived, perhaps because the increase in winter accumulation was not sufficiently offset, over a long time period, by the increase in water temperatures which is demonstrated by the faunas of raised marine deposits in the Canadian Arctic.

The suggested importance of the Milankovitch insolation theory of climatic change has been tested (Andrews, et al., 1972; Andrews (in press)) and is inadequate by itself to explain the glacierization of south-facing accumulation basins (Williams, 1972). There seems no doubt that the south-facing corries must have had higher mass turnovers than their north-facing counterparts and this is qualitatively confirmed by the larger (average) size of the empty south-facing basins. The south-facing corries become glacierized, therefore, during the early stages of regional glacierization when a combination of high winter accumulation and low summer temperatures allow mass gains in the southern basins. The main effect of low summer temperatures is to increase the total energy used in sublimation (i.e., the 1970 summer, Chapter 4). Whether south-facing basins would become ice-filled under the current precipitation regime is difficult to say as the evidence appears to indicate a decrease in snowfall up the northern slopes (Chapter 6). Accordingly, a southerly storm component might be a necessity to begin accumulation in the south-facing basins.

The results given in Chapters 10 and 11 indicate that: 1) the magnitude of the climatic change required to induce glacierization of the corries and ultimately the entire area is equivalent to a fall in the glaciation limit of 200 m, and 2) the climatic change of the last decade is of sufficient intensity to promote this glacierization if the change persists for some reasonable period (perhaps 30-50 years). Figure 12-3 illustrates the current relationship between summit elevation and glacier length in Okoa Bay (solid line) and the possible effect, assuming a linear relationship, of a shift downward of snowline by 200 m (broken line). The effect is to cause an increase in glacier length by a factor of 2.

CONCLUSIONS AND AREAS FOR FUTURE RESEARCH

The interdisciplinary teamwork carried out during this project has proved highly beneficial to all concerned and has laid the firm foundations for a quantitative palaeoclimatological - glaciological understanding of northern Cumberland Peninsula. The area is demonstrably very sensitive to climatic change on a time scale of $10\text{-}10^2$ years, although study of ice caps in the area is essential in order to confirm the wider applicability of this statement. Moreover, we consider eastern Baffin Island to be a lead area with respect to large scale glacierization (Andrews, et al., 1972). It appears critical, therefore, that attention be given to continued monitoring of environmental changes in the area. This should include the establishment of specific measurement sites relation to meteorological and glaciological parameters and the regular inventory of snow and ice cover by air photo and/or satellite coverage at the end of each summer. An evaluation of satellite potential[1] in this respect is intended with ERTS-B due to be launched, November 1973, but the necessary data collection should proceed without delay.

Several aspects of our work have independently shown the importance of winter accumulation on glacier mass balance. This important relationship requires wider examination in the Arctic on both glaciers and ice caps. Extending the analysis spatially will also help to compensate for the limitations of temporal studies at specific sites. The relationships between synoptic climatology, local weather and glacier mass balances are, not surprisingly, complex but even with a short record, certain major features have been identified. However, the radiation climatology of synoptic patterns is of key importance in this respect and this is currently under study by Barry and Jacobs. A specific question relating to glacier mass balance is the relationship between summer temperatures (and other simple indices) and the energy used for sublimation and melting.

Our findings strongly suggest that the climate during the last glaciation changed significantly in this area between the early phase and the middle and late phases. Verification of this in neighbouring areas is desirable. The

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[1] Related studies on arctic energy budgets are already being carried out by Barry and Jacobs under NSF (OPP) grants GV-28218 and GV-28220

relationship of drowned cirques to low stands of sea level during the Wisconsin deserves particular investigation.

