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THE UNIVERSITY OF ALBERTA

THE GLACIAL HISTORY OF MARVIN PENINSULA, NORTHERN ELLESMERE ISLAND, AND WARD HUNT ISLAND, HIGH ARCTIC CANADA

BY

DONALD STANLEY LEMMEN

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled THE GLACIAL HISTORY OF MARVIN PENINSULA, NORTHERN ELLESMERE ISLAND, AND WARD HUNT ISLAND, HIGH ARCTIC CANADA submitted by DONALD STANLEY LEMMEN in partial fulfilment of the requirements for the degree of DOCTOR OF PHILOSOPHY

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ABSTRACT

Marvin Peninsula, northernmost Ellesmere Island, lies adjacent to the Arctic Ocean Basin and is bounded by Disraeli Fiord and M'Clintock Inlet. Ward Hunt Island (ca. 13 km^2) lies at the mouth of Disraeli Fiord. The glacial and relative sea level record of the field area is dominated by features related to the last glaciation. At this ice limit glaciers occupied most of the major valleys on Marvin Peninsula, while the fiords were occupied by the full glacial sea. Ice build-up involved both the expansion of existing cirque and outlet glaciers, and accumulation on plateau and lowland areas that are presently ice-free. Five AMS dates of 11.3 to 31.3 ka BP on redeposited bryophytes indicate that many ice-free areas were biologically productive throughout the last glaciation. Retreat from this ice limit had begun by 9.6 ka BP, although rapid retreat did not start until ca. 8.1 ka BP. An anomalously small amount of emergence observed in inner Disraeli Fiord suggests that neotectonics may represent one component of postglacial emergence.

At many sites meltwater channels incised into deeply weathered bedrock occur above the last ice limit. On the north coast of Marvin Peninsula these channels relate to marine shorelines, up to 117 m asl, which have been AMS dated at 27 to 32 ka BP. These dates likely represent a minimum age estimate of this glacial advance and associated sea levels. Sparse high elevation erratics are believed to relate to a former extensive glaciation of the region. Erratic shell fragments found up to 234 m asl on Ward Hunt Island may relate to this proposed event, the age of which is unknown.

Six cores up to 280 cm long were recovered from Disraeli Fiord. Deglacial(?) sediments relate to the retreat of temperate tidewater glaciers, reflecting a pronounced climatic amelioration in the early Holocene. In the latter part of the Holocene mean sedimentation rates have been 0.05 to 0.1 m/ka, indicating a low energy depositional environment, even in ice-proximal sites.

ACKNOWLEDGEMENTS

It is only 130 km from James Ross Bay to Ward Hunt Island, but it was a journey that took me more than nine years to complete. It has been an incredible experience, and I cannot adequately express my thanks to John England, who has been involved every step of the way. I remember vividly the days John and I spent in a [illegible] Logan in 1979, my introduction not only to the arctic, but to the dreaded "northernmost coast". The next summer saw us on the Hazen Plateau, and in 1982 it was off to NW Greenland. But the beauty and enchantment of the northern Ellesmere coast had infected me, and I was indeed fortunate to be able to return to the University of Alberta in 1984 to once again work with John, and to start the research for this thesis. Most of all I would like to thank John for having confidence in my abilities at the many times I questioned them, for being both a supervisor and a friend. I have been privileged to share my years at the U of A with many outstanding students. It all started in the field in 1979 and 1980 with Jan Bednarski and Tom Stewart. Throughout the last four years the fellow next door, both in the office and the field, has been David Evans. The countless hours of discussion over Pepsi and M&M's, many, many beer, and of course, the SBX-11 have shaped this thesis more than anything else. Discussions with Tom Stewart in the past year about sediments, research and the real world were greatly appreciated. The remaining fellows in the Arctic Boys Club include Tom Morris, Hector Beaudet, Trevor Bell, Philip Friend and Val Sloan. A special thanks to my very good friends Sarah O'Hara, Dirk de Boer, Francien Niekus and Bonnie Gallinger.

Every thesis based upon field work should have an appendix of photographs capturing the truly magic moments of research.. During the first season, 1985, I received the able assistance of Brian Szuster. In 1986 I was fortunate to share the field season with Maria Matishak, whose wisdom, humour and outright silliness were important factors in my most successful summer. Accompanying us that summer was Mark Tushingham from the University of Toronto. The final summer, 1987, was shared with the wonderfully bizarre Richard England and Val Sloan, who has since gone on to solve the mysteries of Canyon Fiord.

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I. INTRODUCTION

The glacial record of the Queen Elizabeth Islands has been the subject of controversy for the past two decades. The primary focus of this debate concerns the style and timing of the last glaciation. Alternative models proposed for this interval include; i) a pervasive ice cover on the Arctic Ocean and high arctic Islands (Hughes et al. 1977; Mayewski et al. 1981; Denton and Hughes 1981); ii) an extensive regional ice sheet, termed the Innuitian Ice Sheet, covering the Queen Elizabeth Islands (Blake 1970, 1975), and iii) a noncontiguous ice cover in the high arctic, termed the Franklin Ice Complex, with many of the major sounds and fiords occupied by a full glacial sea rather than glaciers (England 1976, 1983).

On northern Ellesmere Island (Fig. I.1) several geomorphologic and stratigraphic studies have demonstrated that the ice cover was limited during the last glaciation (England 1978, 1983, 1986a; England and Bradley 1978; England et al. 1978, 1981; Bednarski 1986; Retelle 1986; Evans 1988). At this time glaciers advanced 5 - 40 km beyond their present margins, terminating rapidly upon contact with the full glacial sea (England 1983). Decreased temperatures during this interval were probably associated with extensive, multi-year, landfast sea-ice and the growth of sea-ice ice shelves (England 1983, 1987a). These conditions are analogous to those which characterize the northernmost coast of Ellesmere Island today, where outlet glaciers terminate as floating tongues within many of the fiords, and the sea-ice ice shelves at the fiord mouths are unique within the northern hemisphere. Comparison of contemporary- and paleo-glaciation levels indicates that the present climate of the northernmost coast is more severe than that which characterised the area south of the

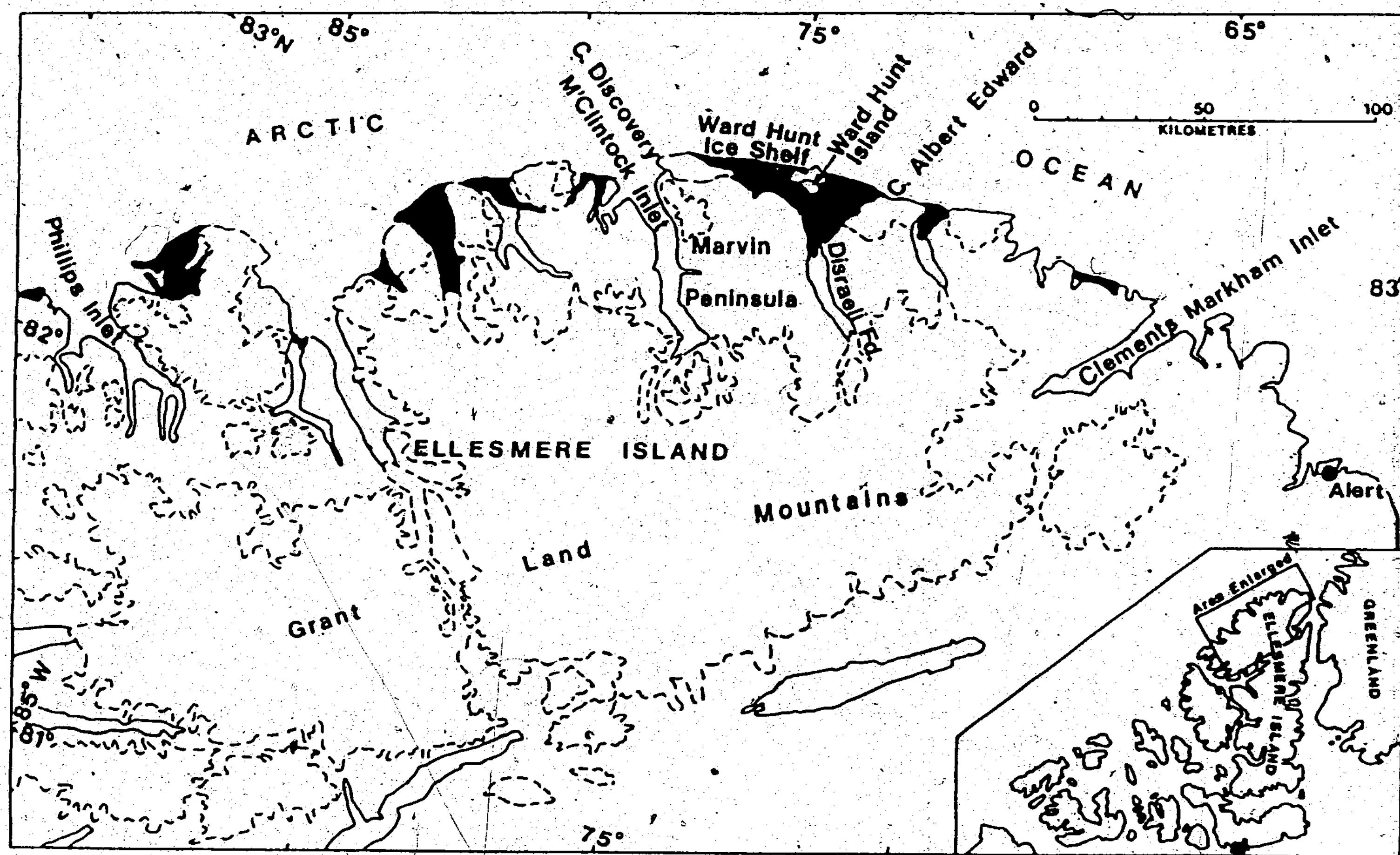

Figure I.1. The north coast of Ellesmere Island. The study area includes Marvin Peninsula, Ward Hunt Island and Disraeli Fiord. Glaciers and icefields are outlined by dashes, ice shelves are black.

Grant Land Mountains (Fig. I.1) during the last glaciation (England 1986b). Field work along Clements Markham Inlet (Fig. I.1) by Bednarski (1984) and Stewart (1988), also of the Department of Geography, University of Alberta, represents the first detailed investigation of the Quaternary history of the north coast region.

This thesis deals with the glacial and sea level history of Marvin Peninsula and Ward Hunt Island (Fig I.1). Previous research from this area concluded that a glacier filled Disraeli Fiord during the last glaciation, overtopping Ward Hunt Island and extending out onto the continental shelf (Lyons and Mielke, 1973). High elevation striae and erratics, as well as proposed moraines on the continental shelf, have been cited as evidence of extensive glaciation in this area (Hattersley-Smith et al. 1955; Crary 1956; Hattersley-Smith 1961; Lyons and Leavitt 1961; Christie 1967). It was felt that detailed research on the Quaternary history of this area was warranted to help fill a large gap in the regional glacioisostatic data base (cf. England and Bednarski 1986), and to test the model of the full glacial sea. Furthermore, this area is characterized by the most severe climate in the northern hemisphere and provides the opportunity to evaluate the relationship between glacial style and the paleoclimatic role of the Arctic Ocean Basin. This research represents an extension of the work of Bednarski (1984) and Stewart (1988), and complements ongoing Quaternary research to the west and southwest on Ellesmere Island (England et al. 1987).

The main objectives in this research were to:

1 - define the limit of the "last" glaciation and the nature of relative sea level adjustments in the field area, and to provide a chronologic framework for ice advance and retreat;

2 - evaluate the evidence for older glaciations and sea level fluctuations within the field area; and

3 - recover and analyze deep water sediments from the adjacent fiords, which would provide a continuous depositional record of paleoenvironmental change as well as insights into the contemporary processes of sedimentation in this unique glaciomarine environment.

This thesis is presented in a paper format, and each of the three papers (chapters II, III, and IV) relates specifically to one of these main objectives. Effort has been made to restrict these chapters to a length which is realistic for final publication. Highly detailed site descriptions are kept to a minimum, with the papers approached from a thematic, rather than spatial, perspective.

Objective 1: Last glaciation and related sea levels

A discussion of the last glaciation in the field area must account for the apparent conflicts between the interpretations of Lyons and Mielke (1973) and the regional model of limited ice cover provided by England (1976, 1983). The methodology employed combines the mapping of surficial geology with data collected regarding the rate and pattern of postglacial emergence (e.g. Dyke 1983). Field work primarily involves the stratigraphic investigation of glacial and raised marine deposits, the collection of marine shells from these deposits, and the surveying of raised marine shorelines. The radiocarbon dating of marine shells, driftwood and terrestrial organics provides chronologic control for the study. Many aspects of the last glaciation of this region remain poorly understood. These include: i) the onset of ice advance and attainment of the last ice limit; ii) the timing of initial deglaciation and its apparent regional differences, which suggest spatial variability in the glacioclimatic regime (England and Bednarski 1986); and iii) the existence of biologically productive refugia throughout the last glaciation (Leech 1966; Brassard 1971; Stewart and England 1986).

<u>Objective 2: Older glaciations and sea levels</u>

While controversy surrounds discussions of the ~~last glaciation~~ in the high arctic, little is known about earlier glaciations. Deposits believed to be of interglacial age have been reported from only two sites in the Queen Elizabeth Islands (Blake 1974; Blake and Matthews 1979), and examples of <u>in situ</u> interstadial deposits have not been found (cf. Blake 1982). The only well-documented record of multiple glacial and sea level events comes from northeast Ellesmere Island (England and Bradley 1978; England <u>et al.</u> 1978, 1981; Retelle 1986). A regional perspective of all glacial events is critical to our understanding of the evolution of the high arctic landscape (cf. England 1987b), and in assessing the role of the Arctic Ocean as a control of glaciation.

<u>Objective 3: Fiord sediments</u>

The fiords along the north coast of Ellesmere Island represent a unique glaciomarine environment in which the processes and products of sedimentation remain matters of speculation (Powell 1984; Syvitski 1986). The study of deep water sediments from these fiords is of direct relevance to the other objectives of this study for two reasons: I) The conditions which characterize these fiords at present (floating glacier tongues; multi-year, landfast sea-ice; and sea-ice ice shelves) are similar to those believed to typify many high arctic fiords during the last glaciation (cf. England 1983). Understanding the processes and products of contemporary sedimentation in these fiords may help in the interpretation of raised marine sequences present throughout the high arctic. II) If sedimentation rates within the fiords are sufficiently low, it should be possible to obtain sediments of full glacial age through coring. If sediments of this age are solely of marine origin, it would provide unequivocal evidence in support of England's (1983) model of limited ice extent and the full glacial sea.

BIBLIOGRAPHY

Bednarski, J. 1984. Glacier fluctuations and sea level history of Clements Markham Inlet, northern Ellesmere Island. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

__________. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 23, 1343-1355.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, 7, 634-664.

__________. 1974. Studies of glacial history in Arctic Canada. II. Interglacial peat deposits on Bathurst Island. Canadian Journal of Earth Sciences, 11, 1025-1042.

__________. 1975. Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada. Geografiska Annaler, 57A, 1-71.

__________. 1982. Terrestrial interstadial deposits, Ellesmere Island, N.W.T., Canada. American Quaternary Association, 7th Biennial Conference, University of Washington, Seattle, Washington. Program and Abstracts, p. 73.

Blake, W., Jr., and Matthews, J.V., Jr. 1979. New data on an interglacial peat deposit near Makinson Inlet, Ellesmere Island, District of Franklin. In, Current Research, Part A, Geological Survey of Canada Paper 79-1A, 157-164.

Brassard, G.R. 1971. The mosses of northern Ellesmere Island, Arctic Canada. Ecology and phytogeography, with an analysis for the Queen Elizabeth Islands. Bryologist, 74, 234-281.

Christie, R.L. 1967. Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic archipelago. Geological Survey of Canada Bulletin 138, 50 p.

Crary, A.P. 1956. Geophysical studies along northern Ellesmere Island. Arctic, 9, 155-165.

Denton, G.H., and Hughes, T.J. 1981. The Arctic Ice Sheet: An outrageous hypothesis. In The Last Great Ice Sheets. Edited by Denton, G.H., and Hughes, T.J. Wiley-Interscience, New York, 437-467.

Dyke, A.S. 1983. Quaternary geology of Somerset Island, District of Franklin. Geological Survey of Canada Memoir 404, 32 p.

England, J. 1976. Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: alternative models. Quaternary Research, 6, 185-202.

__________. 1978. The glacial geology of northeastern Ellesmere Island, N.W.T., Canada. Canadian Journal of Earth Sciences, 15, 603-617.

__________. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, 20, 895-917.

__________. 1986a. On the instantaneous removal of the Innuitian Ice Sheet. 15th Arctic Workshop, Institute of Arctic and Alpine Research, Boulder, Colorado. Abstracts, 18-19.

__________. 1986b. A paleoglaciation level for north-central Ellesmere Island, N.W.T. Arctic and Alpine Research, 18, 217-222.

__________. 1987a. The late Quaternary history of Hall Land, northwest Greenland: reply. Canadian Journal of Earth Sciences, 24, 374-380.

__________. 1987b. Glaciation and evolution of the Canadian high arctic landscape. Geology, 15, 419-424.

England, J., and Bednarski, J. 1986. Postglacial isobases from northern Ellesmere Island and Greenland: new data. Geographie physique et Quaternaire, 30, 299-305.

England, J., and Bradley, R.S. 1978. Past glacial activity in the Canadian high arctic. Science, 200, 265-270.

England, J., Bradley, R.S., and Miller, G.H. 1978. Former ice shelves in the Canadian high Arctic. Journal of Glaciology, 20, 393-404.

England, J., Bradley, R.S., and Stuckenrath, R. 1981. Multiple glaciations and marine transgressions, western Kennedy Channel, Northwest Territories, Canada. Boreas, 10, 71-89.

England, J., Lemmen, D.S., Evans, D.J.A., and Bednarski, J. 1987. The Quaternary of northern Ellesmere Island. International Union for Quaternary Research, XIIth International Congress, Ottawa, Canada. Programme with Abstracts, p. 162.

Evans, D.J.A. 1988. Glacial geomorphology and late Quaternary history of Phillips Inlet and the Wootton Peninsula, northwest Ellesmere Island, Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

Hattersley-Smith, G. 1961. Geomorphological studies in north-western Ellesmere Island. Defence Research Board, Department of National Defence, Canada, Report No. Misc. G-5, 16 p.

Hattersley-Smith, G., and other members of the expedition (A.P. Crary and R.L. Christie). 1955. Northern Ellesmere Island, 1953 and 1954. Arctic, 8, 3-36.

Hughes, T., Denton G., and Grosswald, M.G. 1977. Was there a late-Würm Arctic ice sheet? Nature, 266, 596-602.

Leech, R.E. 1966. The spiders (Areneida) of Hazen Camp 81°49'N, 71°18'W. Quastiones entomologicae, 2, 153-212.

Lyons, J.B. and Leavitt, F.G. 1961. Structural-stratigraphic studies on the Ward Hunt Ice Shelf. Air Force Cambridge Research Laboratories, Final Report Contract AF19(604)-6188, Bedford Massachusetts, 37 p.

Lyons, J.B., and Mielke, J.E. 1973. Holocene history of a portion of northernmost Ellesmere Island, Arctic, 26, 314-323.

Mayewski, P.A., Denton, G.H., and Hughes, T.J. 1981. Late Wisconsin ice sheets of North America. In The Last Great Ice Sheets. Edited by Denton, G.H., and Hughes, T.J. Wiley Interscience, New York, 67-178.

Powell, R.D. 1984. Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. Marine Geology, 57, 1-52.

Retelle, M.J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. Canadian Journal of Earth Sciences, 23, 1001 -1012.

Stewart, T.G. 1988. Deglacial-marine sediments from Clements Markham Inlet, Ellesmere Island, N.W.T., Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

Stewart, T.G., and England, J. 1986. An early Holocene caribou antler from northern Ellesmere Island, Canadian arctic archipelago. Boreas, 15, 25-31.

Syvitski, J.P.M. 1986. DNAG #2. Estuaries, deltas and fjords of eastern Canada. Geoscience Canada, 13, 91-100.

II. THE LAST GLACIATION OF MARVIN PENINSULA AND WARD HUNT ISLAND, N.W.T.

Introduction

Previous research concerning the style and timing of glaciation along the northern coast of Ellesmere Island, bordering the Arctic Ocean Basin, has largely been of a reconnaissance nature (e.g. Hattersley-Smith et al. 1955; Christie 1957, 1967; Lyons and Mielke 1973). In a recent detailed study of Clements Markham Inlet (Fig. II.1) Bednarski (1986) concluded that the last ice limit was restricted to the head of the inlet and major valley systems, with glaciers terminating shortly after contact with the sea. From a regional perspective Bednarski's findings support the model of a limited ice cover during the last glaciation (England 1976), with most of the major fiords and channels being occupied by "the full glacial sea" (England 1983) rather than by glacier ice. Collectively, these studies reject the model of a pervasive regional ice cover, the Innuitian Ice Sheet, originally proposed by Blake (1970) and subsequently accepted by others (e.g. Lyons and Mielke 1973; Hughes et al. 1977; Mayewski et al. 1981).