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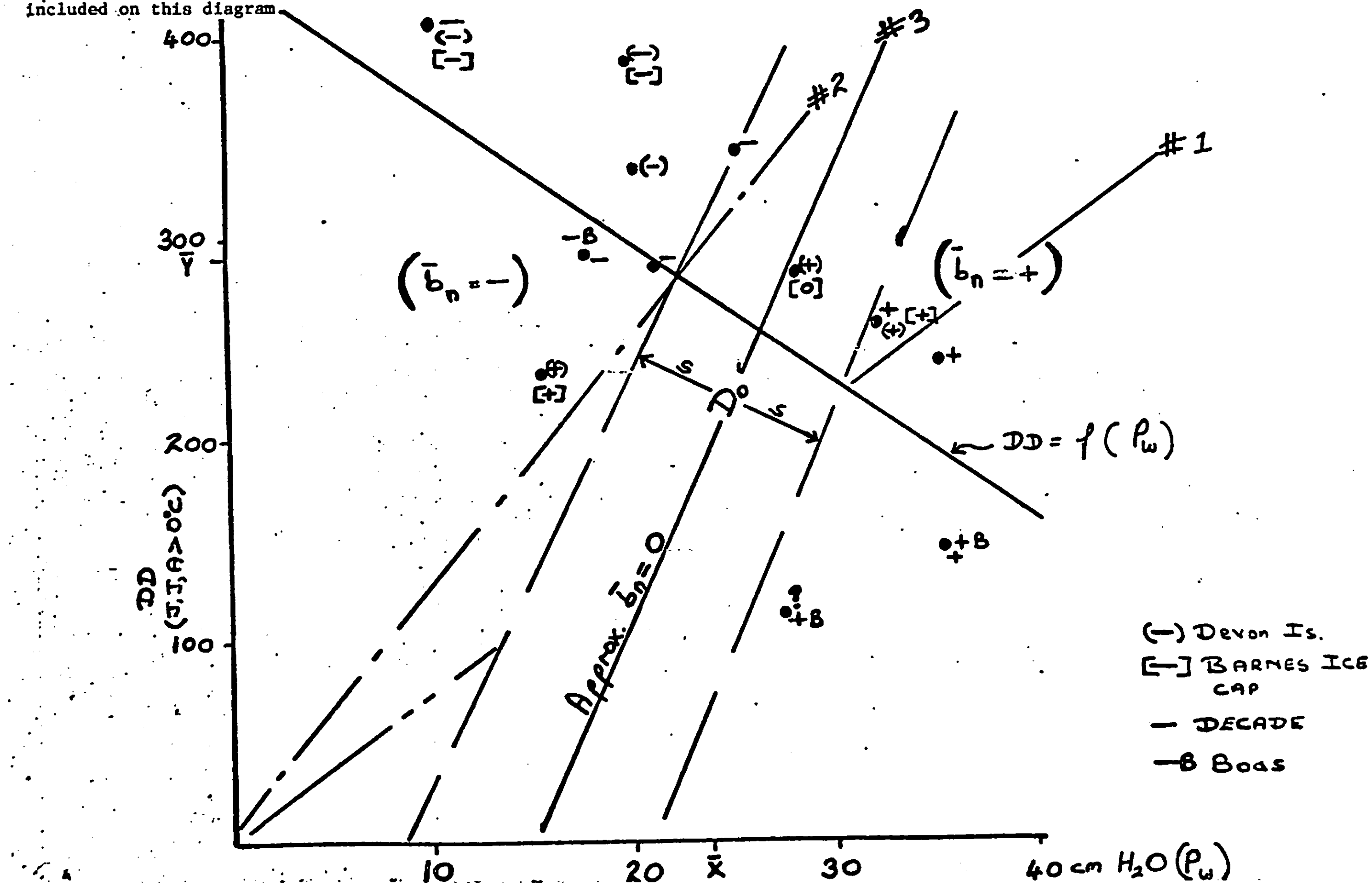
Figure 12-1: Graph of the relationship between winter precipitation on Broughton Island and summer degree days 0°C for the last 12 years (starting in Sept. 1960) and net specific mass balance on the Boas and Decade glaciers (+=positive and −=negative). Line #1 = the multiple regression line when $k_o = 0$; line #2 = multiple regression line for $k \neq 0$ and line #3 = discriminant between positive and negative specific balance years with lines showing + or − 1 standard deviation representing the approximate balance years. Note that mass balance data from the Devon Island Ice Cap (Koerner, 1970) and the Barnes Ice Cap (Sagar, 1966) are included on this diagram.