Presented in two companion papers (Chapters II and III) are the glacial geomorphology and stratigraphy, as well as the relative sea level history, of Marvin Peninsula and Ward Hunt Island (Fig. II.1). These areas constitute the largest ice-free terrain along the north coast west of Clements Markham Inlet. The field area includes many sites discussed in previous reconnaissance studies. The primary

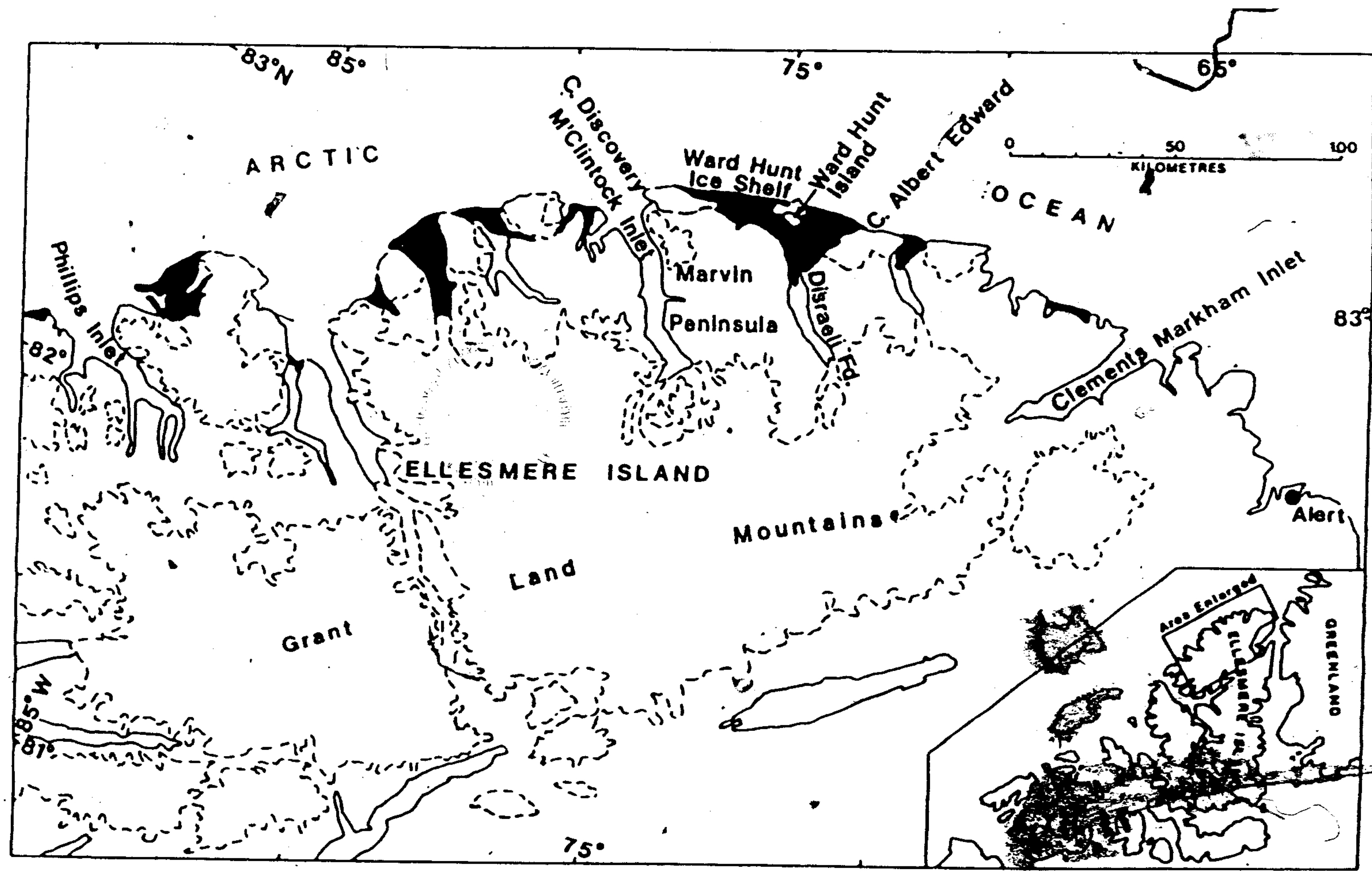

Figure II.1. The north coast of Ellesmere Island with official place names. Glaciers and icefields are dashed, ice shelves are black.

objective in this study is to expand the present data base concerning regional stratigraphy and postglacial emergence, with emphasis placed upon glacial style and deglacial chronology. This paper is focussed upon the last glaciation whereas the second paper concerns the older glacial and sea level record.

The term "last glaciation" has been used informally by numerous workers to refer to the last major ice build-up in the area, subsequently recorded by retreat from moraines contacting dateable raised marine shorelines (e.g. England 1978, 1983, 1985; Hodgson 1985; Bednarski 1986; Retelle 1986; England et al. 1987). Data concerning initiation of the ice advance associated with the last glaciation are sparse (cf. Blake 1982). "Last glaciation" does not necessarily imply the entire period since the last interglaciation, evidence for which has not yet been observed in the terrestrial stratigraphy of northern Ellesmere Island. It is debatable whether the terms "Glacial" and "Interglacial" have any useful application in this extreme high arctic environment, especially given the dominance of precipitation changes over temperature for the build-up of ice (Bradley and England 1978).

Study area

Physiography

Marvin Peninsula borders the Arctic Ocean and is bounded by Disraeli Fiord and M'Clintock Inlet to the east and west, respectively, both ca. 60 km in length (Fig. II.2). A major trough, termed Central Valley, extends NE to SW across the peninsula with a divide at <300 m

NEW RADIOCARBON DATES FROM MARVIN PENINSULA

(No. at left refers to location map)

No.	Lab No.	Material	Date	Sample Elev. (m)	Related Sealevel
	M'CLINTOCK INLET				
1	TO-499	marine shells	8140±90	73	<122
2	TO-267	marine shells	8870±110	84	>84≤120
3	TO-262	marine shells	7770+70	36	?
4	TO-498	subfossil bryophytes	14,730±120	98	>98≤122
5	TO-497	subfossil bryophytes	15,710±180	89	>89≤120
6	TO-263	marine shells	7800±90	46	81
7	TO-265	marine shells	7320±90	32	≥68
	NORTH COAST				
8	TO-861	marine shells	8630±70	42	<67
9a	TO-863	driftwood	5800±50	0.5	?
9b	TO-864	driftwood	5730±60	0.5	?
9c	GSC-4559	driftwood	8850±80	0.5	?
10a	TO-488	marine shells	9560±70	39	?
10b	TO-487	marine shells	9080±110	36	?
10c	TO-489	subfossil bryophytes	23,340±430	41	?
11	TO-490	marine shells	8630±100	46	≤74
	OUTER DISRAELI FIORD				
12a	TO-862	marine shells	9250±80	62	78
12b	TO-857	sandy peat	11,340±70	62	78
	CENTRAL VALLEY EAST				
13	TO-269	marine shells	8150±60	45	≤92
14	TO-270	marine shells	8860±60	46	<97
	TO-500	shell fragments	30,440±330	81	?
	TO-261	driftwood	6510±70	19	>19
	TO-268	marine shells	5450±90	7	?
	TO-264	marine shells	4920±70	15	<40
	TO-266	marine shells	4080±60	3	<59
	INNER DISRAELI FIORD				
15a	TO-496	marine shells	8150±80	64	75
15b	*TO-519	marine shells	7950±60	64	75
15c	*TO-520	marine shells	7800±60	64	75
15d	*TO-521	marine shells	10,[illegible]90±70	64	75
16	TO-493	subfossil bryophytes	7730±70	39	≥68
17	TO-491	marine shells	8010±100	31	≥68
	TO-494	marine shells	7260±80	33	<68
	*TO-525	marine shells	7110±60	32	<68
18	TO-492	subfossil bryophytes	31,360±400	72	lacustrine
19	TO-495	driftwood	6030±70	27	?

* dates courtesy of M. Tushingham, Department of Physics, University of Toronto, unpublished data

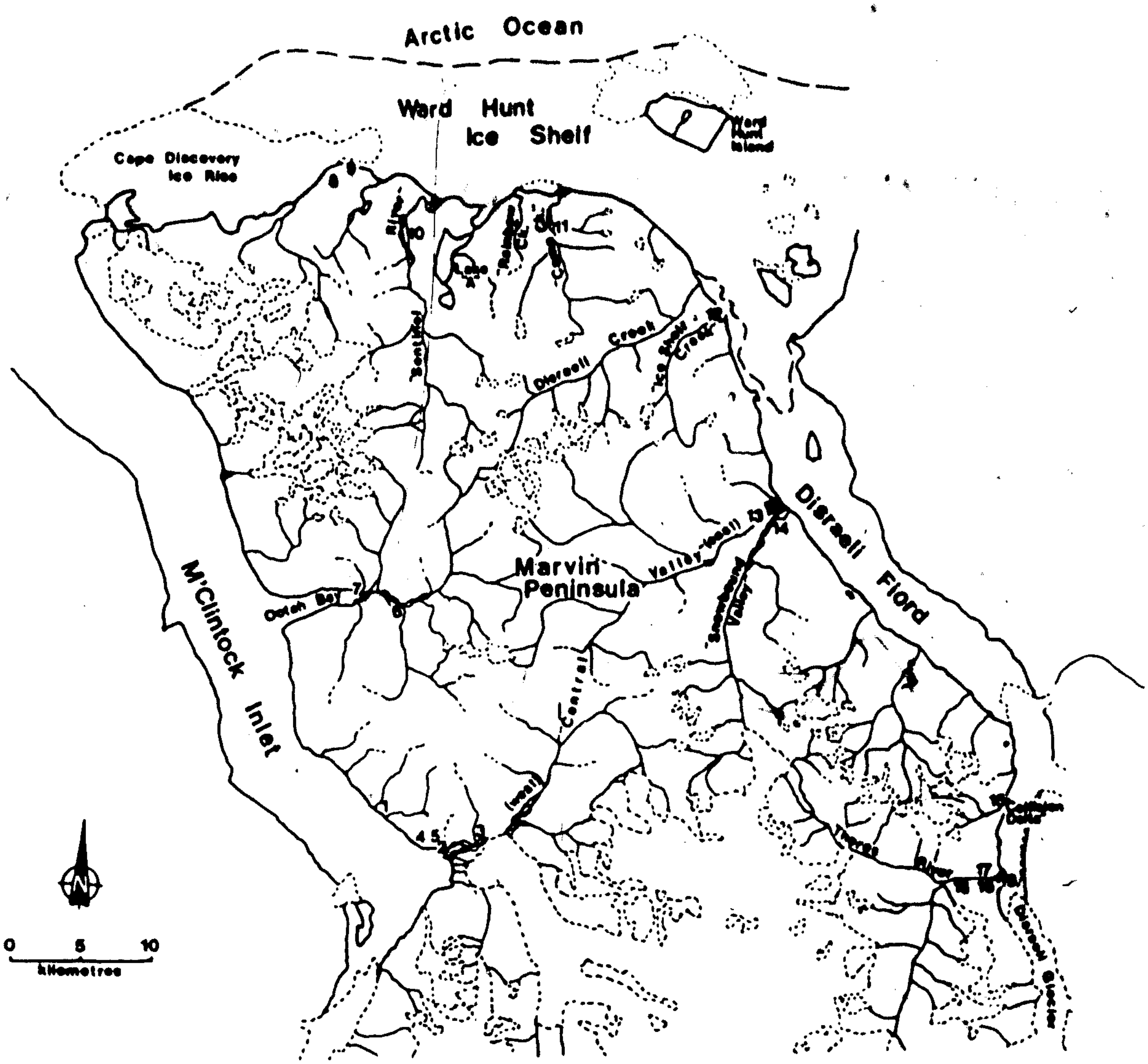

Figure II.2. Marvin Peninsula and Ward Hunt Island. Glaciers and ice rises are outlined by fine dashes. Names in quotations are unofficial. Numbers refer to locations of radiocarbon dated samples listed at left. All "TO" sample numbers refer to AMS dates run by Isotrace Laboratory, University of Toronto. Shell samples received a 30 to 50% preleach. Dates are corrected to a base $^{13}C = 0‰$. . The errors represent one standard deviation (68.3% confidence limits).

asl. To the south, the Grant Land Mountains and icefields reach 2500 m asl with ice thicknesses of 700 m (Hattersley-Smith et al. 1969; Narod et al. 1988). Transection glaciers discharge from these icefields into the fiord heads, forming small glacial ice shelves. North of Central Valley, isolated summits reaching 1100 m asl support small ice caps and cirque glaciers. Ward Hunt Island (83°05' N) lies at the mouth of Disraeli Fiord, 6 km north of Marvin Peninsula. The island is surrounded by the Ward Hunt Ice Shelf, which extends westward from Cape Albert Edward to Cape Discovery, covering ca. 800 km² (Fig. II.1). Ice rises, which constitute sea level glaciers, are formed where the ice shelf becomes grounded along Ward Hunt Island, the Marvin Islands and the north coast of Marvin Peninsula (Fig. II.3).

Climate

Northern Ellesmere Island is a polar desert, with low precipitation and an annual water balance close to zero (Bovis and Barry 1974). Winters are characterized by extreme cold (commonly -35°C) and minimal precipitation, whereas the short summers are characterized by frequent temperature inversions, increased precipitation owing to greater cyclonic activity, and coastal fog associated with runoff and breakup of the pack ice (Maxwell 1982). The mean annual air temperature at Alert (180 km east of Marvin Peninsula) is -18°C and mean summer temperatures (J,J,A) are ca. +2°C with high inter-annual variability (Bradley and England 1978). Westerly winds are dominant through most of the year, although easterly winds tend to be associated with the greatest wind speeds, especially in summer (Maxwell 1982). A steep climatic gradient characterizes the northernmost coast, demonstrating the localized

Figure II.3. Oblique air photo looking east along the north coast of Marvin Peninsula from above the Cape Discovery Ice Rise (CDIR). Other labels identify the Ward Hunt Ice Shelf (WHIS), Ward Hunt Island (WHI) and Disraeli Fiord (DF).

influence of the Arctic Ocean on the glacioclimatic conditions of the field area. As one moves inland the local glaciation level increases from about sea level around the Ward Hunt Ice Shelf to > 1100 m asl over the Grant Land Mountains (Miller et al. 1975; Bednarski 1984).

Pre-Quaternary geology

Geologically, the study area is one of the most complex regions in the Arctic Islands (Trettin 1981). A comprehensive synthesis of the regional geological evolution is presented by Trettin (1987). The area is primarily composed of Precambrian to Upper Silurian lithologies which include; volcanics, intrusive and metamorphic rocks, as well as a wide range of sedimentary rocks, from boulder conglomerates to carbonates (Christie 1964; Trettin 1969, 1981; Frisch 1974). Upper Paleozoic sedimentary rocks (undifferentiated) and minor outcrops of Tertiary conglomerate have also been mapped (Trettin 1981). Since the Ordovician the region has been affected by at least four major orogenies, the most recent being the Eurekan Orogeny of latest Cretaceous to Oligocene time (Trettin 1987). Given the structural and lithological complexity of the field area, it is difficult to identify glacial erratics that definitively record past ice flow directions.

Quaternary history - previous investigations

Much of the early field work in the region focussed upon the history and structure of the Ward Hunt Ice Shelf (cf. Ommanney 1982; Jeffries 1987). Nonetheless, high elevation erratics and other glacial features on Ward Hunt Island and northern Marvin Peninsula apparently demonstrate extensive glaciation of unknown age (Hattersley-Smith et al.

1955, Lyons and Leavitt 1961; Christie 1967; Lyons and Mielke 1973). North of Ward Hunt Island, sparse bathymetric data were interpreted as indicating large moraines on the continental shelf (Crary 1956). Furthermore, the presence of fiords and the discordant relationships between tributary and trunk valleys have all been attributed to glaciation (Crary 1956; Hattersley-Smith 1961; Christie 1967). Lyons and Mielke (1973), using theoretical ice sheet profiles predicated upon their acceptance of an Innuitian Ice Sheet (Blake 1970), concluded that at least 600 m of ice filled Disraeli Fiord during the last glaciation and extended 16 km to the north of Ward Hunt Island (see England 1974). Radiocarbon dates on marine shells (7200±200 BP, L248A, Crary 1960; and 7755±150 BP, SI-718, Lyons and Mielke 1973) collected from a prominent beach at 38 m asl on Ward Hunt Island provided estimates of initial ice retreat and postglacial emergence. Prior to this study, marine shells dated 8130±120 BP (GSC-1850, Blake 1987) provided the oldest date associated with the last glaciation of the field area. A complete list of radiocarbon dates from the field area is contained in Appendix I.

Glacial geomorphology

Although field work concentrated on coastal sites, the mapping of surficial geology from air photographs covered the entire field area (Fig. II.4, Appendix II). In general, small-scale glacial erosional features are not commonly preserved. Glacially molded bedrock is most evident along the west shore of inner Disraeli Fiord. Striae, while common on erratic boulders, are less frequently found on bedrock except

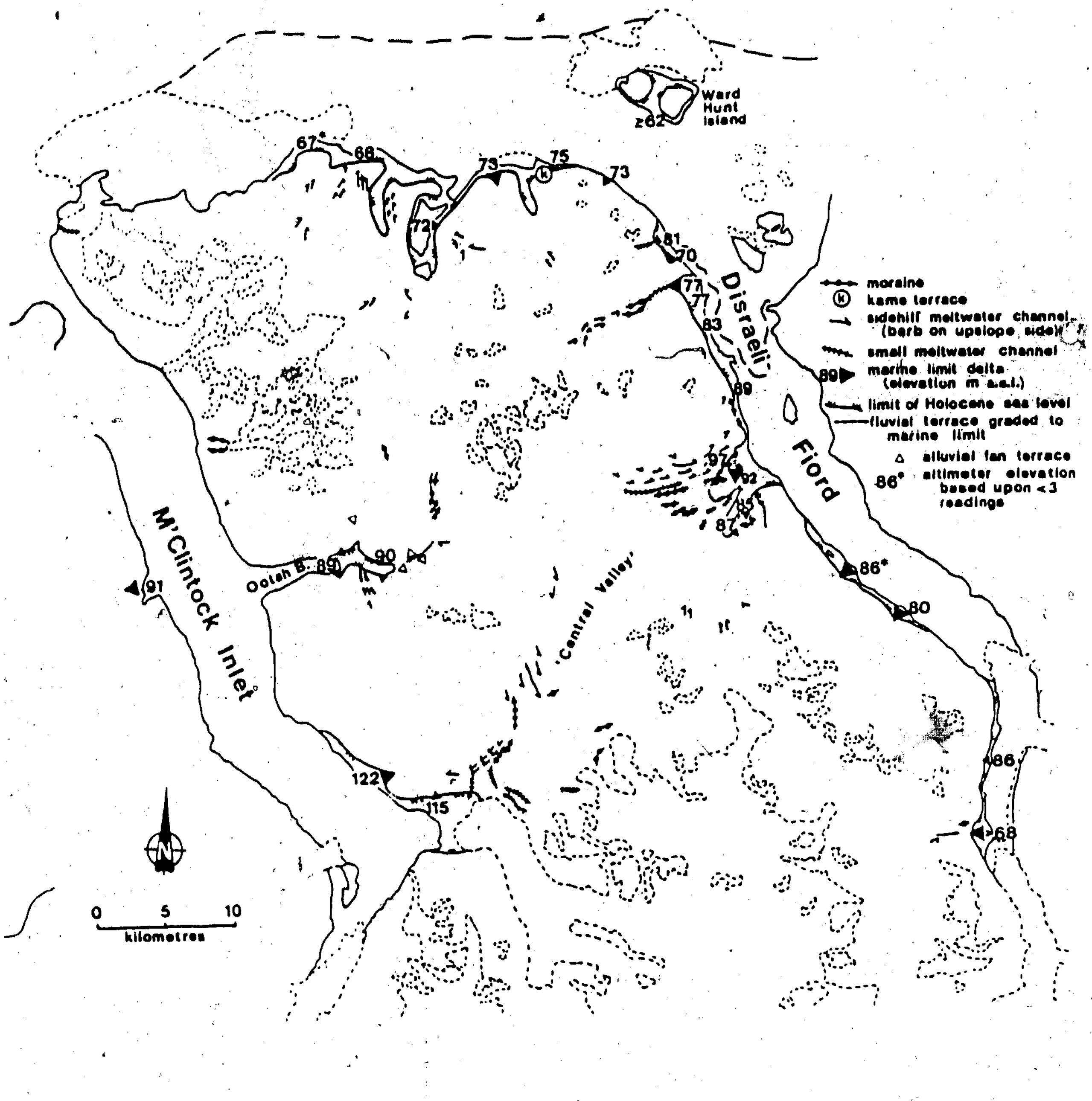

Figure II.4. Prominent glacial and marine features on Marvin Peninsula and Ward Hunt Island. Elevations were determined using a Wallace and Teiran altimeter and believed accurate to ±1.5 m. Unless noted, elevations are based upon more than three independent readings from sea level.

where protected by overlying drift. Where found, striations indicate ice flow parallel to the major valley systems (cf. Christie 1967).

Meltwater channels, usually incised in bedrock, provide the most widespread record of former ice margins. Most glacigenic sediments are confined to the main valley systems, and rarely exceed a thickness of one metre. Moraine systems record a prominent ice limit in three of these valleys: both ends of Central Valley; Ootah Bay and Disraeli Creek (Fig. II.4). Elsewhere ice limits are inferred from the distribution of glacial diamictons, lateral meltwater channels, ice-contact deltas, proglacial fluvial deposits and relative weathering of bedrock.

Moraines along the north wall of Central Valley (Figs. II.4 - II.6) record a former ice flow parallel to the main axis of the valley. The glacier which occupied the trough had a roughly symmetrical NE-SW profile, with a divide located just east of the present drainage divide (<300 m asl). At both ends of the valley the maximum elevation of the outermost lateral moraines is ca. 205 m asl. The moraines consist of silty diamicton with coarse clasts which are generally rounded. At the east end of Central Valley these moraines form a terminal loop marked by a thick wedge of glaciomarine diamicton. This landform is interpreted as a morainal bank deposit (Powell 1981) marking the grounding line of a tidewater glacier in Disraeli Fiord. Shell fragments collected from the diamicton dated 30,440±300 BP (TO-500). The terminal position of ice flowing southwest towards the west end of Central Valley is not as clearly recorded. At its maximum extent this glacier contacted a second lobe of ice blocking the mouth of the valley which, in turn, terminated in M'Clintock Inlet.

Figure II.5. Vertical air photo mosaic of the east end of Central Valley denoting major glacial and marine features. Abbreviations: moraines (M), meltwater channels (mwc), morainal bank deposit (MB), marine limit (ml). Arrows indicate the direction of former ice flow. The edges of marine deltas are outlined, elevations in metres asl. Dots show the location of radiocarbon dated samples.

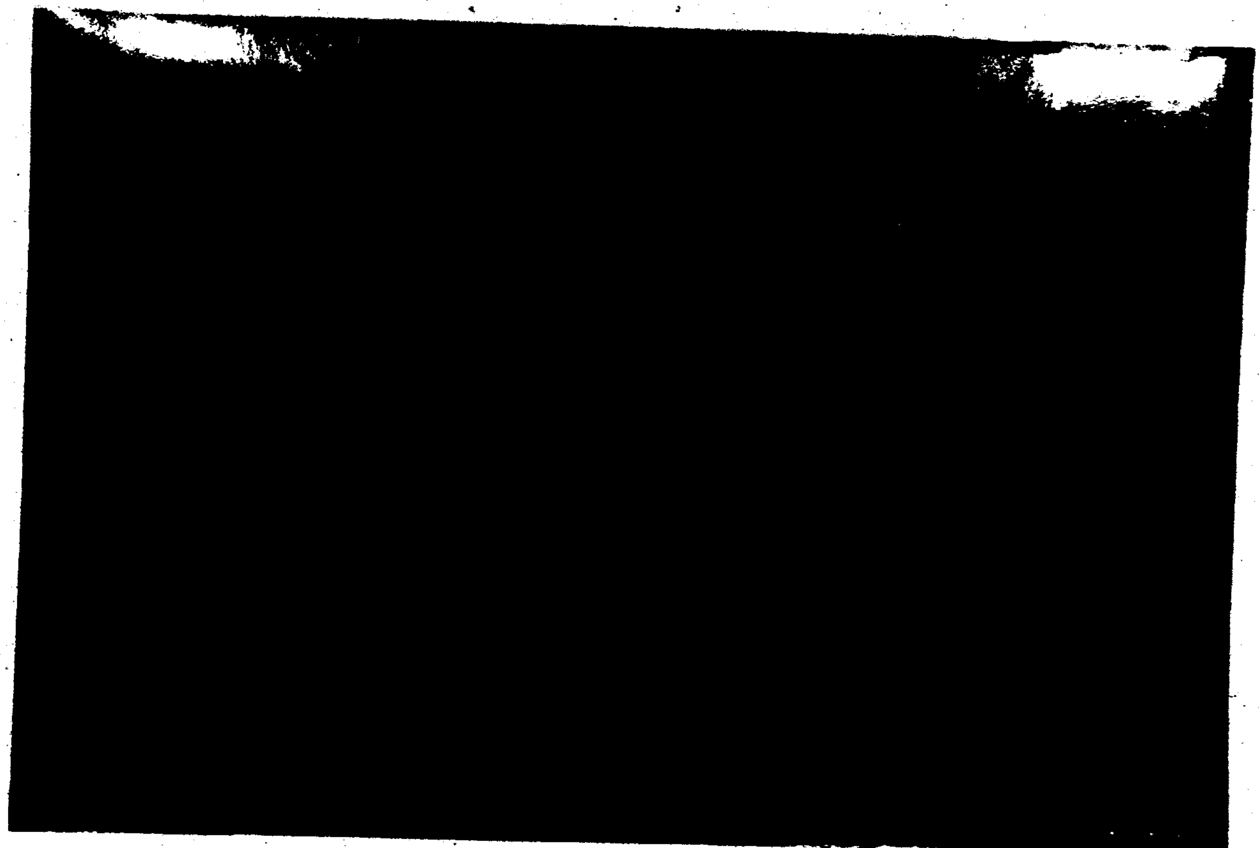

Figure II.6. Vertical air photo of the west end Central Valley. Dashed arrow shows location of cross-valley which leads to 122 m asl delta. Other symbols and abbreviations are the same as in Fig. II.5.

Inland of Ootah Bay (Fig. II.4), prominent moraines occur about 10 km up valley. Distal to these moraines is a discontinuous glaciofluvial terrace which grades to ca. 90 m asl in the lower valley. Inactive alluvial fan surfaces, related to this major terrace, are present at the mouths of all tributary valleys.

Ice-contact deltas are another important ice marginal indicator, particularly, at the east end of Central Valley and along the west coast of outer Disraeli Fiord. All of these deltas are related to local valley glaciers which formerly contacted the fiord. A prominent kame terrace at ca. 83 m asl along the north coast of Marvin Peninsula relates to the flow of ice out of the Camp Creek Valley (Figs. II.2 and II.4). This terrace, initially described by Lyons and Mielke (1973), is composed predominantly of glaciofluvial boulders.

Overstep sequences of glacial marine silt abruptly overlying till provide stratigraphic evidence of glaciers grounded below sea level and the transgression of the sea upon ice retreat (Bednarski 1988). Sediments exposed in outer Sentinel Valley and Snowbound Valley (Fig. II.2) show such a relationship. Elsewhere, stratigraphic sections containing till are rare, with the most common sequence observed consisting of marine or fluvial sediments directly overlying bedrock.

Above the uppermost extent of significant glacigenic deposits, the surficial material is dominated by bedrock, colluvium and gelifluctng residuum. Tors are common, and are found <3 m above the marine limit at the east end of Central Valley. North of Ootah Bay tors >6 m in height are found in volcanic flow rocks on a summit ca. 800 m asl. Although some summits show no evidence of glaciation, sparse erratics are found at almost every elevation in the field area. Meltwater channels occur up to elevations of ca. 400 m asl, with the higher channels incised within highly weathered bedrock. On Ward Hunt Island, weathering and slope processes make the recognition of glacigenic deposits difficult. However, highly weathered erratics are common and occur within 2 m of the summit of the island (415 m asl), while tiny shell fragments (probably glacially transported, Chapter III) occur up to 234 m asl.

Glacial marine sediments

Like elsewhere in the high arctic the mouths of most of the major valleys on Marvin Peninsula are occupied by glaciomarine deltas which record progradation along coastlines that have emerged following deglaciation. The local marine limit, which usually coincides with the highest delta, ranges from 122 m asl in inner M'Clintock Inlet to roughly 62 m asl on Ward Hunt Island (Fig. II.4). However, many small valleys which drain directly into the fiords did not produce sufficient sediment during deglaciation to prograde deltas to the marine limit (England 1982, 1983; Bednarski 1988). At these sites, the highest marine features are beaches or washing limits at elevations somewhat below the marine limit recorded by deltas in adjacent valleys. The problem of recognizing the marine limit is particularly evident on Ward Hunt Island, where marine sediments, beaches and washing limits are rare. At many localities fine-grained, fossiliferous, marine sediments are common below the marine limit, but it can be problematic relating these to a specific former sea level. Vast expanses of dissected marine silt as described elsewhere on northern Ellesmere Island (cf. Christie 1967; Bednarski 1986), are only found at the east end of Central Valley.

The elevation of marine limit descends from Central Valley (122 m asl at west end, 97 m asl at east end) to the north coast, where it reaches its lowest elevation of ca. 62 m asl on Ward Hunt Island (Fig. II.4). South of Central Valley, along Disraeli Fiord, marine limit descends to ca. 86 m and shows no systematic gradient southward to the fiord head. At Thores River, which drains to the inner west shore of Disraeli Fiord (Fig. II.2), sections feature foreset beds extending to the surface of the marine limit delta at 68 m asl. Because the topset beds have been removed by erosion, this surface must represent a minimum estimate of the local marine limit. It is also possible that fluvial incision has removed sediments recording the true marine limit in other confined valley systems.

Prominent deltas were surveyed in 18 different valleys while beaches and washing limits were surveyed at more than 15 other, intervening localities. Together with 34 new accelerator mass spectrometry (AMS) radiocarbon dates, these data record the pattern of postglacial emergence and relative sea level changes in the field area.

Relative sea level history

This section presents a discussion of the evidence for, and ages of, former sea levels. A site-specific approach is taken, proceeding in a clockwise circuit from the head of M'Clintock Inlet northward to Ward Hunt Island, and then inland to the head of Disraeli Fiord.

M'Clintock Inlet

At the northwest extremity of Central Valley, a prominent delta at 122 m asl records marine limit near the head of M'Clintock Inlet (Figs. II-4, II-6). This delta was constructed when the mouth of Central Valley was blocked by the large outlet glacier which enters the valley from the south (Fig. II.6). Advance of this glacier diverted meltwater through a small cross-valley leading to M'Clintock Inlet. A distinct break in slope occurs between ca. 100 and 115 m asl along the north side of outer Central Valley and records glaciomarine deposition beyond the retreating ice margin. A second large delta was constructed when relative sea level fell to 115 m asl, at which time drainage through the cross-valley was abandoned. Distal to the 122 m delta whole valves of Hiatella arctica were collected and dated at 8140±90 BP (TO-499) whereas a second sample, between the 122 and 115 m deltas, dated 8870±110 BP (TO-267; sites 1 and 2, Fig. II-2). The date of 8870 BP provides a minimum age estimate on both deglaciation and the 122 m sea level. A third date of 7770±70 BP (TO-262) was obtained on whole valves of Hiatella arctica within sandy marine silt <300 m beyond the present ice margin (site 3, Fig. II-2).

East of Ootah Bay, along central M'Clintock Inlet (Fig. II.2), the marine limit was not observed within the main valley. Even small tributary valleys did not generate sufficient sediment during deglaciation to produce well defined deposits at marine limit. Deltas at the mouths of these valleys have a cone-like morphology. The highest marine features in the valley, measured at five different sites, are wave-washed surfaces at 90±2 m asl (Fig. II.4). Marine shells are sparse, but they record the transgression of the sea at least 3.5 km upvalley. Two shell samples from foreset beds of different deltas were dated (TO-263 and TO-265; sites 6 and 7 respectively, Fig. II.2). The oldest of these dates (7800±90 BP) is related to a former sea level at 81 m asl, and provides a minimum estimate on ice retreat within the tributary basins.

North coast

Along the north coast, between Lake A and the Cape Discovery Ice Rise (Fig. II.2), a series of prominent beaches that terminate at ca. 68 m asl are considered to mark the maximum height of the sea during the Holocene (Fig. II.4). A minimum age estimate on this sea level of 8630±70 is provided by paired bivalves collected from marine silts behind the eastern margin of the ice rise (TO-861; site 8, Fig. II.4). In outer Sentinel River Valley, the local marine limit is defined by wave-washed surfaces which extend to ca. 117 m asl. However, associated radiocarbon dates demonstrate that this higher shoreline predates the last glaciation (Chapter III). Two km up-valley of this high sea level, from an elevation of <40 m asl, marine silt abruptly overlies glacial diamicton. Complete bivalves of Portlandia arctica from these silts dated 9560±70 BP (TO-488) and 9080±110 BP (TO-487), providing excellent control on the timing of initial deglaciation within the valley (site 10, Fig. II.2).

Immediately east of Lake A, at the mouth of Rainbow Creek, a large delta at 73 m asl records the local marine limit (Fig. II.4). Shells associated with marine limit were not found at this site, and are rarely observed along the northernmost coast. Four km further east a small cone of foreset bedded sands abuts the eastern end of a kame terrace.

The outer lip of this delta has an elevation of 75 m asl (Fig. II.4). Shells were not found at this site, however, three km up-valley, along the east side of Camp Creek, whole bivalves of Hiatella arctica collected at 46 m asl dated 8630±110 BP (TO-490) (Site 11, Fig. II.2). This date provides a minimum estimate for the age of the marine limit recorded at the mouth of the valley. This sample is from the same general location as shells dated 7045±190 BP (SI-725) from about 30 m asl reported by Lyons and Mielke (1973).

Ward Hunt Island

The steep slopes and small drainage basins of Ward Hunt Island make recognition of the marine limit difficult. A prominent beach berm on the north shore of the island at 38 - 42 m asl has been described by numerous authors. Marine shells are abundant on this beach, however, dates of 7200±200 BP (L248A, Crary 1960) and 7755±150 BP (SI-720, Lyons and Mielke 1973) suggest that they represent a mix of different aged bivalves that were not restricted to this sea level. Wave-washed benches along the south shore of the island extend to 62 m asl (Fig. II.4), representing a minimum estimate of the marine limit. Large fragments of marine shells that predate the last glaciation were collected from between 52 and 115 m asl (Chapter III).

Outer Disraeli Fiord

Along outer Disraeli Fiord the marine limit is well defined by a series of deltas and intervening beaches (Fig. II.4). Immediately north of Disraeli Creek a large delta complex at 81 m asl provides the best estimate of the local marine limit. This delta is unrelated to any significant drainage basin, but meltwater channels incised into bedrock indicate that it once received meltwater from ice occupying a plateau immediately to the west. Within Disraeli Creek proper the highest delta occurs at 70 m asl. To the south, in the basin between Disraeli and Ice Shelf Creeks, a large delta occurs at ca. 77 m asl (Fig. II.4). A broad, coarse-grained outwash plain indicates that this delta received meltwater primarily from ice occupying Disraeli Creek. Abundant bivalves of Portlandia arctica, collected from a large subaqueous channel deposit immediately distal to the 77 m delta, dated 9250±80 BP (TO-862; site 12, Fig. II.2).

Between Ice Shelf Creek and Central Valley the marine limit is recorded by discontinuous beaches, whereas small deltas related to minor drainage basins usually are not graded to marine limit. Immediately south of Ice Shelf Creek the highest prominent beach occurs at 77 m asl, whereas ca. 2.5 km further south beaches extend to 83 m asl (Fig. II.4). Meltwater channels and ice-contact deposits indicate that, at some sites, local ice from these small valleys terminated in the sea. Unfortunately most of this area was not traversed.

Disraeli Fiord - Central Valley

The marine limit along Disraeli Fiord attains a maximum elevation at the northern margin of Central Valley. Here, a wave-washed terrace at 97 m asl occurs distal to the outermost moraines in the lower valley, and records the penetration of the sea in a narrow strip between the mountain slope and former glacier margin (Figs. II.4, II.5). Following retreat of the ice an ice-contact delta was constructed at 92 m asl inside the moraines. Complete bivalves of Mya truncata from clayey-silt distal to, and apparently underlying, this delta dated 8150±60 BP (TO-269; site 13, Fig. II.2). On the south side of the valley Portlandia arctica dating 8860±60 BP (TO-270) were collected from marine silts overlying glacial diamicton (site 14, Fig II.2). This date provides a minimum estimate on the establishment of the 97 m marine limit as well as the timing of initial ice retreat. Ice-contact deltas at 87 and 85 m asl within Snowbound Valley record continued retreat of ice up-valley (Figs. II.4, II.5).

Inner Disraeli Fiord

Two large deltas located 10 and 14 km south of Central Valley record local marine limits of ca. 86 and 80 m asl, respectively, within large valley systems (Fig II.4). These sites have not been radiocarbon dated. Further up-fiord, the marine limit becomes less well defined, despite the presence of many small deltas. At Collision Delta (Fig. II.2) silt underlying colluvium extends up to a prominent break in slope at 86 m asl, marking a minimum elevation of the local marine limit. Sandy deltaic sediments related to the retreat of a local cirque glacier, and not a trunk glacier in Disraeli Fiord, are inset against the indurated silt at 75 m asl. In situ bivalves of Hiatella arctica from the gravelly-sand foresets of this delta dated 8150±80 BP (TO-496) (site 15, Fig. II.2). Three other dates are also available on this 75 m delta, from a sample of Hiatella arctica collected by a second researcher. The dates are; 7800±60 BP (TO-520), 7950±60 BP (TO-519) and 10,090+70 BP (TO-521; M. Tushingham, personal communication 1988). Given that all of the dated shells were found in situ and collected from the same bed, and indeed within centimetres of each other, the 10,090 BP date is anomalous. On the basis of the reasonable consistency of the other Holocene dates in this area, the date of 10,090 BP is rejected.

At the head of Disraeli Fiord, along the south side of the Thores River (Fig. II.2), a large delta at 68 m asl provides a minimum estimate on the height of the local marine limit (Fig. II.4). A date of 7730±70 BP (TO-493) was obtained on bryophytes which had been redeposited in bedded sand beneath the marine limit delta (site 16, Fig. II.2). A date of 8010±100 BP (TO-491) was obtained on Portlandia arctica collected from contorted silt and clay at 31 m asl on the north side of the valley (site 17, Fig. II.2). However, the stratigraphic relationship between this sample and the marine limit is not clear. The ca. 8000 BP sediments were likely deposited within metres of the grounding line of the Disraeli Glacier, and perhaps were overridden by the fluctuating ice margin. The marine limit delta, on the other hand, clearly relates to sedimentation from within Thores Valley and was prograded at a slightly younger date, following retreat of the Disraeli Glacier across the mouth of the valley.

Two km upvalley (west) of the marine limit delta in Thores Valley, 45 m thick sections of bedded sand and gravel extends ca. 110 m asl. These sediments are considered to record deposition in a lake dammed by the Disraeli Glacier when it was grounded across the mouth of the valley (Matishak and Lemmen 1987). This interpretation is based upon the absence of marine fauna in the sediments and the abundant ice-rafted debris at the base of the sections nearest Disraeli Fiord. A pre-Holocene radiocarbon date was obtained on redeposited terrestrial organics found within these sediments (see next section). Small pockets of glaciolacustrine sediment occur at least 2 km further west along the north valley wall, at an estimated height of 150 m asl.

Pre-Holocene organics

Five samples of subfossil bryophytes which predate the Holocene were found within waterlain sediments on Marvin Peninsula. The oldest of these samples, dating 31,360±400 BP (TO-492), was found within the ice-dammed lake sediments in the lower Thores River valley (site 18, Fig. II.2). The bryophytes were dispersed within horizontally bedded, medium sand, and overlain by ca. 20 m of coarsening-upwards sediments that record the infilling of the lake. Part of a bird feather, as yet unidentified, was found along with the bryophytes.

A sample of subfossil bryophytes from marine sediments within the Sentinel River valley provided a date of 23,340±430 BP (TO-489; site 10, Fig. II.2). These organics were obtained from horizontally bedded sand 2-5 m above the silt containing marine shells dated [illegible]000 to 9500 BP (TO-487 and TO-488). Similar-looking organics are common within sand exposed in numerous sections along the east river bank. A similar, anomalously old date has been obtained on the basal sediments of a small lake in the Sentinel River Valley (M. Retelle, personal communication, 1987).

Radiocarbon dates of 14,730±120 BP (TO-498) and 15,710±180 BP (TO-497) were obtained from bryophytes incorporated within marine sand near the head of M'Clintock Inlet (sites 4 and 5, Fig. II.2). The 14,730 BP sample was collected from a small section exposed along a gully at 98 m asl. The uppermost sand of this section contain symmetrical bedforms

Figure II.7. Subfossil bryophytes contained within cross-laminated marine sands. Sample from this site (location 5, Fig. II.2) at 89 m asl was dated 15,710±180 BP (TO-497). Scale = 17 cm.

provisionally interpreted as wave ripples. There is no significant marine deposition found above 100 m asl at this site, although the local marine limit is 115 to 120 m asl (Fig. II.4). The 15,710 BP sample was found at 89 m asl within coarse to medium sand dominated by current ripples and climbing ripples (Fig. II.7). Paleocurrent directions indicate flow parallel to the contemporary slope. The section fines upward into silt which is laterally equivalent to that dated 8870 BP (TO-267, site 2, Fig. II.2). Analysis of this sample reveals eight bryophyte species (Appendix III), with very good preservation. All of the species are extant on northern Ellesmere Island, and are characteristic of a high arctic wet meadow community (Brassard 1971; LaFarge-England 1988).

The youngest of the five dates, 11,340±70 BP (TO-857), was obtained on dispersed organics deposited in sandy-silt distal to the marine limit delta of Ice Shelf Creek (site 12, Fig. II.2). Marine shells from the same bed as the organics dated 9250±80 BP (TO-862).

Driftwood

Previous studies concerning driftwood penetration within the field area have concentrated on low elevation samples which may be related to the formation of the Ward Hunt Ice Shelf (Crary 1960; Hattersley-Smith 1973; Stewart and England 1983; Blake 1987). Previously dated Holocene wood from this area ranges from 3000±200 (L254D; Crary 1960) to 6280±140 BP (SI-568, Mielke and Long 1969; Appendix I). Six other samples have yielded "greater than" radiocarbon dates (Mielke and Long 1969; Blake 1987). All driftwood above ca. 3 m asl found during this study had clearly been moved downslope subsequent to its original deposition. Along the north coast of Marvin Peninsula, about 3 km east of the Cape Discovery Ice Rise, more than twenty pieces of wood were found between 0.4 and 1 m asl. Dates on three of these samples were 8850±50 BP (GSC-4559), 5800±50 BP (TO-863), and 5730±60 BP (TO-864; site 9, Fig. II.2). From inner Disraeli Fiord, a sample found at 27 m asl along the southern margin of the Thores River delta was dated at 6030±70 BP (TO-266; site 19, Fig. II.2). This site is 11 km inside the margin of the floating glacier blocking the inner fiord, and therefore ice was upvalley of its present position ca. 6000 BP.

Interpretation

Last glacial ice limits

The widespread presence of sparse erratics at all elevations in the field area indicates that most, if not all, of Marvin Peninsula was subject to extensive glaciation at some unknown time in the past. However, it is possible that some of the highest peaks in central Marvin Peninsula remained unglaciated. For the last glaciation one cannot unequivocally define the limit of ice based upon morphological evidence alone. For example, at the east end of Central Valley a prominent depositional break corresponding to a mappable ice margin has been identified. Immediately upslope are meltwater channels incised into weathered bedrock, marking ice margins parallel to those defined by moraines in the lower valley. This depositional break may have temporal significance, or may relate to the dynamic characteristics of the glacier, reflecting changes in basal thermal regime or distribution of sediment within the ice.

The limit of the last glaciation at any site must be defined collectively on the basis of i) glacial geomorphology, ii)relative sea level history, and iii) absolute chronology. By utilizing all these criteria for the east end of Central Valley it is suggested that the outermost moraine closely corresponds to the last ice limit, as depicted on a paleogeographic map for 9500 BP (Fig. II.8). A discontinuous moraine that loops into the lower valley is defined by three segments with maximum elevations of ca. 105, 103 and 91 m asl over ca. 1 km, indicating that the ice had a gentle surface slope near its terminus. The glacier likely terminated in a tidewater front, as the grounding bank extends to ca. 91 m asl, whereas the local marine limit is 97 m asl. Distal to the moraine a thick wedge of glaciomarine diamicton contains shell fragments dated 30,440 BP, providing a maximum date for attainment of the ice limit (Fig. II.8). Similar sequences of glaciomarine sediments are not found elsewhere in the valley. Inside the moraine loop the local marine limit is an ice-contact delta at 92 m asl, deposited in a basin formed behind the moraine with retreat of the glacier. The date of 8860±60 BP (TO-270) overlying glacial diamicton on the south side of the valley provides a minimum estimate of initial ice retreat. Although this date relates to the retreat of a tongue of ice occupying Snowbound Valley, rather than Central Valley, the two glaciers were likely coalescent in the lower valley prior to this transgression (Fig. II.8).