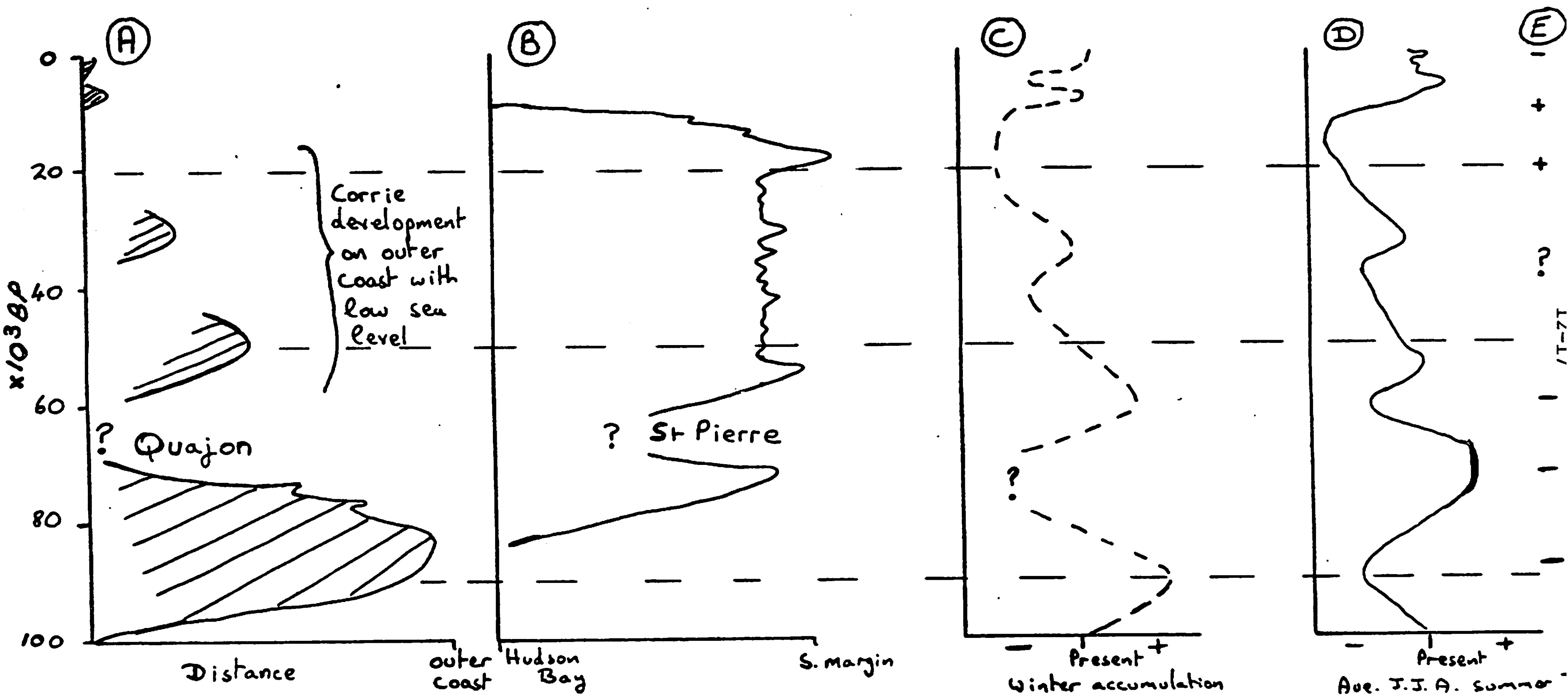


Figure 12-2: Schematic diagrams of:

A: time/distance diagram for eastern Baffin Island (NE Laurentide margin)
B: time/distance diagram for the southern margin of the Laurentide Ice Sheet.
C: suggested relative changes (relative to today) of winter accumulation in northern Cumberland Peninsula.
D: suggested relative changes of summer (J,J,A) temperatures for same area.
E: sign of the correlation between winter accumulation and summer temperature