The ca. 8900 BP date demonstrates that we are dealing with an ice margin related to the last glaciation, and there is no evidence to

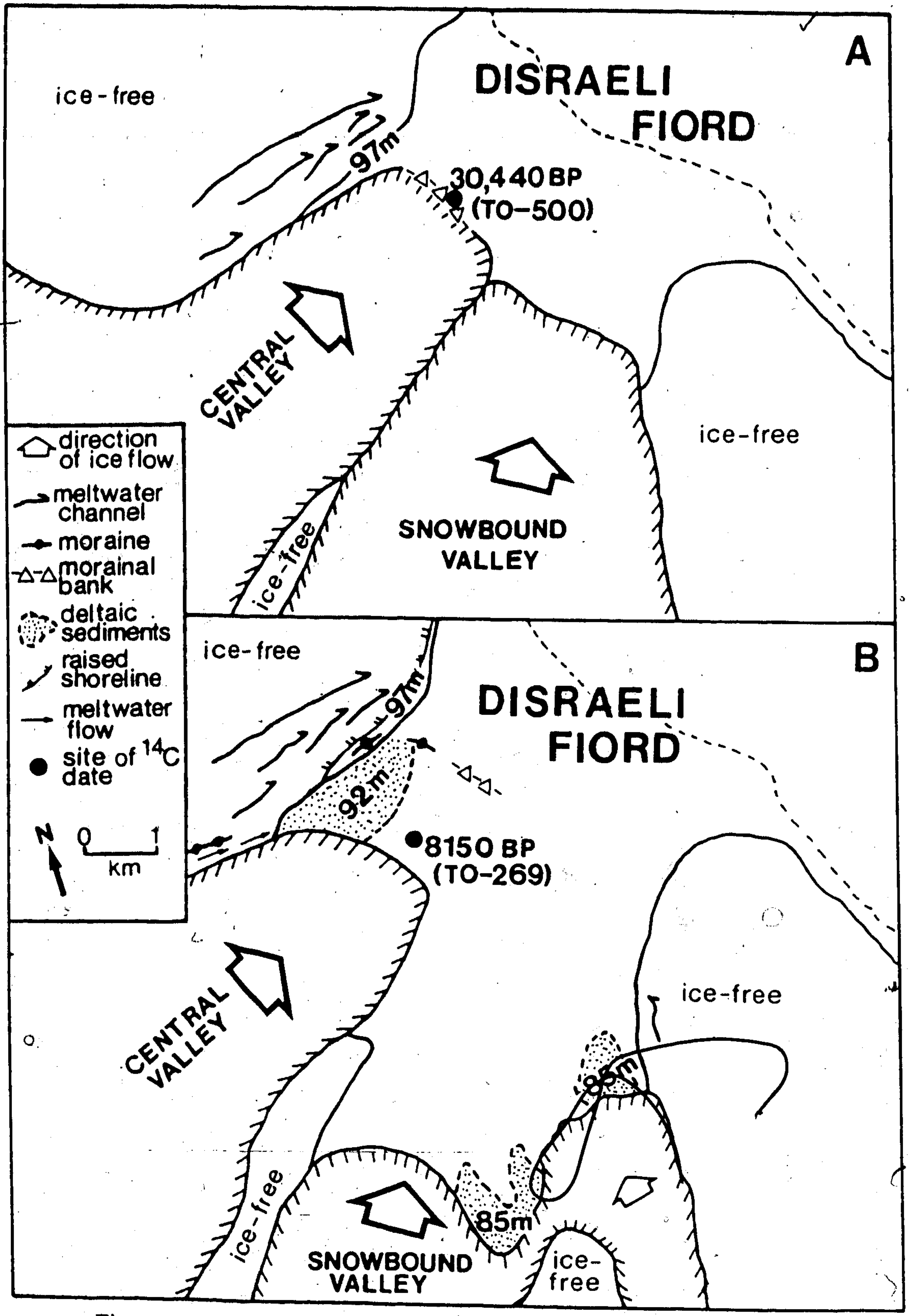

Figure II.8. Schematic diagram of the paleogeography of the east end of Central Valley for ca. 9500 BP (A) and ca. 8000 BP (B).

suggest that this position represents a recession from a more extensive limit in outer Disraeli Fiord (beyond the limit shown in Fig. II.8). If this were the case, one would expect to find equally abundant glaciomarine sediments and ice-contact deposits dating >8900 BP distal to this ice margin. However, these were not observed. Additionally, the abrupt decline in local marine limit observed on the proximal side of the moraine at the east end of Central Valley should also be found out Disraeli Fiord if a more extensive trunk glacier had retreated to the moraine, allowing successively younger and lower transgressions to follow it (cf. Andrews 1970). To the contrary, the marine limit north of Central Valley is generally smoothly inclined, suggesting that it represents a shoreline undisrupted by retreating ice. With respect to terrestrial deposits, erratics found above the limit of the outermost moraines in Central Valley have considerably enhanced micro-relief compared with the same lithologies inside the moraines. As no quantitative studies of weathering rates have been conducted in this area I am reluctant to use this as a criteria for defining a temporal boundary. However, the relationship is supportive of the other evidence arguing against a more extensive ice limit. Finally, the idea that the outermost moraines at the east end of Central Valley record the last ice limit is reinforced by the extent of moraines at the west end of the valley, which indicate a symmetrical advance of ice from a common dispersal centre.

Application of the same criteria throughout the study area produces a pattern of ice limits with glaciers terminating near the mouths of the present valleys (Fig. II.9). In many of the major valleys, ice contacting the sea is recorded by marine limits lower than those from

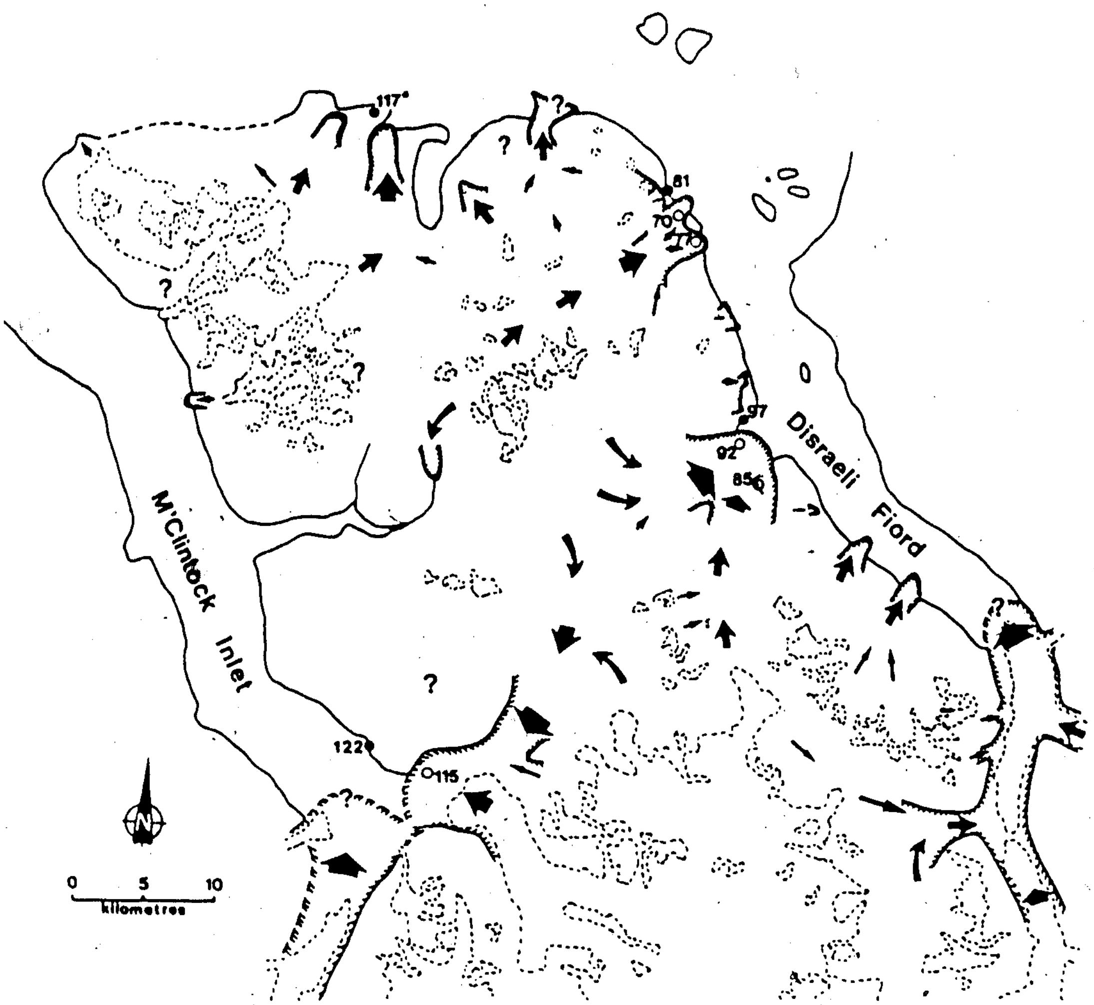

Figure II.9. Provisional map of the last ice limit in the field area. Arrows note direction and relative magnitude of ice flow. Elevations (m asl) contrast the marine limit inside (open circle) and outside (black circle) this ice limit. Asterisk denotes marine limit which predates the last glaciation. Dashed lines show the present glaciers and ice caps.

adjacent sites. For example, the marine limit in Disraeli and Ice Shelf creeks are 70 and 77 m asl, respectively, whereas a delta 2 km further north occurs at 81 m asl. This suggests that the moraines present along Disraeli Creek record recessional ice positions. Along the north coast, in the lower Sentinel Valley, the preservation of marine sediments (up to 117 m asl) which predate the last glaciation (Chapter III) unequivocally demonstrates that the last ice limit lay upvalley of this site (Fig. II.9).

For both inner Disraeli Fiord and M'Clintock Inlet the last ice limit is uncertain. Geomorphic evidence of ice extent is sparse, largely due to the steep and active fiord walls. Nonetheless, the decline of the marine limit in Disraeli Fiord, south of Central Valley, suggests that the major tributary valleys were deglaciated later than those further north (cf. Andrews 1970). It is not clear, however, whether this relates to retreat of a trunk glacier occupying the fiord or simply local ice debouching from side valleys. The best evidence for a trunk glacier in the fiord is provided by lake sediments in the lower Thores River Valley as well as striations and glacially molded bedrock along the west shore of the inner fiord. Unfortunately, for much of this area chronological control is unavailable. Where dates are available there is evidence of marine fauna in the innermost fiord by 8100 BP, associated with relative sea levels of <85 m. This presents apparent anomalies in the relative sea level data which are addressed at a regional scale in the following section.

Postglacial emergence

Establishing detailed emergence curves for the study area is not presently possible due to the sparse chronological control, particularly below marine limit. This is a result of two factors: i) the absence of an adequate sediment supply to the fiords since deglaciation, precluding the deposition of younger deltas; and ii) the general scarcity of driftwood, and complete absence of wood dating <3000 BP. Radiocarbon dates on driftwood are commonly used as control points for constructing emergence curves in arctic regions (cf. Blake 1975). Despite these limitations it is possible to reconstruct emergence patterns given two assumptions: i) that no significant emergence occurred prior to ca. 10,000 BP; and ii) that the elevation of the sea prior to initial emergence (the full glacial sea, England 1983) is recorded by the marine limit at sites lying beyond the last ice margin. The first assumption is supported by radiocarbon dates on marine shells from inside this ice margin, the oldest of which is 9560±70 (TO-488). The second assumption is supported by the fact that the marine limit declines abruptly inside the last ice limit.

Given these assumptions, provisional isobases can be drawn on the 10,000 BP shoreline. Control points are provided at numerous sites north of Central Valley. Using the 97 m marine limit distal to moraines at the east end of Central Valley as a starting point, it is clear that this isobase must run south of Ootah Bay (90 m ml) and north of west Central Valley (122 m ml). Assuming a linear gradient between these points, the resultant trend of the isobase is roughly perpendicular to the axis of Disraeli Fiord. When extrapolated on a regional scale (Fig. II.10a), isobases drawn on the 10,000 BP shoreline are consistent with patterns observed further east (England and Bednarski 1986).

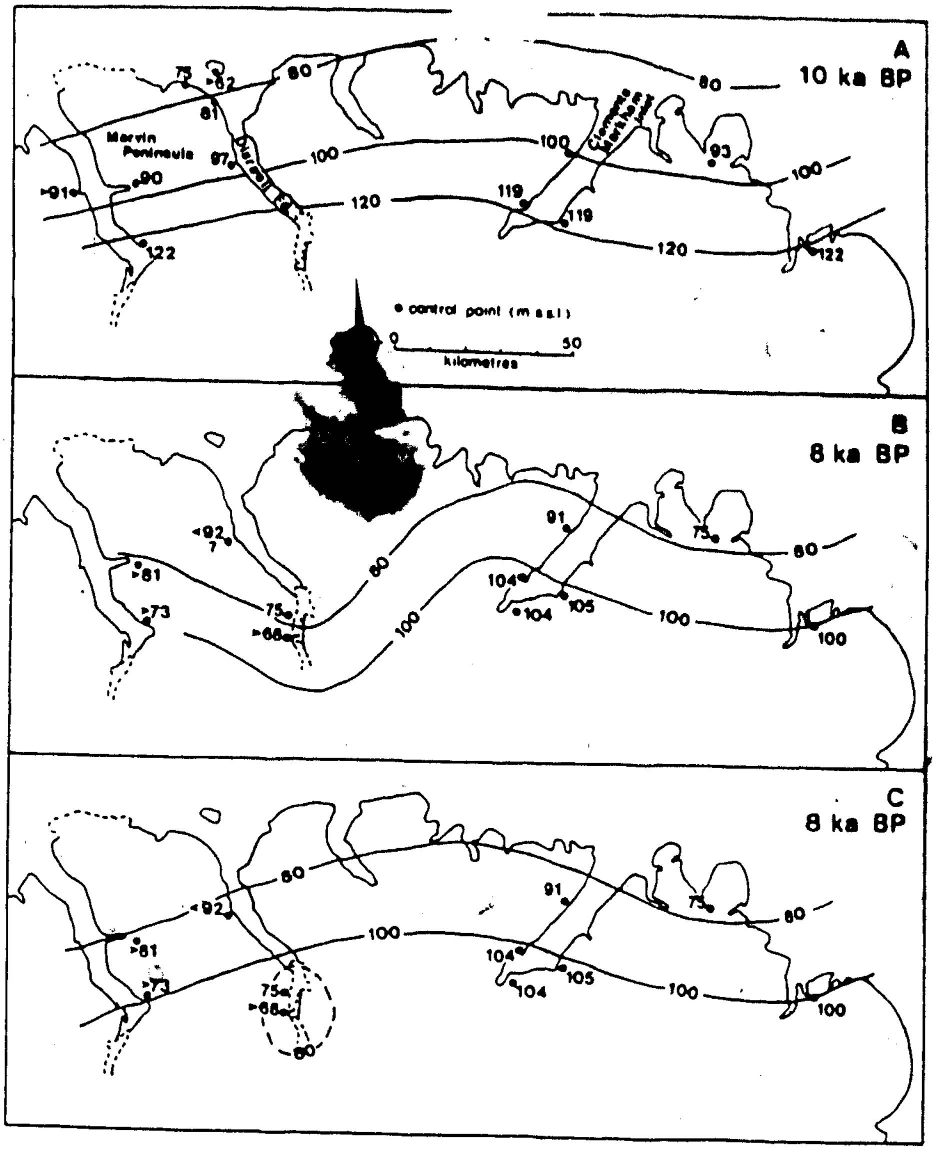

Figure II.10. Regional isobases on the 10.000 BP (A) and 8000 BP (B and C) shorelines. Data east of the field area are from England and Bednarski (1986). See text for discussion of B and C.

Along outer Disraeli Fiord the 10,000 BP shoreline is observed to rise upfiord at roughly 1.0 m km^{-1}. South of Central Valley, where glaciers displaced the 10,000 BP shoreline at many sites, the pattern of isobases for this interval is unclear. Extrapolation of the gradient found in the outer fiord indicates a 10,000 BP sea level for Thores River at ca. 127 m asl. This value is a conservative estimate because the profile of any shoreline will be curvilinear toward the centre of the former ice load (Andrews 1970).

Data are available regarding the height of the 8000 BP shoreline within inner Disraeli Fiord. At Thores River the marine limit of >68 m asl dates 8010±100 BP (TO-491) whereas 5 km further north, at Collision Delta, four dates on a 75 m delta range from 7800±60 BP (TO-520) to 8150±80 BP (TO-496; Fig. II.2). The implication here is that both the 80 and 100 m isobases on the 8000 BP shoreline must be located south of Thores River (Fig. II.10b). Elsewhere in the field area strong stratigraphic control for the 8000 BP shoreline is unavailable. Shells dated 8150±60 (TO-269) are believed to relate to a 92 m delta inside the last ice limit at the east end of Central Valley. To the west, near Ootah Bay, shells dated 7800±90 BP (TO-263) are interpreted as dating a delta at 81 m asl. These dates suggest that the 80 m isobase on the 8000 BP shoreline should lie north of both of these sites, trending WSW (Fig. II.10c), roughly parallel to the isobases drawn previously for the 10,000 BP shoreline. A third relevant date, 8140±90 (TO-499) on shells at 73 m asl at the west end of Central Valley, provides a minimum elevation for the ca. 8000 BP shoreline at that site and is consistent with the data from Ootah Bay and the east of Central Valley. It is impossible to resolve all the available observations into one simple isobase diagram. On a regional scale isobases on the 8000 BP shoreline from the east (England and Bednarski 1986) conform readily to the ca. 90 m sea level at the east end of Central Valley at this time (Fig. II.10c). However, the ca. 8000 BP dates from the inner fiord would require that the isobases bend sharply to the south (Fig. II.10b). This resulting pattern is inconsistent with the regional integration of a glacioisostatic load given a normal flexural parameter of ca. 180 km (cf. Walcott 1970), and does not seem plausible.

Compared to the rest of the field area, there appears to be a disproportionately small amount of emergence since 8000 BP in inner Disraeli Fiord. This is difficult to account for in terms of the glacial history. It might be suggested that the shells which date ca. 8000 BP at Thores River and Collision Delta relate to higher relative sea levels which were not recognized in the field. For example, England (1987b) discusses the occurrence of shells of full glacial age having lived beneath former ice shelves. These shells relate to a sea level about 35 m higher than the local marine limit, which was formed after the breakup of the ice shelf. England does, however, find evidence of the higher sea level in areas which remained ice-free adjacent to the tributary valleys where the shells were collected. However, in the case of Disraeli Fiord there is no geomorphic evidence to suggest that an ice shelf may have occupied the fiord head (cf. England <u>et al</u>. 1978; Sugden and Clapperton 1981; Lemmen and Evans 1987). Alternatively, it could be argued that the stratigraphy of the 8000 BP samples is correct and that their association with much lower sea levels reflects a preceding interval of significant emergence (at least 50 m) prior to ca. 8000 BP, occasioned by the retreat of grounded ice up Disraeli Fiord. Although initial emergence rates of up to 100 m/1000 yrs have been proposed for Greenland (Washburn and Stuiver 1962; Clark 1976), Bednarski (1986) reports less than 15 m of emergence prior to 8000 BP in inner Clements Markham Inlet (Fig. II.1). Similarly, data from the north coast of the field area show only limited emergence prior to 8600 BP, while dates from the east end of Central Valley suggest as little as 5 m of emergence prior to 8150 BP. Because the stratigraphic interpretation of the ca. 8000 BP dates from inner Disraeli Fiord appears correct, and because there is no evidence of significant glacioisostatic unloading prior to this time elsewhere in the field area, it may be necessary to conclude that the observed pattern of postglacial emergence is not simply a product of glacial isostasy. Neotectonics may also be an important factor (cf. England 1987a).

<u>Glacial style and paleoclimate</u>

During the last glaciation existing glaciers advanced up to 20 km, while presently empty cirques and ice-free plateaus were occupied by ice in response to the lowering of the local glaciation level. Based upon the maximum elevation of lateral moraines at either end of Central Valley, the paleoequilibrium line altitude (paleo ELA) in this area was about 205 m asl, which represents a lowering of 300 to 400 m from present ELA's (cf. Miller et al. 1975). This would put the paleo ELA <100 m above full glacial sea level, such as occurs along the northernmost coast today. In Central Valley, where the present divide lies <300 m asl and there is only a limited highland component to the drainage basins, this low paleo ELA allowed accumulation of ice within the valley itself.

Despite the substantial lowering of the regional glaciation level during the last glaciation, ice extent on Marvin Peninsula remained limited. This indicates that aridity was as great, or greater, than it is at present (cf. England 1986). Given these constraints of low temperatures and severe aridity, several meteorological factors operating in the area today probably combined to control the last ice limit. These include; i) wind drifting of snow - which enhances accumulation in protected sites; ii) fog - which reduces summer ablation at low elevations and iii) summer temperature inversions - which diminish ablation at low elevations.. In addition to these general factors, the physiography of Marvin Peninsula itself produces climatic constraints on ice extent. For example, the large ice-free area which existed inland from Ootah Bay throughout the last glaciation likely represented a rain-shadow in the lee of the peaks to the northwest, the highest terrain (1100 m asl) north of Central Valley. Although the extent of ice in the uplands between Ootah Bay and the west end of Central Valley is unknown, it may have constituted a similar rain shadow. In contrast, the development of extensive ice within Camp Creek, Disraeli Creek and Central Valley (Fig. II.9) reflects favorable exposure to precipitation from the northeast. This does not infer a significant change from present weather patterns (Maxwell 1982), but does imply that summer precipitation occurred primarily as snow.

<u>Ice build-up and refugia</u>

Few data are available regarding the onset of the last glaciation. The date of 30,440±330 BP (TO-500) on marine shell fragments within glaciomarine diamicton at the east end of Central Valley provides a maximum age for establishment of that ice margin, assuming that this represents an accurate, finite date (cf. Chapter III). Redeposited terrestrial organics dated 31,360±400 BP (TO-492), found in a section along the Thores River, were probably also transported by a glacier, and deposited in a high energy glaciolacustrine environment. Finally, organics dated 23,340±430 BP (TO-489) in the Sentinel River valley have also been redeposited at a much younger date, as they overlie marine silts dated 9080±110 BP (TO-487). These dates on redeposited organics indicate that plants grew in the study area at these times and, assuming that they have been glacially transported, the ice advance must have occurred subsequently. Therefore, it is likely that the last glacial maximum was attained <25,000 BP.

It is also likely that some vegetation persisted beyond the ice limit throughout the last glaciation. This is evidenced by the two samples from the west end of Central Valley dated 14,730±120 BP (TO-498) and 15,710±180 BP (TO-497), and the sample from Ice Shelf Creek dated 11,340±70 BP (TO-857). Although wood providing infinite dates, commonly interpreted to be Tertiary, has been found along Disraeli Fiord (Mielke and Long 1969; Blake 1987), it is unlikely that the organic dates in this study have been contaminated by inclusion of such older material, as only clean, subfossil bryophytes were submitted. The oldest of these samples (15,710 BP) was found within sand which stratigraphically underlies marine silt dated 8870 BP, and therefore the date on the organics is a reasonable estimate of the time of deposition of these sediments. The 11,340 BP organics were collected 16 m below the marine limit at Ice Shelf Creek, and were associated with marine shells dated 9250 BP. Because these shells are <u>in situ</u> and the organics are allocthanous, the dates are consistent and indicate that nonvascular vegetation was present in the drainage basin at least 2000 years before local deglaciation. There is no evidence of significant ice retreat prior to ca. 9500 BP in the study area, and therefore it is unreasonable to suggest that these plants migrated into the site following regional deglaciation. Therefore, it is concluded that bryophytes existed locally in ice-free areas throughout the last glaciation. It is noteworthy that the species composition of the bryophytes samples indicates a climate not radically different from that of the present (C. LaFarge-England, personal communication 1988), an observation consistent with the conclusions regarding full glacial climate presented above.

Deglaciation

The oldest date recording the onset of deglaciation in the field area is 9560±70 BP (TO-488) from Sentinel River (Fig. II.2). Elsewhere, dates on initial ice retreat range from 9250 BP (Ice Shelf Creek) to 8000 BP (Thores River). Ice retreat may have occurred in response to: i) a climatic amelioration, and/or ii) a rise in relative sea level, which would destabilize the marine component of glaciers (cf. Thomas 1979). Evidence of an early Holocene climatic amelioration includes: i) sea-ice push features and well developed beaches throughout the field area at or near marine limit; ii) driftwood dated 8850±50 BP (GSC-4559) along the north coast; and iii) marine shells dated 8630±70 (TO-861) inland of the Cape Discovery Ice Rise; all of which reflect more open water during initial deglaciation and preclude the existence of an ice shelf along the north coast at that time. In contrast, there is no evidence of a marine transgression at sites which lay beyond the last ice limit.