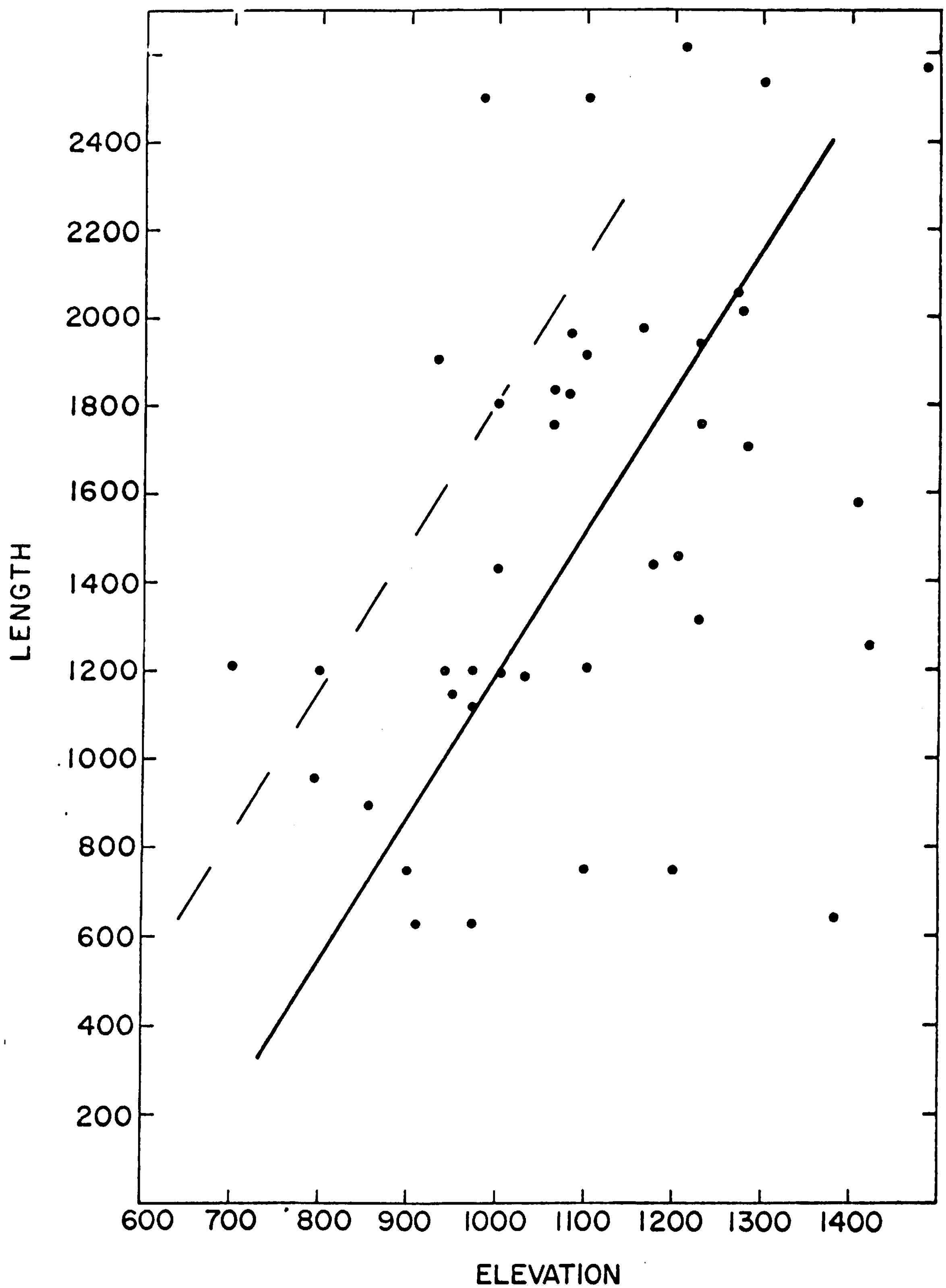


Figure 12-3: Relationship of summit elevation and glacier length in Okoa Bay and effect on length of 200 m fall in snowline.

OCCASIONAL PAPERS

INSTITUTE OF ARCTIC AND ALPINE RESEARCH

Occasional Paper No. 1: The Taxir Primer, R. C. Brill, 1971.

Occasional Paper No. 2: Present and Paleo-climatic Influences on the Glacierization and Deglacierization of Cumberland Peninsula, Baffin Island, N.W.T., Canada. J. T. Andrews and R. G. Barry and others, 1972.