It appears that initial glacioisostatic unloading, a function of ice thinning and retreat, was slow. This is reflected in the amount of emergence recorded by the marine limit on either side of dated ice margins. Such data are available from three sites: i) Ice Shelf Creek, where 5 m of emergence occurred prior to 9250 BP; ii) the west end of Central Valley, where 7 m of emergence occurred prior to 8870 BP; and iii) the east end of Central Valley, where only 5 m of emergence is evident prior to 8150 BP. Although emergence at the east end of Central Valley is anomalously low, these amounts demonstrate very slow rates of initial emergence until ca. 8000 BP. Subsequently, ice retreat was apparently rapid. Dates on marine shells collected near the margins of glaciers indicate that ice had retreated upvalley of present positions by 7770±70 BP (TO-262) in the west end of Central Valley and by 7200±250 BP (L-248B, Crary 1960; Christie 1967) along outer M'Clintock Inlet. Driftwood dates demonstrate that the head of Disraeli Fiord was free of glaciers by 6000 BP. Glaciers at the fiord head have since readvanced beyond the site of this driftwood sample (Fig. II.2).

Discussion

The style of the last glaciation on Marvin Peninsula parallels the conditions reported from other localities on northern Ellesmere Island (England 1978, 1983; England et al. 1978; Bednarski 1986; Retelle 1986; Evans 1988). Limited ice coverage is evidenced by glacial geomorphology, relative sea level history, preservation of sea levels which pre-date the last glaciation, and radiocarbon dates of full glacial age on terrestrial organics. During the last glaciation, ice advanced <5 km beyond present margins where it presently terminates in or near the sea. Where ice presently terminates well inland, glaciers advanced up to 20 km beyond present margins. Ice growth involved not only the expansion of existing glaciers, but also accumulation in plateau and lowland areas that are presently ice-free. For example, glaciers completely filled the 30 km length of Central Valley, with much of the accumulation occurring within the valley on a site that is ice-free today. These observations suggest a similar style of glaciation to that proposed for west-central Ellesmere Island by Hodgson (1985, model C).

At all localities ice terminated rapidly upon contact the with sea, as evidenced by abundant glaciomarine sediments at the mouths of most valleys but not elsewhere along the fiords. Stewart (1988) concludes that such extensive raised marine sediments are reliable indicators of the last ice limit in the fiords of the high arctic. These ice-free fiords were occupied by the full glacial sea (England 1983), but it should be noted that unequivocal dates of full glacial age on marine fauna outside the last ice limit were not found in this study. This is a problem which persists at the regional scale (England 1987c). Calving was likely an important ablation process within the full glacial sea, especially if the activity index of the glaciers was similar to the low values of the present glacioclimatic regime.

Field evidence indicates that Disraeli Fiord and M'Clintock Inlet were not filled with grounded ice during the last glaciation. In addition, there is no evidence of non-local ice on Ward Hunt Island during this time. Thus the local model of ice coverage proposed by Lyons and Mielke (1973) is rejected, as are regional models of pervasive ice cover (Blake 1970; Hughes et al. 1977; Mayewski et al. 1981). Lyons ~~and Mielke (1973) cite~~ undatable erratics and striations as evidence for extensive glaciation are and ascribe these features to the last glaciation based upon their acceptance of Blake's (1970) proposal for an Innuitian Ice Sheet. Their observations are, however, consistent with a model of more extensive glaciation of the field area prior to the last glaciation (Chapter III). The rebound curve published for Ward Hunt Island by Lyons and Mielke (1973) is meaningless because of the lack of stratigraphic control for the dated samples (England 1974). Unfortunately, the same problems were encountered in this study and therefore a reliable emergence curve remains unavailable.

Non-geomorphological studies also suggest that glaciers were limited in extent on northern Ellesmere Island during the last glaciation. Ice cores from the Agassiz Ice Cap on northeast Ellesmere Island indicate that ice cap divides were no more than 200 m thicker than today's during the last glaciation (Koerner et al. 1987), precluding a pervasive regional ice sheet. Furthermore, preliminary interpretations of Quaternary sediments recovered from the continental shelf bordering northwest Axel Hieberg Island do not support the concept of grounded glaciers in that area during the last glaciation (Mudie et al. 1988). The hypothesis that ice-free areas on Ellesmere Island served as biological refugia throughout the last glaciation has been previously discussed by Leech (1966) and Brassard (1971), and supported by the subsequent mapping of the last ice limit (England 1978, 1983;

Bednarski 1986). Further support for a biological refugia is provided by a date of 8415±135 BP (S-2501) on a caribou antler from Clements Markham Inlet (Stewart and England 1986). This date shows that caribou were at the northern limit of their range at the onset of local deglaciation, when the Laurentide Ice Sheet still occupied northern mainland Canada (Dyke and Prest 1987). Furthermore, it is suggested that four radiocarbon dates between ca. 10,000 and 25,000 BP presented by Volk (1980), from northwest Ellesmere Island, also likely record the existence of biological refugia. In this study five dates, ranging from ca. 11,000 to 32,000 BP, on subfossil bryophytes are presented. These provide stratigraphic and chronologic evidence that strongly support the refugium hypothesis, and it is therefore concluded that northern Ellesmere Island was biologically viable throughout the last glaciation.

Evidence presented in this paper suggests that the last glacial maximum was locally attained after 25,000 BP in the study area. Elsewhere on northern Ellesmere Island organic dates from beneath glacial diamicton of 30,250±1100 BP (S-2650, Bednarski 1987) and 39,270±640 BP (TO-485, Evans 1987) have been obtained. Dates on detrital organics within marine sediments of 28,100±380 BP (GSC 1656, England 1978), 25,300±580 (H 5622-5164, Volk 1980) and 23,850±850 BP (S-2140, Bednarski 1988) have also been presented. These dates suggest that glaciers on northern Ellesmere Island attained their last ice limit during stage 2 of the marine isotope record. This chronology of ice advance places constraints on the possibility of attaining glacioisostatic equilibrium prior to deglaciation, as has been suggested by England (1983, 1985) for northeast Ellesmere Island and northwest Greenland. The growing data base on past ice margins and related sea levels from northern Ellesmere Island provides an interesting framework for testing the applicability of existing geophysical models to this area.

Deglaciation of Marvin Peninsula had begun by 9560±70 BP (TO-488). This is very similar to the chronology reported from Clements Markham Inlet by Bednarski (1986), where the oldest date on the last ice limit is 9845 BP. Bednarski (1986) suggested that rapid postglacial emergence did not start in Clements Markham Inlet until ca. 8000 BP, while a date of 8150±60 BP (TO-269) from the east end of Central Valley suggests that a similar pattern of emergence occurred on Marvin Peninsula. Isobases along the north coast of Ellesmere Island show a regionally consistent pattern of emergence for ca. 10,000 BP. However, data for 8000 BP from inner Disraeli Fiord do not fit into this regional pattern. This apparent anomaly may be a product of neotectonics (England 1987a). Finally, it must be recognized that all reconstructions of postglacial emergence are constrained by the reliability of the chronologic control, and may reflect the limitations of radiocarbon dating as much as geophysical processes.

There is strong evidence for a marked climatic amelioration during deglaciation of the north coast, where driftwood dating 8850 BP (GSC-4559) was collected inland of the Ward Hunt Ice Shelf. This is about the same age as the oldest Holocene driftwood found previously in the Canadian high arctic, 8915±115 BP (S-2211), which was collected near the head of Clements Markham Inlet (Stewart and England 1983). Stewart (1988) concludes that deglacial sedimentation in Clements Markham Inlet was dominated by processes characteristic of temperate tidewater glaciers which produced large volumes of sediment-laden subglacial meltwater, in contrast to the present cold-based (subpolar) glaciers. A preliminary investigation of deep water sediments within Disraeli Fiord suggests that the same processes were dominant there during deglaciation (Chapter IV). Based upon geomorphic data, Evans and Lemmen (1987) suggested that the Holocene climatic optimum along the north coast occurred ca. 7000 - 9000 BP. Preliminary interpretations of ice core records from northern Ellesmere also suggest a significant climatic amelioration beginning ca. 10,000 BP (R.M. Koerner, personal communication, 1987). The fact that rapid ice retreat did not begin until ca. 8000 BP is consistent with a lag time in the glacial system of about 2500 years, as suggested by England and Bednarski (1986).

These observations demonstrate a consistent pattern in the timing and extent of major glacial events along much of the north coast of Ellesmere Island. They also serve to emphasize an apparently pronounced difference in the timing of deglaciation between the north and south sides of the Grant Land Mountains (England and Bednarski 1986). On northeast Ellesmere Island, England (1983) found no evidence of emergence until ca. 8000 BP, with significant ice retreat not beginning until ca. 6200 BP. As both areas received ice flowing out of the Grant Land Mountains, the observed differences cannot simply be a product of climate, but must also be a function of ice dynamics and topographic controls.

BIBLIOGRAPHY

Andrews, J.T. 1970. A geomorphological study of postglacial uplift with particular reference to Arctic Canada. Institute of British Geographers, Special Publication No. 2, 156 p.

Bednarski, J. 1984. Glacier fluctuations and sea level history of Clements Markham Inlet, northern Ellesmere Island. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

__________. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 23, 1343-1355.

__________. 1988. The geomorphology of glaciomarine sediments in a high Arctic fiord. Geographie physique et Quaternaire, 42, 65-74.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, 7, 634-664.

__________. 1975. Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada. Geografiska Annaler, 57A, 1-71.

__________. 1982. Terrestrial interstadial deposits, Ellesmere Island, N.W.T., Canada. American Quaternary Association, 7th Biennial Conference, Seattle, Washington. Program and Abstracts, p. 73.

__________. 1987. Geological Survey of Canada radiocarbon dates XXVI. Geological Survey of Canada Paper 86-7, 60 p.

Bovis, M.J., and Barry, R.G. 1974. A climatological analysis of north polar desert areas. In, Polar Deserts and Modern Man. Edited by, Smiley, T.L., and Zumberge, J.H. The University of Arizona Press, Tucson, Arizona, 23-31.

Brassard, G.R. 1971. The mosses of northern Ellesmere Island, Arctic Canada. Ecology and phytogeography, with an analysis for the Queen Elizabeth Islands. Bryologist, 74, 234-281.

Bradley, R.S., and England, J. 1978. Recent climatic fluctuations of the Canadian high arctic and their significance for glaciology. Arctic and Alpine Research, 4, 715-731.

Christie, R.L. 1957. Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin Northwest Territories. Geological Survey of Canada Paper 56-9, 40 p.

__________. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. Geological Survey of Canada Memoir 331, 79 p.

Christie, R.L.. 1967. Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic archipelago. Geological Survey of Canada Bulletin 138, 50 p.

Clark, J.A. 1976. Greenland's rapid postglacial emergence: a result of ice-water gravitational attraction. Geology, 4, 310-312.

Crary, A.P. 1956. Geophysical studies along northern Ellesmere Island. Arctic, 9, 155-165.

__________. 1960. Arctic ice island and ice shelf studies. Part II. Arctic, 13, 32-50.

Dyke, A.S., and Prest, V.K. 1987. Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. Geographie physique et Quaternaire, 41, 237-263.

England, J. 1974. A note on the Holocene history of a portion of northernmost Ellesmere Island. Arctic, 27, 154-157.

__________. 1976. Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: alternative models. Quaternary Research, 6, 185-202.

__________. 1978. The glacial geology of northeastern Ellesmere Island, N.W.T., Canada. Canadian Journal of Earth Sciences, 15, 603-617.

__________. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, 20, 895-917.

__________. 1985. The late Quaternary history of Hall Land, northwest Greenland. Canadian Journal of Earth Sciences, 22, 1394-1408.

__________. 1986. A paleoglaciation level for north-central Ellesmere Island, N.W.T. Arctic and Alpine Research, 18, 217-222.

__________. 1987a. Glaciation and evolution of the Canadian high arctic landscape. Geology, 15, 419-424.

__________. 1987b. <u>In situ</u> shells of full glacial age beneath former ice shelves. 16th Arctic Workshop, Boreal Institute for Northern Studies, University of Alberta, Edmonton, Alberta. Abstracts, 20-22.

__________. 1987c. The late Quaternary history of Hall Land, northwest Greenland: Reply. Canadian Journal of Earth Sciences, 24, 374-380.

England, J., and Bednarski, J. 1986. Postglacial isobases from northern Ellesmere Island and Greenland: new data. Geographie physique et Quaternaire, 40, 299-305.

England, J., Bradley, R.S., and Miller, G.H. 1978. Former ice shelves in the Canadian high Arctic. Journal of Glaciology, 20, 393-404.

England, J., Lemmen, D.S., Evans, D.J.A., and Bednarski, J. 1987. The Quaternary of northern Ellesmere Island. International Union for Quaternary Research, XIIth International Congress, Ottawa, Canada. Programme with Abstracts, p. 162.

Evans, D.J.A. 1987. The last glaciation of northwest Ellesmere Island. 16th Arctic Workshop, Boreal Institute for Northern Studies, University of Alberta, Edmonton, Alberta. Abstracts, 24-28.

_________. 1988. Glacial geomorphology and late Quaternary history of Phillips Inlet and the Wootton Peninsula, northwest Ellesmere Island, Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

Evans, D.J.A., and Lemmen, D.S. 1987. Indicators of Holocene climatic change from Wootton Peninsula/Phillips Inlet and Marvin Peninsula/Disraeli Fiord, Northern Ellesmere Island. International Union for Quaternary Research, XIIth International Congress, Ottawa, Canada. Programme with Abstracts, p. 163.

Frisch, T. 1974. Metamorphic and plutonic rocks of northernmost Ellesmere Island, District of Franklin. Geological Survey of Canada Bulletin 229, 87 p.

Hattersley-Smith, G. 1961. Geomorphological studies in north-western Ellesmere Island. Defence Research Board, Department of National Defence, Canada, Report No. Misc. G-5, 16 p.

_________. 1973. Ice shelf and Fiord ice problems in Disraeli Fiord, northern Ellesmere Island, N.W.T. Defense Research Establishment Ottawa, Technical Note no. 72-34, 12 p.

Hattersley-Smith, G., and other members of the expedition (A.P. Crary and R.L. Christie). 1955. Northern Ellesmere Island, 1953 and 1954. Arctic, 8, 3-36.

Hattersley-Smith, G., Fuzesy, A., and Evans, S. 1969. Glacier depths in northern Ellesmere Island: airborne radio echo sounding in 1966. Defence Research Board, Canada, Defence Research Establishment Ottawa Technical Note No. 69-6, 23 p.

Hodgson, D.A. 1985. The last glaciation of west-central Ellesmere Island, Arctic Archipelago, Canada. Canadian Journal of Earth Sciences, 22, 347-368.

Hughes, T., Denton G., and Grosswald, M.G. 1977. Was there a late-Würm Arctic ice sheet? Nature, 266, 596-602.

Jefferies, M.O. 1987. The growth, structure and disintegration of arctic ice shelves. Polar Record, 23, 631-649.

Koerner, R.M., Fisher, D.A. and Paterson, W.S.B. 1987. Wisconsinan and pre-Wisconsinan ice thicknesses on Ellesmere Island, Canada: inferences from ice cores. Canadian Journal of Earth Sciences, 24, 296-301.

LaFarge-England, C. 1988. The contemporary and Holocene moss assemblages of a high arctic plateau, southern Piper Pass, northern Ellesmere Island, N.W.T., Canada. M.Sc. thesis, The University of Alberta, Edmonton, Alberta.

Leech, R.E. 1966. The spiders (Areneida of Hazen Camp 81°49'N, 71°18'W. Quastiones entomologicae, 2, 153-212.

Lemmen, D.S., and Evans, D.J.A. 1987. Contemporary and ancient ice shelf environments, Disraeli Fiord, Ellesmere Island, N.W.T. Canadian Association of Geographers - Western Division, Annual Meeting, Simon Fraser University, Burnaby, B.C. Abstracts, p. 30.

Lyons, J.B. and Leavitt, F.G. 1961. Structural-stratigraphic studies on the Ward Hunt Ice Shelf. Air Force Cambridge Research Laboratories, Final Report Contract AF19(604)-6188, Bedford Massachusetts, 37 p.

Lyons, J.B., and Mielke, J.E. 1973. Holocene history of a portion of northernmost Ellesmere Island, Arctic, 26, 314-323.

Matishak, M.A., and Lemmen, D.S. 1987. Deglacial sedimentation in the Thores River Valley, northern Ellesmere Island, N.W.T. Canadian Association of Geographers - Western Division, Annual Meeting, Simon Fraser University, Burnaby, B.C. Abstracts, p. 33.

Mayewski, P.A., Denton, G.H., and Hughes, T.J. 1981. Late Wisconsin ice sheets of North America. In The Last Great Ice Sheets. Edited by Denton, G.H., and Hughes, T.J. Wiley Interscience, New York, 67-178.

Maxwell, J.B. 1982. The climate of the Canadian arctic islands and adjacent waters. Climatological Studies, No. 30, volume 2, 589 p.

Mielke, J.E., and Long, A. 1969. Smithsonian Institution radiocarbon measurements V. Radiocarbon, 11, 163-182.

Miller, G.H., Bradley, R.S. and Andrews, J.T. 1975. The glaciation level and lowest equilibrium line altitude in the high Canadian arctic: Maps and climatic interpretation. Arctic and Alpine Research, 7, 155-168.

Mudie, P.J., Hein, F.J., and Van Wagoner, N.A. 1988. Marine geology studies on the Canadian polar margin. Geological Association of Canada, Joint Annual Meeting, Program with Abstracts, p. A88.

Narod, B.B., Clarke, G.K.C., and Prager, B.T. 1988. Airborne UHF radar sounding of glaciers and ice shelves, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 25, 95-105.

Ommanney, C.S.L. 1982. Bibliography of Canadian glaciology. Bibliography no. 2: Ellesmere Island glaciers and ice shelves. National Hydrology Research Institute, Ottawa, 53 p.

Powell, R.D. 1981. A model for sedimentation by tidewater glaciers. Annals of Glaciology, 2, 129-134.

Retelle, M.J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. Canadian Journal of Earth Sciences, 23, 1001 -1012.

Stewart, T.G. 1988. Deglacial-marine sediments from Clements Markham Inlet, Ellesmere Island, N.W.T., Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta, 229 p.

Stewart, T.G., and England, J. 1983. Holocene sea-ice variations and paleoenvironmental change, northernmost Ellesmere Island, N.W.T., Canada. Arctic and Alpine Research, 15, 1-17.

__________. 1986. An early Holocene caribou antler from northern Ellesmere Island, Canadian arctic archipelago. Boreas, 15, 25-31.

Sugden, D.E., and Clapperton, C.M. 1981. An ice-shelf moraine, George VI Sound, Antarctica. Annals of Glaciology, 2, 135-141.

Thomas, R.H. 1977. Calving bay dynamics and ice sheet retreat up the St. Lawrence valley system. Geographie physique et Quaternaire, 21, 347-356.

Trettin, H.P. 1969. Geology of Ordovician to Pennsylvanian rocks, M'Clintock Inlet, north coast of Ellesmere Island, Canadian Arctic Archipelago. Geological Survey of Canada Bulletin 183, 93 p.

__________. 1981. Geology of Precambrian to Devonian rocks, M'Clintock Inlet area, District of Franklin (NTS 340 E, H) -preliminary geological map and notes. Geological Survey of Canada, Open File Report 759.

__________. 1987. Pearya: a composite terrane with Caledonian affinities in northern Ellesmere Island. Canadian Journal of Earth Sciences, 24, 224-245.

Volk, H.R. 1980. Records of emergence around Ooboyah Bay and Neil Peninsula in connection with the Wisconsin deglaciation pattern, Ellesmere Island, N.W.T., Canada: a preliminary report. Polarforschung, 50, 29-44.

Walcott, R.I. 1970. Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, 7, 716-727.

Washburn, A.L., and Stuiver, M. 1962. Radiocarbon-dated post-glacial deleveling in northeast Greenland and its implications. Arctic, 15, 66-73.

III. MULTIPLE GLACIATIONS AND RELATED SEA LEVEL ADJUSTMENTS, NORTHERNMOST ELLESMERE ISLAND, N.W.T.

Introduction

Despite considerable research over the past 15 years on the Quaternary history of northern Ellesmere Island (Fig. III.1) little is known about the chronology of glacial events prior to the last glaciation[1]. The best record of multiple glaciations and related sea level adjustments comes from the northeast coast bordering Nares Strait (Fig. III.1, England and Bradley 1978; England et al. 1978, 1981; Retelle 1986; England 1987). Elsewhere, extensive glaciation(s) is recorded by high elevation erratics, meltwater channels and striae (Christie 1967; Hattersley-Smith 1969; Lyons and Mielke 1973; England 1978; Hodgson 1985; Bednarski 1986; Chapter II). Because the last glaciation in this area was of limited extent, these high elevation features must relate to an older glacial event(s) (England 1976, 1978, 1983, 1985, 1987; Bednarski 1986; Retelle 1986; England et al. 1987; Evans 1988; Chapter II). However, stratigraphic evidence of this older glaciation(s) is largely absent (cf. England 1987).

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[1] The term "last glaciation" is used to refer to the last major ice build-up in the area, subsequently recorded by retreat from dated ice-contact deposits and the start of glacioisostatic emergence during the Holocene (cf. England 1978, 1983; Hodgson 1985; Bednarski 1986). The term does not embrace the entire period since the last interglacial (Chapter II). Furthermore, raised marine features associated with the older, more extensive, glaciation(s), which should be preserved beyond the last ice limit, have not been recognized in most areas.

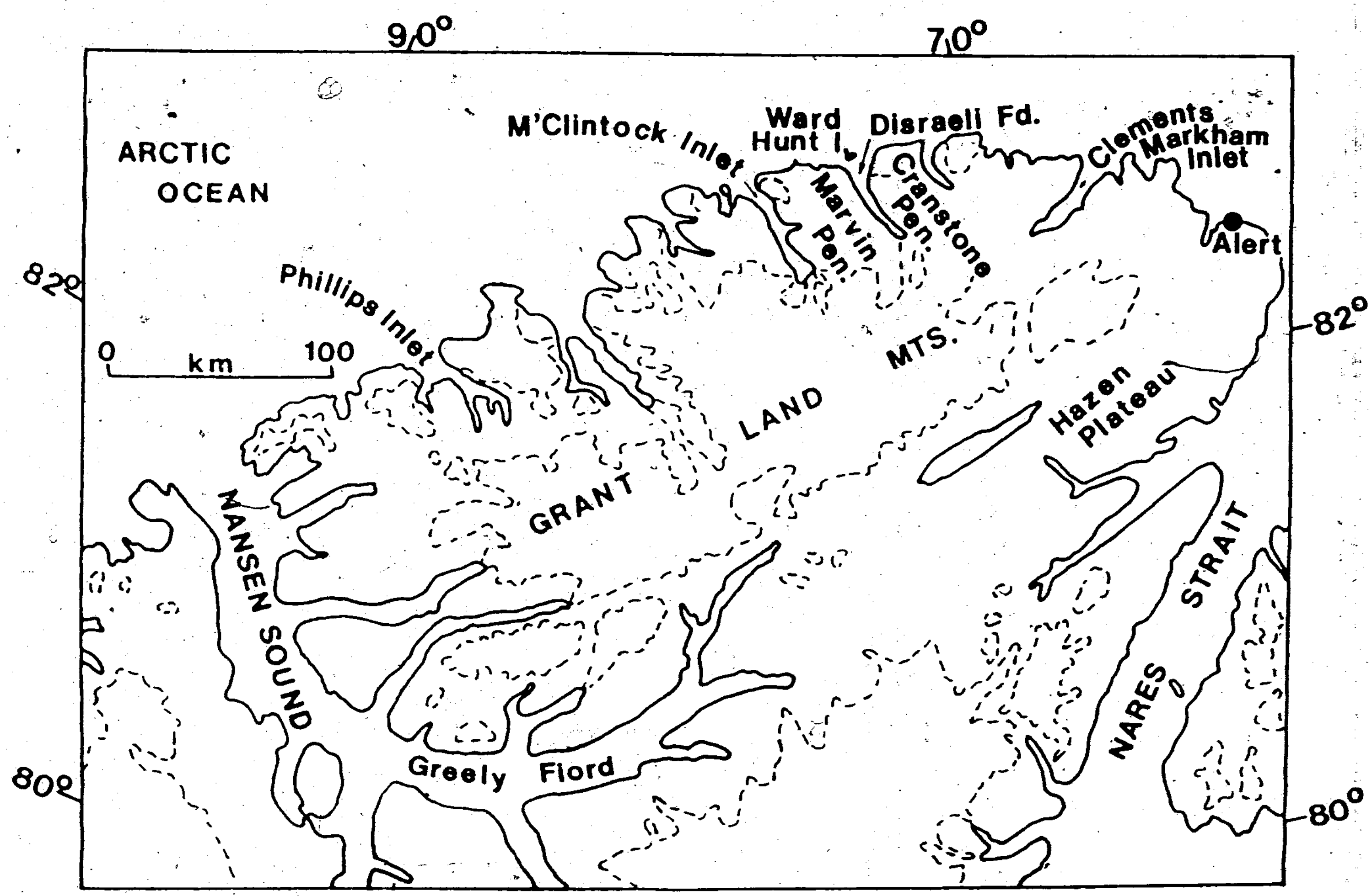


Figure III.1 Northern Ellesmere Island and official place names. Glaciers and icefields are outlined by dashes, the ice shelves are not shown.

This paper discusses glacial and sea level events recorded along the north coast of Ellesmere Island which predate the last glaciation (discussed in Chapter II). The present study provides further evaluation of ideas presented by England (1987) regarding the role of glaciation in the geomorphic evolution of the high arctic, as well as the relationship between glacial style, tectonics and sea level changes through time. Furthermore, it provides new field evidence regarding glacial history immediately adjacent to the Arctic Ocean Basin. This is of particular interest given previous attempts to correlate the glacial record with Arctic Ocean sediment cores (Clark et al. 1984; Dalrymple and Maas 1987).

Study area

The north coast of Ellesmere Island is a rugged fiord landscape bordering the Arctic Ocean Basin (Fig. III.1). Climatically, the region is a polar desert, with annual precipitation ranging from 2.5 to 13 cm a^{-1}, and a net annual water balance close to zero (Bovis and Barry 1974; Maxwell 1982). The mean annual air temperature at Alert (Fig. III.1) is $-18°C$. A steep climatic gradient characterizes the northernmost coast, with the local glaciation level increasing from about sea level along the Arctic Ocean to > 1100 m over the Grant Land Mountains to the south (Miller et al. 1975).

Field work for this study was conducted on Marvin Peninsula and Ward Hunt Island (Fig. III.2). Marvin Peninsula (ca. 2700 km^2) is bounded by Disraeli Fiord and M'Clintock Inlet to the east and west, respectively, which extend ca. 60 km inland from the Arctic Ocean. A

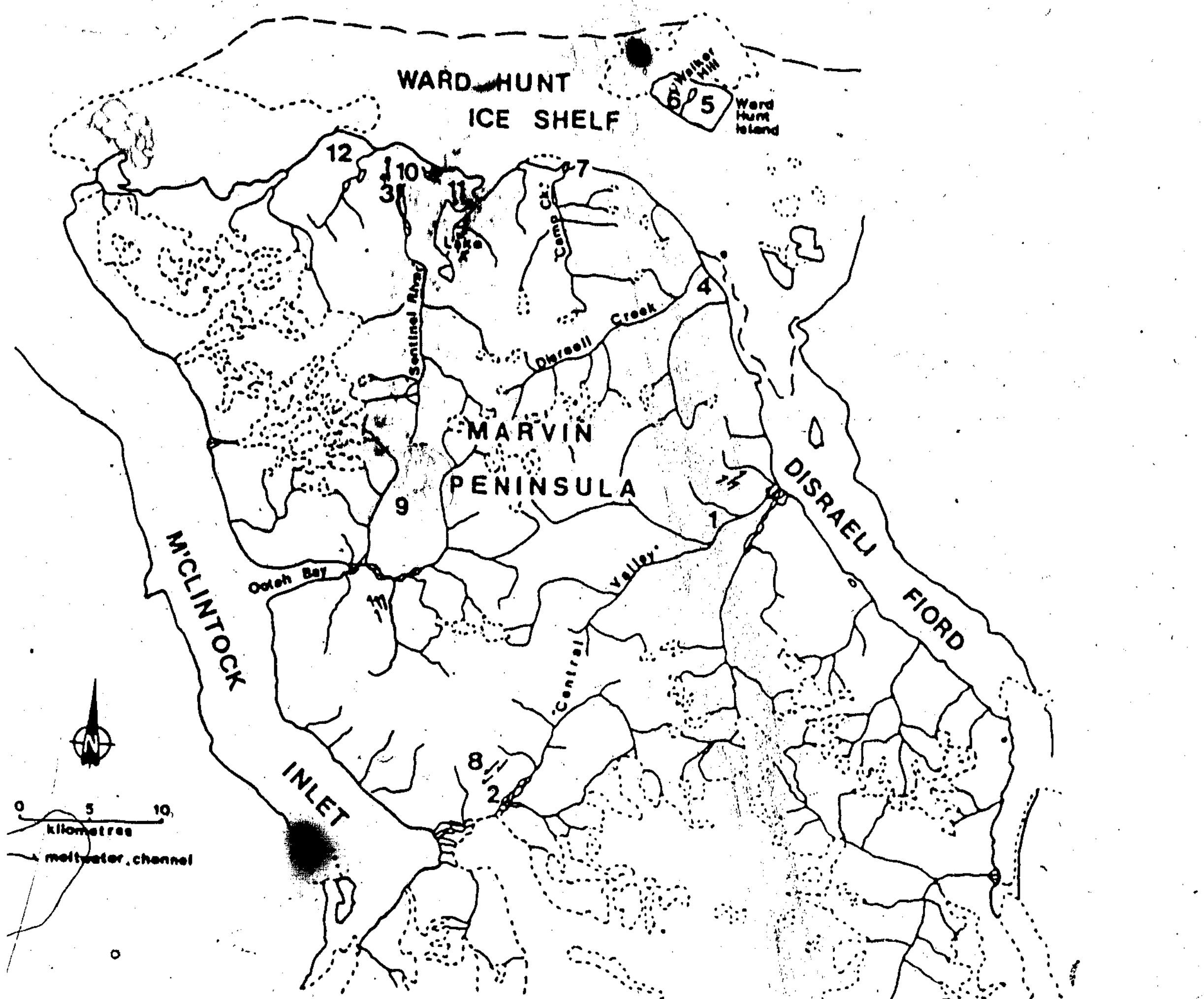


Figure III.2. Marvin Peninsula and Ward Hunt Island. Glaciers and ice rises are outlined by fine dashes. Names in quotations are unofficial. Numbers refer to sites discussed in text.

major trough, termed Central Valley, crosses the peninsula from NE to SW (Fig. III.2). South of Central Valley, the Grant Land Mountains and icefields reach 2500 m asl with ice thicknesses of 700 m (Hattersley-Smith et al. 1969; Narod et al. 1988). Large outlet glaciers descend from the icefields into the fiord heads, where they terminate as small glacial ice shelves. North of Central Valley summits up to 1100 m support small ice caps and cirque glaciers. The investigation of local Quaternary history focussed upon the numerous large valleys which radiate from the mountainous terrain of Marvin Peninsula and terminate at the sea (Chapter II).

Ward Hunt Island, which lies at the mouth of Disraeli Fiord, has an area of ca. 13 km^2. A low elevation valley trending NNE - SSW through the middle of the island separates Walker Hill (415 m asl) to the west and three lower summits (<250 m asl) to the east. The island is surrounded by the Ward Hunt Ice Shelf, which is grounded along its W and NW coast as well as offshore to the north (Fig. III.2).

Pre-Quaternary geology

A synthesis of the geological evolution of northern Ellesmere Island is presented by Trettin (1987a). The field area is composed of three major geologic units: i) Pearya, which in turn may be divided into four successions (Trettin 1987a); ii) the Franklinian Mobile Belt; and iii) the Sverdrup Basin (Fig. III.3). The distribution of these units is important with regard to Quaternary history, as some contain distinctive lithologies which may serve as diagnostic erratics.

The oldest rocks which outcrop in the field area are from succession II of Pearya (Upper Proterozoic to Lower Ordovician). Most

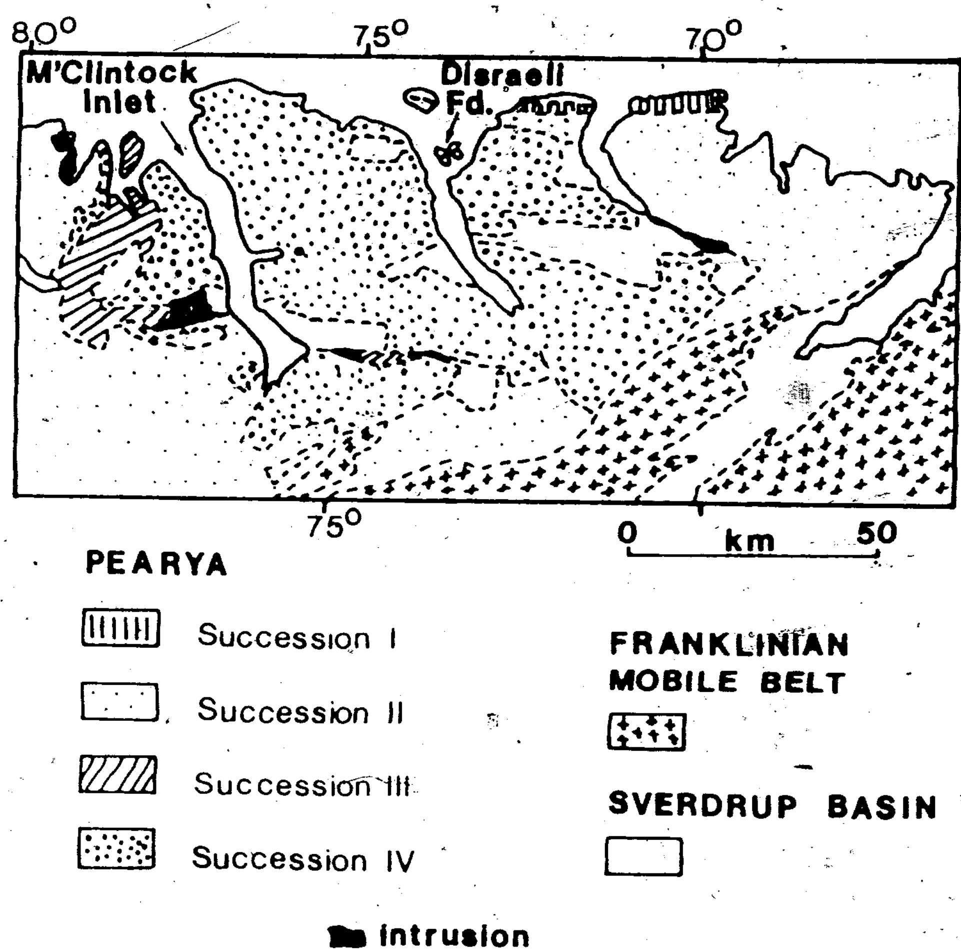


Figure III.3. *Major geological units of the north coast of* Ellesmere Island (modified from Trettin 1987a).

of Marvin Peninsula is composed of succession IV (Pearya), Middle Ordovician to Upper Silurian sedimentary and volcanic rocks (Fig. III,3, Trettin 1987a, 1987b). The only intrusive lithologies found on the peninsula are two small bodies of ultramafic to granitic rocks associated with succession III (Pearya). The youngest rocks which comprise a significant part of the field area are Upper Paleozoic. These sedimentary rocks represent part of the Sverdrup Basin, which was uplifted and deformed during the Eurekan Orogeny (Late Cretaceous to Early Tertiary; Thorsteinsson and Tozer 1970). Small outcrops of Tertiary conglomerates occur on ridgetops west and south of the head of M'Clintock Inlet, and are believed to represent syntectonic deposition associated with the Eurekan Orogeny (U. Mayr, personal communication, 1988). Late Tertiary rocks which may be correlative with the Beaufort Formation (Tozer 1956; Thorsteinsson and Tozer 1970) have not been observed in the field area.

Quaternary history - previous investigations

Evidence for regional glaciation at some unknown time was reported by investigators conducting reconnaissance studies along the north coast of Ellesmere Island (Hattersley-Smith et al. 1955; Crary 1956; Hattersley-Smith 1961; Lyons and Leavitt 1961; Christie 1967; Lyons and Mielke 1973). Based upon large-scale geomorphic features (U-shaped valleys, truncated spurs and hanging valleys), Hattersley-Smith (1961) suggested that the north coast fiords are the product of glacial over-deepening of preglacial valleys. However, Christie (1967) noted that although glacial striae and grooves tend to parallel major valley systems, the valley patterns reflect preglacial fluvial erosion with relatively minor glacial modification.

Recent detailed studies from Clements Markham Inlet (Bednarski 1986), ca. 100 km east of Marvin Peninsula, and Phillips Inlet (Evans 1988), ca. 200 km to the west (Fig. III.3), demonstrated that the last glaciation in those areas was of limited extent. Beyond the last ice limit, striae and sparse erratics (both rocks and shell fragments) are evidence of an older, more extensive glaciation in these areas. However, neither study found evidence of sea levels which might be attributed to the unloading of this larger ice cover; sea levels which should occur above the Holocene marine limit.

Evidence of extensive ice cover within the field area was reported by Hattersley-Smith et al. (1955), who found erratics up to 335 m asl on Ward Hunt Island and striae up to 760 m asl along the north coast of Marvin Peninsula. Crary (1956) interpreted shoals on the continental shelf, 15 to 20 km north of Ward Hunt Island, as moraines. Christie (1967) suggested that ice once flowed northward across the entire field area, but did not speculate about the age of this event. Based uponacceptance of Blake's (1970) model of an Innuitian Ice Sheet, Lyons and Mielke (1973) concluded that Disraeli Fiord was filled by a trunk glacier during the last glaciation and that this ice extended at least 16 km north of Ward Hunt Island to the vicinity of Crary's proposed moraines. However, as shown in Chapter II, during the last glaciation ice was of limited extent on Marvin Peninsula, precluding the advance of glaciers onto Ward Hunt Island from the mainland during this interval.

Geomorphology

Meltwater channels are found up to 400 m asl above the last ice limit on Marvin Peninsula. The most prominent channels are found in three localities: i) along the north wall of Central Valley; ii) within a tributary valley south of Ootah Bay; and iii) near the mouth of the Sentinel River (Fig. III.2). These channels occur within highly weathered bedrock, distinguishing them from fresh channels which are related to the last glaciation. These younger meltwater channels commonly occur parallel to, but downslope from, the older channels. Erratics of chert and quartzite pebble conglomerate within the older channels commonly show surface micro-relief of >10 mm due to the weathering of the fine-grained matrix. The same lithologies within deposits from the last glaciation show little to no surface micro-relief. Orientations of all meltwater channels show that ice flow was topographically controlled at the time they were formed.

The stratigraphic record for multiple glaciations in this area is limited. However, given the scarcity of thick glacigenic sediments related to the last glaciation (Chapter II), this is perhaps not surprising. A river bank section at ca. 65 m asl near the eastern end of Central Valley features ca. 2 m of interbedded sand and gravel overlain by 1.3 m of massive diamicton, which is interpreted as till (Fig. III.4; site 1, Fig. III.2). The entire section has been partially lithified with a carbonate-rich matrix. These sediments underlie, and border laterally, unconsolidated sands. Similar sediments are found at the west end of Central Valley up to 155 m asl (site 2, Fig. III.2). Although the stratigraphy is not as clear at this site, it is apparent

Figure III.4. Indurated, bedded, sand and gravel overlain by diamicton exposed along river cut at the east end of Central Valley. Scale = 1m.

that both the bedded and massive cemented units predate the last glaciation, as both occur commonly as erratics in surface till down valley. The relationship of these valley floor sediments (predating the last glaciation) to the "old" meltwater channels, or to high elevation erratics, is unknown.

Erratics are common beyond the last ice limit throughout the field area. They are usually highly weathered, and frequently buried within felsenmeer. However, owing to the area's structural and lithological diversity (cf. Trettin 1981), as well as the reconnaissance nature of the geological mapping, only a few lithologies are considered to represent diagnostic erratics that can be traced to known source areas. In general, erratics seem to indicate a south to north direction of transport. Fine-grained quartzite cobbles, likely from the Pearya succession II rocks which outcrop south of the fiord heads (H.P. Trettin, personal communication, 1988), were found within the "old" meltwater channels of outer Sentinel River, at 300 m asl south of Disraeli Creek, and up to 230 m asl on Ward Hunt Island (sites 3, 4, and 5, respectively, Fig. III.2). Sandstones, conglomerates and volcanics associated with Pearya succession IV (H.P. Trettin, personal communication, 1988), and which outcrop on northern Marvin Peninsula, are found up to 400 m asl on Ward Hunt Island (site 6, Fig. III.2). Greenschist, which outcrops on southern Ward Hunt Island, is found as erratics on the sedimentary rocks which form the north half of the island. Granitic boulders occur within the "old" meltwater channels of outer Sentinel River Valley, although they were not observed higher on the adjacent hillside. It is unlikely that these granites relate to the small intrusions which occur on Marvin Peninsula (H.P. Trettin, personal communication 1988), with a more likely source being the Cape Columbia Group which outcrops east of field area along the north coast (Christie 1964; Frisch 1974). A small granite pebble was also found on the surface of a kame terrace related to the last ice limit at the mouth of Camp Creek, near the elevation of the local marine limit (site 8, Fig. III.2).

Ten summits were ascended in the field area. On the highest of these (900 m asl; site 8, Fig. III.2), a subrounded cobble of quartz-muscovite schist was found among the limestone felsenmeer. On the northernmost peak (Walker Hill on Ward Hunt Island, 415 m asl; Fig. III.2) numerous erratics including volcanics, quartzite pebble conglomerate and greenschist were found to within 2 m of the summit. However, other summits feature spectacular tors and show no evidence of having ever been glaciated (Fig. III.5; site 9, Fig. III.2). No consistent pattern was observed to the uppermost elevation of erratics, which might relate to a former ice profile.

Sea-level record

Generally, few shorelines related to older glaciations have been recognized on northern Ellesmere Island (England 1987). At the mouth of the Sentinel River Valley (site 10, Fig. III.2) the maximum height of the Holocene sea level is recorded by prominent beaches with well-preserved sea-ice push ridges at ca. 68 m asl. This elevation is consistent with those of the highest deltas along the north coast of Marvin Peninsula, where deglaciation began >9500 BP (Chapter II). On a bedrock-controlled upland (sill) west of the lower river, marine

Figure III.5. Tors eroded in the volcanic flow rocks of the M'Clintock Formation, at ca. 750 m asl. Inset is of the same tors from a distance, showing mature periglacial slope. Erratics were not found among felsenmeer at this site. Location is site 9, Fig. III.2.

shorelines extend to at least 117 m asl. At ca. 93 m asl, a prominent wave-cut terrace extends laterally for about 100 m (Fig. III.6). The surface of this terrace, which is cut into bedrock, is highly oxidized and boulders exhibit well developed weathering pits or are highly frost-shattered. Thermal contraction cracks have developed within the bedrock at this site. To the east and upvalley of this terrace are a number of gravel cones and ridges which overlie finer sediments. This gravel surface extends to ca. 105 m asl, and shell fragments collected from a small ridge at ca. 100 m asl dated 30,010±300 BP (TO-486; site 1, Fig. III.7). About 200 m west of this site two complete valves of Astartae borealis were found on the surface of silty-sand at 117 m asl. A date of 27,170±210 BP (TO-860; site 2, Fig. III.7) was obtained from a single valve. Numerous large shell fragments were also found at the site. There is, however, no geomorphic evidence of sea level having extended above this elevation anywhere in the valley.

High elevation surfaces of probable marine origin are present at other sites along the north coast of Marvin Peninsula. Immediately east of Sentinel Valley the bedrock-controlled ridge north of Lake A (site 11, Fig. III.2) appears to be wave-washed to its crest at 86 m asl. Two large fragments of Hiatella arctica were found beside a bird perch at the highest point on this heavily polyganized terrain. No other shells were observed above 35 m asl in this area. Four km west of Sentinel Valley (site 12, Fig. III.2) a broad, level plain cuts across the geologic structure at ca. 110 m asl. The surficial sediment is dominantly fine grained and features well-developed thermal contraction cracks. Local relief of <10 cm over an area of >100 m² suggests

Figure III.6. Wave-cut terrace (arrow) at 93 m asl at the mouth of the Sentinel River Valley. Photo obscured by fog, which is typical of the northernmost coast.

SITE[1]	MATERIAL	SAMPLE[2] ELEVATION (m)	AMS[3] SAMPLE NO.	AMS DATE	alle/Ile[4] SAMPLE NO.	alle/Ile (free)	alle/Ile (total)
1	shell fragments	100	TO-486	32,010±300	UA-2452	0.133	0.023
2	shell fragments	117	TO-860	27,170±210	UA-2455	0.151	0.033
3	shell fragments	52	TO-859	27,560±200			
4	shell fragments	97-115	[illegible]-858	29,790±200	UA-2456	0.122	0.021
5	shell fragments	81	TO-500	30,440±330	UA-2454	0.780	0.208
6	whole valve *Hiatella arctica*	86			UA-2453	NA	0.058
7	shell fragments	136			UA-2458	0.170	0.034
8	shell fragments	150±			UA-2459	0.184	0.038
9	shell fragments	159			UA-2[illegible]	0.467	NA
9	shell fragments	234			UA-2457	0.210?	0.065
10	whole valves *Hiatella arctica*	46	TO-490	8630±100	UA-2461	0.367**	0.074**
11	whole valves *Hiatella arctica*	84	TO-267	8870±110	UA-2462	0.112	0.009

1 - site number refers to accompanying map

2 - all elevations determined using a Wallace and Tiernan altimeter and based on at least three independent readings from sea level. Values believed accurate to ±1.5 m.

3 - all AMS dates run by Isotrace Laboratory, University of Toronto. All samples were corrected to a base of $\delta^{13}C = 0‰$, equivalent to a reservoir correction of 410 years. The errors represent 1s.d.(68.3% confidence limits). All samples except TO-858 and TO-486 were given a 30 to 50% preleach.

4 - all alle/Ile ratios run by the Amino Acid Racemisation Laboratory, University of Alberta.

NA - ratio not available

** - sample was likely incorrectly labeled during preparation

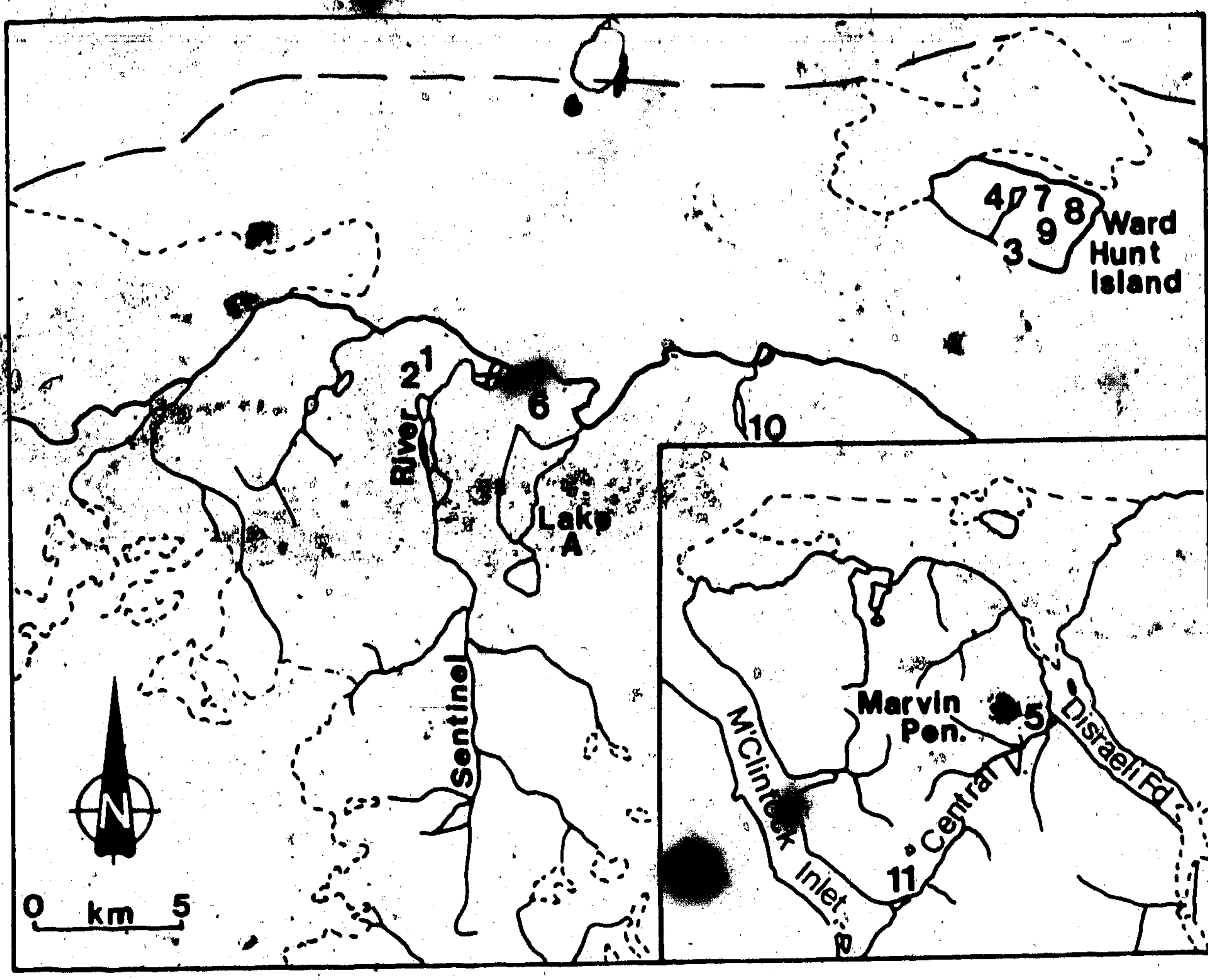


Figure III.7. Sites which provide chronologic control on glacial and sea level events which pre-date the last glaciation. Numbers refer to table at left. Sites 1[illegible] and 11 are Holocene, and used as comparison for aIle/Ile analysis.

planation by wave activity, but unequivocal evidence of a marine environment is absent.

On Ward Hunt Island steep slopes and small drainage basins make identification of marine surfaces above 45 m asl difficult. A minimum estimate of the Holocene marine limit on the south shore of the island is 62 m asl (Chapter II). Large shell fragments of Hiatella arctica found within silty colluvium at 52 m asl dated 27,560±200 BP (TO-859; site 3, Fig. [illegible]). Other shell fragments, found between 97-115 m asl among coarse talus on the east flank of Walker Hill, dated 29,790±200 (TO-858; site [illegible], Fig. III.7). Tiny fragments of marine shells, which may be erratic and not related to sea level history, were observed up to 23[illegible] m asl on eastern Ward Hunt Island.

Interpretation

The following discussion considers the evidence for multiple glacial and sea level events in the field area in three sections: i) the distribution of erratics; ii) geomorphic evidence from the Sentinel Valley; and iii) chronological evidence obtained through radiocarbon dating and preliminary amino acid racemization analysis.

Distribution of erratics

While erratics are common beyond the late limit throughout the field area few feature striations or facetted edges which unequivocally demonstrate that they relate to glacial transport. In contrast, most high elevation erratics which have not been shattered by frost processes tend to be subrounded in shape, cobble-size and smaller, with highly-weathered surfaces. Where erratics are not found in association with meltwater channels, striations, or other evidence of glaciation, the possibility that they are fluvial in origin cannot be discounted (cf. England 1987). For example, England (1974) interprets the highest erratics on the outer Hazen Plateau (Fig. III.1) as being fluvial in origin.

Fluvial transport. Clearly the interpretation of erratics found on summits as having a fluvial origin requires that they were transported across a landscape which was totally unrelated to the present physiography. As little is known about the mid-Tertiary to Holocene tectonic evolution of the high arctic landscape (England 1987), evaluation of the fluvial alternative is difficult. This is particularly true on Marvin Peninsula where syntectonic sedimentation related to this interval has not been recognized. However, one of the most curious physiographic features of Marvin Peninsula is believed to relate to the Tertiary drainage system in this area. Central Valley with its pass at <300 m asl, is clearly related to a similar valley which cuts across Cranstone Peninsula to the east of Disraeli Fiord (Fig. III.1), generally following the contact between the Ordovician volcanics and Upper Paleozoic sedimentary rocks. If Central Valley was part of the Tertiary drainage system it is noteworthy that the valley trends generally east-west, and has been truncated by the northerly oriented fiords. If the proposed source areas for many of the erratics are correct and reflect a south to north transport across Central Valley, then, for the erratics to be fluvial in origin they must predate both the formation of the fiord and the earlier establishment of this Tertiary drainage system.

Evidence from the field area supporting the interpretation of fluvial transport includes: i) the occurrence of erratics at elevations higher than any unequivocal glacial features; ii) the presence of tors and absence of glacial features on many summits; iii) the fact that the uppermost elevation of erratics does not appear to relate to a former ice profile; iv) the absence of high elevation moraines and meltwater channels similar to those found on NE Ellesmere Island, which are related to the most extensive glaciation of that area (e.g. England 1987); and v) the absence of shorelines or other marine deposits which relate to the demise of a pervasive ice cover. Again, these features are well preserved on NE Ellesmere Island.

Glacial transport. If the erratics have been glacially transported their regional distribution provides a record of ice extent and flow direction during a maximum glaciation. However, the redistribution of these erratics by subsequent glacial advances limits their usefulness for defining the extent of specific glacial advances. Additionally, it may be incorrect to use present topography when interpreting the maximum distribution of erratics, as the earliest glaciations may predate major tectonic modification of the high arctic landscape (England 1987).

Field evidence supporting the interpretation of glacial transport includes: i) the distribution of erratics, which reflects south to north transport parallel to the fiords and perpendicular to the proposed Tertiary drainage system, from a source area in the Grant Land Mountains; ii) the absence of significant deposits in the field area that represent syntectonic fluvial deposition associated with the Eurekan Orogeny or younger events; iii) the sparse occurrence of high elevation erratics, which does seem consistent with fluvial transport; and iv) the presence of tiny shell fragments up to 234 m asl on Ward Hunt Island. These fragments occur more than 100 m above the highest shorelines observed anywhere along the north coast of Ellesmere Island, and are therefore interpreted as erratics which could only have been transported by glaciers.

It is tentatively concluded that the high elevation erratics observed in the field area, including those on Ward Hunt Island, are primarily glacial in origin. Although meltwater channels were not measured above 400 m asl, and high elevation striations were not observed in this study, it should be noted that Hattersley-Smith et al. (1955) observed striations at 760 m asl along the north coast of Marvin Peninsula. This is not inconsistent with the glacial transport of erratics to 900 m asl at the fiord heads and 415 m asl on Ward Hunt Island. Erratics near the summit of Walker Hill suggest that ice overtopped Ward Hunt Island, and must have extended a considerable distance onto the continental shelf. Nonetheless, it must be emphasized that moraines and meltwater channels related to this pervasive ice cover are not observed, while an extensive advance of Greenland ice onto northeast Ellesmere Island is well recorded by such geomorphic features (England and Bradley 1978; England et al. 1981; Retelle 1986).

Of all the erratics observed in the field area the most difficult to account for are the granites at Camp Creek and Sentinel Valley. All examples observed occur below the marine limit (ca. 117 m asl) recorded within Sentinel Valley. If these erratics are derived from the Cape Columbia Group it is most probable that they were rafted by sea-ice or icebergs to the north coast of Marvin Peninsula. Similar low elevation erratics along the north coast of Greenland are considered to have been transported by glacial ice shelves flowing parallel to the coast. However, the possibility that they have been ice-rafted was apparently not considered (Dawes 1987).

Sentinel Valley

Sentinel Valley (Fig. III.8) is the only site in the field area where the relationship between glacial and sea level events can be documented, but the absence of an associated stratigraphic record hinders interpretation. Within the northwest part of Sentinel Valley, lateral meltwater channels, incised up to 10 m in bedrock, record the oldest definable ice margin. These channels descend to ca. 95 m asl, which serves as a maximum estimate of relative sea level at that time. This sea level is consistent with the greater isostatic load associated with an ice cover more extensive than that of the last glaciation, whose highest relative sea level is 68 m asl. The topography of the valley requires that the glacier responsible for the meltwater channels was pinned over the sill at the mouth of the valley, and grounded below sea level for a distance of at least 5 km upvalley (Fig. III.9a). It is possible that deltaic sedimentation related to discharge from these channels is marked by coarse rock-glacierized gravels which descend to sea level at the extreme NW corner of the valley.

The highest sea level recorded in the Sentinel Valley occurs inside the ice margin marked by the meltwater channels, and therefore it must postdate the channels. The fossiliferous sediments at 117 m asl are

Figure III.8. Oblique air photo of outer Sentinel Valley showing "old" meltwater channels (mwc), ca. 105 m asl delta surface (D, dashed line) and limit of Holocene beaches (H, dotted line). Meltwater channels related to the last glaciation are not present on this side of the valley. The wave-cut terrace in Fig. III.6 is just to the right (west) of this photo.

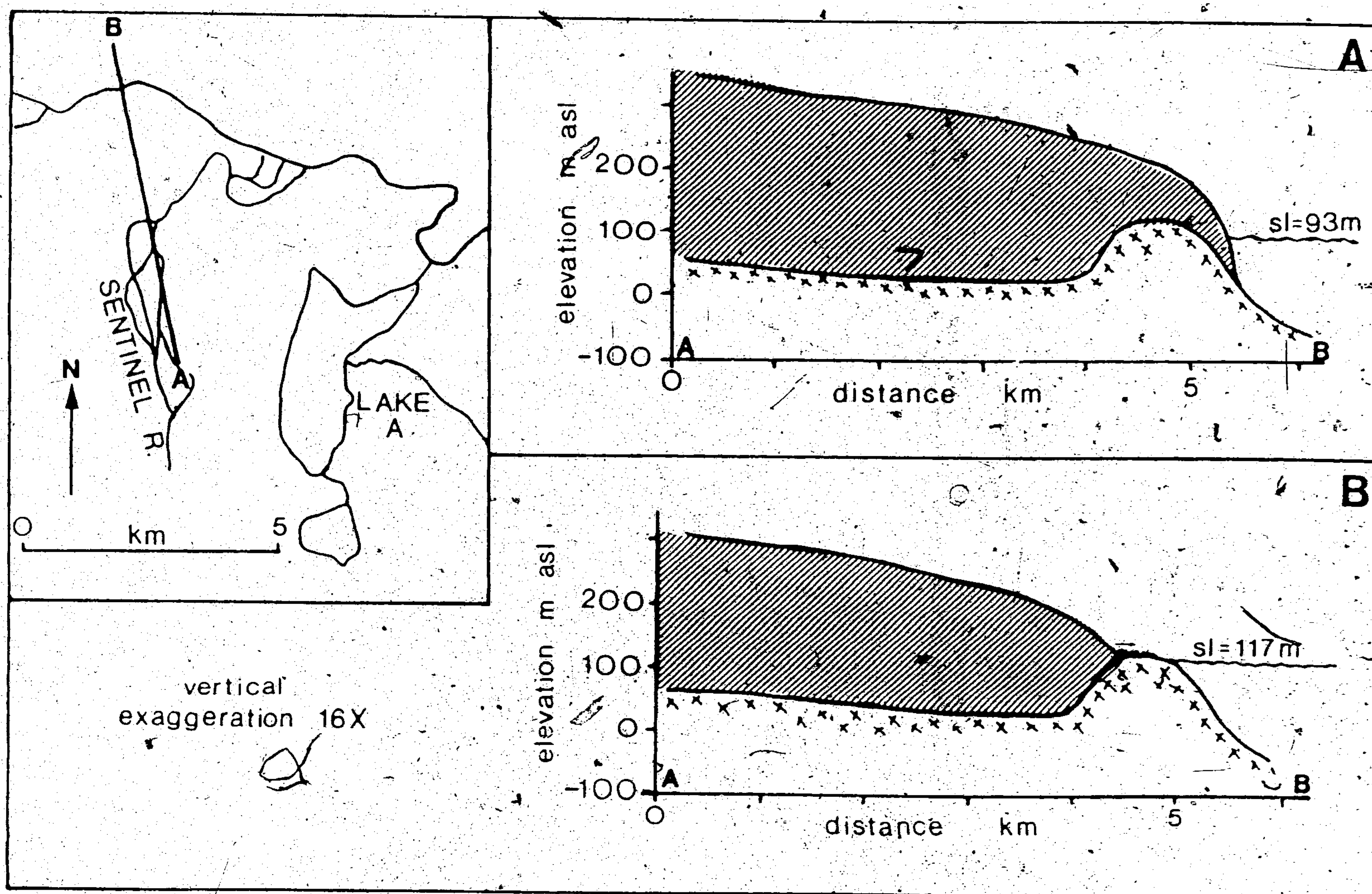


Figure III.9. Schematic cross-sections of ice extent in Sentinel Valley at time outer meltwater channels were cut (A), and transgression to 117 m asl as ice retreated (B). Delta at 105 m asl was deposited with further, ice retreat and the emergence of the 117 m shoreline.

interpreted as littoral deposits. A bedrock knoll immediately upvalley has protected these sediments from subsequent fluvial erosion. Within the main valley the gravel cones and ridges extending to ca. 105 m asl record an extensive marine delta that has been severely altered by weathering and slope processes. The distinct surface morphology of this delta is similar to that of altered marine terraces at the mouth of the Seegloo River, northeast Ellesmere Island, which has been radiocarbon dated at > 39 ka BP (see Figs. 5 and 6 in England et al. 1981).

Formation of the 105 m delta likely occurred when a glacier abutted the southern edge of the bedrock sill and was grounded below sea level within the main valley (Fig. III.9b). The alternative, that the entire valley was filled with deltaic sediment to >105 m asl which was then removed by subsequent glacial and fluvial activity, is considered unlikely. Additionally, the presence of a significant ice load allows for isostatic depression of the crust which, in the absence of tectonic uplift, is necessary to account for the high relative sea level. It is likely that this delta was deposited during the retreat of the glacier that previously produced the lateral meltwater channels. This necessitates a marine transgression of >20 m between the time that the meltwater channels were carved and the littoral sediments were deposited at ca. 117 m asl. This transgression may represent a rise in eustatic sea level, or simply record an increasing ice load from the Grant Land Mountains at a time when the glacier in Sentinel Valley was undergoing minor retreat.

Wave-cut terraces similar to the one observed at ca. 93 m asl in Sentinel Valley do not occur elsewhere in the field area, and are found only rarely on northern Ellesmere Island. The formation of this platform would require a period of reduced sea-ice severity (i.e. less than at present), and preclude the presence of an ice shelf along the north coast. It would also require a period of relative sea-level stability, although the length of this stable phase is unknown. England (personal communication, 1988) has found wave-cut terraces along Greely Fiord (Fig. III.1) associated with the full glacial sea, which remained stable for at least 2000 years (England 1983, 1987). Again, the elevation of the terrace requires an increased glacioisostatic load compared to the present, while prolonged sea-level stability implies that a state of isostatic equilibrium had been attained. The sea level related to the terrace is unknown, and if the platform is a product of wave erosion the associated sea level may be as much as 20 m higher (Fairbridge 1968). Therefore it is possible that this terrace relates to the same relative sea level as the 105 m delta.

Chronology

Degree of weathering is useful for distinguishing between deposits related to the last glaciation and those of older glacial events. However, beyond the last ice limit well defined weathering zones such as those described elsewhere in the arctic (cf. Pheasant and Andrews 1973; England et al. 1981) were not observed. The presence of high tors and thick felsenmeer on some summits of Marvin Peninsula indicates surfaces of considerable antiquity.

Of the forty radiocarbon dates on marine shells and shell fragments available from the field area, 85% are <10,000 BP (Fig. III.10). Five dates, all determined using accelerator mass spectrometry (AMS), ranging

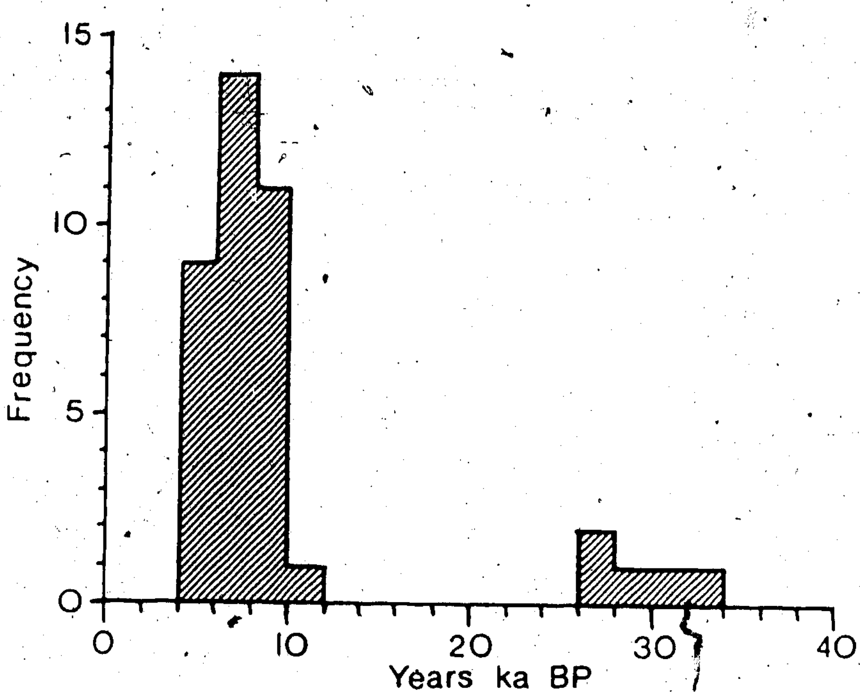


Figure III.10. Histogram of ^{14}C dates on marine shells from Marvin Peninsula and Ward Hunt Island. Dates are plotted without standard error.

from 27,170±210 BP (TO-860) to 32,010±300 BP (TO-486) have also been obtained (Fig. III.7). Interpretation of these "old but finite" radiocarbon dates is critical to understanding the timing and dynamics of high latitude glaciation. Owing to problems associated with altered carbonates, the limit for reliable radiocarbon dating of marine shells using conventional methods (ß - counting) is generally accepted to fall around 20 to 25 ka (Lowden 1985; Bradley 1985). It remains to be demonstrated whether AMS dates beyond this age range are reliable. It is clear that AMS dates are vulnerable to modern contaminants, owing to the small size of the sample used. The Geological Survey of Canada recommends that only samples anticipated to be younger than 30 - 35 ka be submitted for AMS dating (D. St-Onge, personal communication, 1988).

Three samples collected above the limit of the Holocene sea were dated by AMS in this study, and all fall within this zone of questionable authenticity. None of the shells were found <u>in situ</u>, and only the whole valves collected at 117 m asl in Sentinel Valley (27,170±210 BP, TO-860) can be confidently related to a relative sea level. Fragments collected from deltaic sediments within the same valley, and dated 32,010±300 BP (TO-486), may have been redeposited by glacial ice. Reworking of older marine shells has been demonstrated along Disraeli Fiord, where shell fragments dating 30,440±330 BP (TO-500; site 5, Fig. III.7) are found within a morainal bank deposit which records the last ice limit (Chapter II). The two dates of 27,560±200 (TO-859) and 29,790±200 (TO-858) on shell fragments from Ward Hunt Island are not believed to have been glacially reworked, based upon the size and abundance of fragments and the fact that they occur below the highest sea level recorded in Sentinel Valley. Using the same criteria, the tiny shell fragments found between 136 and 234 m asl on Ward Hunt Island are believed to represent erratics deposited by a glacier.

If these old, finite radiocarbon dates are accepted, then the date of ca. 27 ka BP in Sentinel Valley records a marine transgression to 117 m asl associated with local ice retreat. The date of ca. 32 ka BP from the same valley represents shells living during the transgression caused by this advance, and which subsequently were reworked by the glacier and redeposited in deltaic sediments during ice retreat. Dates from Ward Hunt Island lend additional support to the concept of a high sea level at ca. 27 - 30 ka BP. Combined with the data regarding the timing and extent of the last glaciation, one would conclude that there were both Middle and Late Wisconsinan glacial advances in the field area, the former being the more extensive.

Despite the consistency of the AMS dates, it is argued that the true age of these samples is likely beyond the range of radiocarbon dating for the following reasons.

A - The relative weathering of bedrock, erratics and marine surfaces which predate the last glaciation, in comparison to Holocene features, suggest that they they have been exposed to weathering processes for a considerable period of time. Features which predate the last glaciation are highly weathered and deeply oxidized, while the same lithologies inside the last ice limit appear fresh. The surface of the delta at 105 m in Sentinel Valley has been highly altered by weathering processes, in contrast to Holocene deltas and beaches which commonly feature well-preserved sea-ice push features. As noted previously, the morphology of this delta is similar to altered marine terraces on NE Ellesmere Island previously dated > 39 ka BP. While no quantitative data are available regarding rates of weathering in the area, it seems unlikely that the marked differences observed could be attributed to only a three-fold longer exposure to weathering processes.

B - Ratios of allo-isoleucine (aIle) to isoleucine (Ile) provide an independent relative dating technique on shell samples (cf. Miller et al. 1982; Rutter et al. 1985). Ten samples from this study were submitted for analysis, with preliminary results now available (Fig. III.7). These data show that two samples from the Sentinel Valley, which relate to the 117 m asl sea level, are of similar age to fragments collected between 97 and 115 m asl on Ward Hunt Island, and that all 3 samples of these samples are significantly older than Holocene shells collected along the north coast. All samples collected above 136 m asl on Ward Hunt Island have higher aIle/Ile ratios (both fractions) than the samples from the Sentinel Valley, again reflecting greater relative age. However, it is impossible to differentiate populations of widely varying age in the data, and hence it cannot be concluded that a number of distinct sea level (or glacial) events are recorded. The fact that all pre-Holocene samples collected in this study consisted of unidentified shell fragments adds considerable uncertainty to their interpretation, as ratios are known to vary within individual valves and by as much as 50% between shell species (J. Brigham-Grette, personal communication, 1988). Ratios on fragments from central Disraeli Fiord (Site 5, Fig. III.7) of 0.780 (free fraction) and 0.208 (total acid hydrolysate) appear highly anomolous. Further clarification is not possible until additional aIle/Ile analyses have been completed.

Discussion

Glacial and sea-level record

Glacial geomorphology and sea level history clearly record two distinct glacial advances in the field area. The most recent of these, the last glaciation, occupied most of the major valleys on Marvin Peninsula but terminated rapidly upon contact with the sea. Abundant geomorphic and chronologic evidence demonstrates that many areas remained ice-free throughout the last glaciation (Chapter II). In the Sentinel Valley an older and slightly more extensive glacial advance is recorded by "old" meltwater channels and associated high sea-levels. AMS radiocarbon dates, if accepted as being accurate, place this advance in the middle Wisconsinan, but relative weathering and preliminary aIle/Ile analysis suggest that it may be considerably older. A third glacial advance, recorded by high elevation erratics, has also been proposed. Constructional landforms, either glacial or marine, which may relate to this proposed event have not been recognized in the field area. However, erratic cobbles near the summit of Ward Hunt Island suggest that this oldest event involved a nearly pervasive ice cover. The age of this proposed glacial advance is unknown, but aIle/Ile ratios on high elevation erratic(?) shell fragments from Ward Hunt Island indicate a greater relative age than the outermost glacial advance recorded in the Sentinel Valley.

Evidence of multiple glacial and sea level events is available from other sites in the region. Along northeastern Ellesmere Island three discrete glaciations have been defined. The oldest of these involved the advance of Greenland ice onto Ellesmere Island, which is recorded by erratics, well-preserved moraines and abundant meltwater channels (England and Bradley 1978; England et al. 1978, 1981). Subsequent retreat of the Greenland Ice was associated with a relative sea level at 286 m asl (England 1985; Retelle 1986). aIle/Ile ratios on shells related to this high sea level suggest an age of 500 ka to 1 Ma (England 1987). A subsequent advance of Ellesmere Island ice cross-cuts the Greenland erratics, and in places contacted the sea at elevations up to 175 m asl. Radiocarbon dates on this sea level are >39 ka BP (England and Bradley 1978; England et al 1978, 1981). Further north along the same coast, Retelle (1986) reports radiocarbon dates of 31,300±900 and >32 ka on an indeterminant sea level between 90 and 286 m asl. Mean aIle/Ile ratios from these samples are 0.218±0.03 (free) and 0.063±0.011 (total), and suggest that the radiocarbon dates should be viewed as minimum age estimates only (Retelle 1986). Elsewhere in the region, Bednarski (1987) reports multiple tills above marine sediments dated 30,250±1100 BP on western Ellesmere Island. Geomorphic evidence of sea levels up to 170 m asl, and shell fragments to 280 m asl, are present along eastern Nansen Sound (Fig. III.1; Bednarski 1988).

Extensive fieldwork has demonstrated that the style of the last glaciation was similar throughout northern Ellesmere Island (England et al. 1987), therefore, it would seem reasonable to assume that the timing and extent of previous glacial events would also demonstrate regional similarity. The most extensive ice limit recorded in Sentinel Valley is of similar magnitude to the most extensive Ellesmere ice advance recorded on the northeast coast (England and Bradley 1978; England et al. 1978, 1981). In Sentinel Valley the 117 m sea-level lies ca. 50 m above the maximum Holocene shoreline whereas along northeastern Ellesmere Island the observed difference is ca. 55 m. ΔIle/Ile ratios may be interpretted as supporting correlation of these events, although further analysis is needed. The absolute age of this advance remains unknown.

The oldest glacial advance proposed in this study, involving the northward flow of a nearly pervasive ice cover from the Grant Land Mountains, is marked by high elevation erratics and possibly erratic shell fragments on Ward Hunt Island. Moraines, meltwater channels and sea level features associated with this proposed advance were not observed. Nonetheless, similar conclusions have been made by Bednarski (1986) working along the east coast of Clements Markham Inlet (Fig. III.1), where a high elevation erratic is felt to record an extensive ice cover with flow parallel to, but not diverted by, the inlet (Bednarski 1986). Reconstructive ice sheet modeling by Reeh (1982, 1984) indicates that even if ice were grounded offshore along the 600 m isobath, the Grant Land Mountains would still serve as a major ice divide. It is tempting to make a tentative correlation between the advance ~~of~~ Greenland ice on northeastern Ellesmere Island and the proposed pervasive glaciation of Marvin Peninsula and Ward Hunt Island. However, there are neither geomorphic nor chronologic data to justify such a correlation. Clearly, given our limited understanding of high latitude glacial history and the difficulty in making correlations on a regional scale, attempts to correlate between the terrestrial record and sediment cores from the Arctic Ocean (Clark et al. 1984; Dalrymple and Maas 1987) should be considered premature.

Tectonic influences

Trettin (personal communication, 1988) considers the geological evolution of northern Ellesmere Island to have been largely passive since the Oligocene. In contrast, England (1987) believes that considerable regional uplift, accompanied by crustal extension, has occurred during the Late Tertiary/Quaternary. In a major departure from previous discussions, England (1987) questions the significance of glaciation in the geomorphic evolution of the high arctic. He introduces several concepts which provide a new perspective in evaluating the older glacial record in the region. These include; i) that the major fiords and inter-island channels are a product of tectonics and not glacial erosion, and that many have never been occupied by glaciers; ii) that tectonics and not climate has been the major control on glacial style through time; and iii) that it is incorrect to use present topography when interpreting the maximum distribution of erratics which, in turn, could be of Tertiary fluvial origin.

One important component of England's (1987) thesis is the distribution of shoreline features which pre-date the last glaciation. He notes that shorelines associated with two older glaciations are well preserved along Nares Strait but are not recognized within the adjacent fiords, despite extensive field traverses. It is also clear that many of the fiords remained free of glaciers throughout the last glaciation, and thus the absence of the "old" shorelines cannot be explained through glacial erosion. Nor is it logical to suggest that they would be preferentially removed by weathering. England suggests that the old marine shorelines were never formed in these areas and that the associated glacial events actually pre-date the formation of the fiords, which are a product of tectonism. This study serves to support England's observations. Pre-Holocene marine shoreline are observed along the northern coast of Marvin Peninsula, but similar features are not found along the coastlines of Disraeli Fiord and M'Clintock Inlet. Like Nares Strait, the north coast of Marvin Peninsula falls along the perimeter of Ellesmere Island and represents a coastal margin which dates back to well into the Tertiary. In contrast, the age of fiords is completely unknown.

Evaluation of the relative significance of tectonism versus glacial/glacioisostatic processes in the field area is difficult. The most striking physiographic features, the fiords, have long been assumed to be the product of glacial erosion (cf. Hattersley-Smith 1961). However, the limited evidence found for a pervasive ice cover in the area does not include either large-scale erosional or depositional features which one would expect to be associated with a highly erosive ice sheet. Schei (1904) and Gregory (1913) first noted this paradox of fiords in a landscape with little evidence of glaciation, and concluded that the fiords must be of tectonic origin. Christie's (1967) observation that valleys along the north coast show only minor glacial modification is particularly relevant. If Disraeli Fiord is a product of selective linear erosion (cf. Sugden 1978), the presence of bedrock islands in the central and outer fiord with adjacent water depths of >400 m indicates that erosion was selective indeed!

Several observations from this study are relevant to the ideas presented by England (1987).

A - If the shell fragments found at high elevations on eastern Ward Hunt Island are correctly interpreted as being erratics, they indicate that the glacier which reached the island had passed through a marine channel. While this does not indicate that Disraeli Fiord was either as deep or extensive as it is today, neither does it necessitate that extensive glaciation occurred on a landscape radically different from that of today.

B - Fluvial gravels and till which predate the last glaciation are exposed along a river cut section at 65 m asl in central Marvin Peninsula. This indicates that base level at the time these sediments were deposited was below the elevation of this site.

C - The difference between the highest sea level and the Holocene marine maximum recorded in Sentinel Valley is roughly proportional to the difference in ice load between the last glaciation and that defined by older melt water channels. This suggests that tectonic uplift between these two glacial events was minimal.

These observations neither unequivocally support or refute England's (1987) hypothesis. Point B suggests that if recent regional uplift has occurred, it has not been spatially uniform. Both A and C suggest that tectonic factors do not need to be invoked in order to explain the old glacial and sea level features observed in the field area. However, it should be noted that the pattern of postglacial emergence observed on Marvin Peninsula cannot be accounted for by glacioisostatic factors alone, and suggests that neotectonism has been active during the

Holocene influencing the elevations of the entire sea level record (Chapter II).

The Arctic Ocean

From the time of Ewing and Donn (1956; also see Donn and Ewing 1966), the role of the Arctic Ocean as a control of glaciation has been the subject of considerable controversy. Among the most imaginative ideas are those which propose a glacial ice cap on the Arctic Ocean during full glacial time (Mercer 1970; Broecker 1975; Denton and Hughes 1981). Although this is an interesting concept from a theoretical perspective (see Hughes 1986) there is simply no field evidence to support that model.

During the last glaciation the climate of northern Ellesmere Island was characterized by increased cold and extreme aridity (England 1976, 1983, 1986; England and Bradley 1978; Chapter II). The limited extent of glaciers at this time was largely a result of limited precipitation, as it is today (Bradley and England 1978). Calving in the sea was another important factor limiting ice extent during the last glaciation (England 1987). If glaciers were able to advance through major marine channels in the past, as evidenced by the advance of Greenland Ice onto northeastern Ellesmere Island and the proposed advance of Ellesmere Ice onto Ward Hunt Island, then: i) the activity index of those glaciers must have been significantly higher than at present in order to overcome the efficiency of calving as an ablation process; and/or ii) the topography at that time was less accentuated than it is at present. Regardless, extensive regional glaciation at some time in the past necessitated increased moisture availability (Bradley and England 1978).

The stability of the sea-ice cover on the Arctic Ocean, which would greatly influence its effectiveness as a moisture source, has been the subject of considerable controversy. Clarke (1982) argued for a stable sea-ice cover since the middle Cenozoic, whereas Herman and Hopkins (1980) believed a permanent ice-cover was not established until after 1 Ma. Funder et al. (1985) demonstrated the existence of a forest tundra environment on northern Greenland ca. 2 Ma, and supported the idea of an ice-free Arctic Ocean at that time. Carter et al. (1986), based upon effective diagenetic temperatures calculated from amino acid analysis, considered the perennial sea-ice cover of the Arctic Ocean to have been established in association with the first major glaciation of the northern hemisphere, about 2.5 Ma. Since that time they consider the ice cover to have been continuous except during the warmest interglacial episodes. Given the unknown age of the most extensive glaciation on northern Ellesmere Island it remains possible that it was associated with an ice-free Arctic Ocean. The lesser extent of subsequent glaciations may reflect the establishment of a perennial ice-cover on the Arctic Ocean Basin and the resulting increased aridity of the region. Whether or not the oldest glaciation(s) occurred on a landscape significantly different from that of the present (cf. England 1987) remains a controversy.

BIBLIOGRAPHY

Bednarski, J. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 23, 1343-1355.

_________. 1987. Glacial advances in Otto Fiord, Ellesmere Island, N.W.T. 16th Arctic Workshop, Boreal Institute for Northern Studies, University of Alberta, Abstracts, 4-5.

_________. 1988. Quaternary history of Nansen Sound; preliminary observations. Canadian Association of Geographers Annual Meeting, Saint Mary's University, Halifax, Nova Scotia. Programme and abstracts, p. 48.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, 7, 634-664.

_________. 1987. Geological Survey of Canada radiocarbon dates XXVI. Geological Survey of Canada Paper 86-7, 60 p.

Bovis, M.J., and Barry, R.G. 1974. A climatological analysis of north polar desert areas. In, Polar Deserts and Modern Man. Edited by, Smiley, T.L., and Zumberge, J.H. The University of Arizona Press, Tucson, Arizona, 23-31.

Bradley, R.S. 1985. Quaternary paleoclimatology: methods of paleoclimatic reconstruction. Allen and Unwin, Boston, Mass., 472 p.

Bradley, R.S., and England, J. 1978. Recent climatic fluctuations of the Canadian high arctic and their significance for glaciology. Arctic and Alpine Research, 4, 715-731.

Carter, L.D., Brigham-Grette, J., Marincovich, L., Jr., Pease, V.L., and Hillhouse, J.W. 1986. Late Cenozoic Arctic Ocean sea ice and terrestrial paleoclimate. Geology, 14, 675-678.

Christie, R.L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. Geological Survey of Canada Memoir 331, 79 p.

_________. 1967. Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic archipelago. Geological Survey of Canada Bulletin 138, 50 p.

Clark, D.L. 1982. The Arctic Ocean and post-Jurassic paleoclimatology. In, Climate in Earth History. Edited by, Berger W.H., and Crowell, J.C. National Academy Press, Washington, D.C., 133-139.

Clark, D.L., Vincent, J.-S., Jones, G.A., and Morris, W.A. 1984. Correlation of marine and continental glacial and interglacial events, Arctic Ocean and Banks Island. Nature, 311, 147-149.

Crary, A.P. 1956. Geophysical studies along northern Ellesmere Island. Arctic, 9, 155-165.

Dalrymple, R.W., and Maass, O.C. 1987. Clay mineralogy of late Cenozoic sediments in the CESAR cores, Alpha Ridge, central Arctic Ocean. Canadian Journal of Earth Sciences, 24, 1562-1569.

Dawes, P.R. 1986. Glacial erratics on the Arctic Ocean margin of north Greenland: implications for an extensive ice shelf. Bulletin of the Geological Society of Denmark, 35, 59-64.

Denton, G.H., and Hughes, T.J. 1981. The Arctic Ice Sheet: An outrageous hypothesis. In The Last Great Ice Sheets. Edited by Denton, G.H., and Hughes, T.J. Wiley-Interscience, New York, 437-467.

Donn, W.L., and Ewing, M. 1966. A theory of ice ages III. Science, 152, 1706-1712.

England, J. 1974. The glacial geology of Archer Fiord/Lady Franklin Bay, northeastern Ellesmere Island, N.W.T. Ph.D. thesis, University of Colorado, Boulder, Colorado, 234 p.

_________. 1976. Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: alternative models. Quaternary Research, 6, 185-202.

_________. 1978. The glacial geology of northeastern Ellesmere Island, N.W.T., Canada. Canadian Journal of Earth Sciences, 15, 603-617.

_________. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, 20, 895-917.

_________. 1985. The late Quaternary history of Hall Land, northwest Greenland. Canadian Journal of Earth Sciences, 22, 1394-1408.

_________. 1986. A paleoglaciation level for north-central Ellesmere Island, N.W.T. Arctic and Alpine Research, 18, 217-222.

_________. 1987. Glaciation and evolution of the Canadian high arctic landscape. Geology, 15, 419-424.

England, J., and Bradley, R.S. 1978. Past glacial activity in the Canadian high arctic. Science, 200, 265-270.

England, J., Bradley, R.S., and Miller, G.H. 1978. Former ice shelves in the Canadian high Arctic. Journal of Glaciology, 20, 393-404.

England, J., Bradley, R.S., and Stuckenrath, R. 1981. Multiple glaciations and marine transgressions, western Kennedy Channel, Northwest Territories, Canada. Boreas, 10, 71-89.

England, J., Lemmen, D.S., Evans, D.J.A., and Bednarski, J. 1987. The Quaternary of northern Ellesmere Island. International Union for Quaternary Research, XII[th] International Congress, Ottawa, Canada. Programme with Abstracts, p. 162.

Evans, D.J.A. 1988. Glacial geomorphology and late Quaternary of Phillips Inlet and the Wootton Peninsula, northwest Ellesmere Island, Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

Ewing, M., and Donn, W.L. 1956. Theory of ice ages. Science, 129, 463-465.

Fairbridge, R.W. 1968. The Encyclopedia of Geomorphology. Reinhold Book Corporation, New York, 1295 p.

Frisch, T. 1974. Metamorphic and plutonic rocks of northernmost Ellesmere Island, District of Franklin. Geological Survey of Canada Bulletin 229, 87 p.

Funder, S., Abrahamsen, N., Bennike, O., and Feyling-Hanssen, R.W. 1985. Forested Arctic: evidence from north Greenland. Geology, 13, 542-546.

Gregory, J.W. 1913. The nature and origin of fiords. J. Murray, London, 542 p.

Hattersley-Smith, G. 1961. Geomorphological studies in northwestern Ellesmere Island. Defence Research Board of Canada, Ottawa, Miscellaneous Report No. G-5.

__________. 1969. Glacial features of Tanquary Fiord and adjoining areas of northern Ellesmere Island, N.W.T. Journal of Glaciology, 8, 23-50.

Hattersley-Smith, G., and other members of the expedition (A.P. Crary and R.L. Christie). 1955. Northern Ellesmere Island, 1953 and 1954. Arctic, 8, 3-36.

Hattersley-Smith, G., Fuzesy, A., and Evans, S. 1969. Glacier depths in northern Ellesmere Island: airborne radio echo sounding in 1966. Defence Research Board, Canada, Defence Research Establishment Ottawa Technical Note No. 69-6, 23 p.

Herman, Y., and Hopkins, D.M. 1980. Arctic Ocean climate in late Cenozoic time. Science, 209, 557-562.

Hodgson, D.A. 1985. The last glaciation of west-central Ellesmere Island, Arctic Archipelago, Canada. Canadian Journal of Earth Sciences, 22, 347-368.

Hughes, T.J. 1986. The marine ice transgression hypothesis. Geografiska Annaler, 69A, 237-250.

Lowden, J.A. 1985. The Geological Survey of Canada radiocarbon dating laboratory. Geological Survey of Canada Paper 84-24, 19 p.

Lyons, J.B. and Leavitt, F.G. 1961. Structural-stratigraphic studies on the Ward Hunt Ice Shelf. Air Force Cambridge Research Laboratories, Final Report Contract AF19(604)-6188, Bedford Massachusetts, 37 p.

Lyons, J.B., and Mielke, J.E. 1973. Holocene history of a portion of northernmost Ellesmere Island, Arctic, 26, 314-323.

Maxwell, J.B. 1982. The climate of the Canadian arctic islands and adjacent waters. Climatological Studies, No. 30, volume 2, 589 p.

Mercer, J.H. 1970. A former ice sheet in the Arctic Ocean? Palaeogeography, Palaeoclimatology, Palaeoecology, 8, 19-27.

Miller, G.H., Bradley, R.S. and Andrews, J.T. 1975. The glaciation level and lowest equilibrium line altitude in the high Canadian arctic: Maps and climatic interpretation. Arctic and Alpine Research, 7, 155-168.

Narod, B.B., Clarke, G.K.C., and Prager, B.T. 1988. Airborne UHF radar sounding of glaciers and ice shelves, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 25, 95-105.

Pheasant, D.R., and Andrews, J.T. 1973. Wisconsin glacial chronology and relative sea level movements, Narpaing Fiord, Broughton Island area, eastern Baffin Island, N.W.T. Canadian Journal of Earth Sciences, 10, 1621-1641.

Reeh, N. 1982. A plasticity theory approach to the steady-state shape of a three-dimensional ice sheet. Journal of Glaciology, 28, 431-455.

________. 1984. Reconstruction of the glacial ice covers of Greenland and the Canadian arctic islands by three-dimensional, perfectly plastic ice-sheet modelling. Annals of Glaciology, 5, 115-121.

Retelle, M.J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. Canadian Journal of Earth Sciences, 23, 1001-1012.

Rutter, N.W., Crawford, R.J., and Hamilton, R.D. 1985. Amino acid racemization dating. In Dating Methods of Pleistocene Deposits and their Problems. Edited by Rutter, N.W. Geoscience Canada, Reprint Series 2, 23-30.

Schei, P. 1904. In, Sverdrup, O. New land, four years in the Arctic regions. Volume II, appendix I. Longmans, Green and Company, London, 455-465.

Sugden, D.E. 1978. Reconstruction of the morphology, dynamics and thermal characteristics of the Laurentide Ice Sheet at its maximum. Journal of Glaciology, 20, 367-391.

Thorsteinsson, R., and Tozer, E.T. 1970. Geology of the arctic archipelago. In, Geology and Economic Minerals of Canada. Edited by Douglas, R.J.W. Geological Survey of Canada, Economic Geology Report 1, 548-590.

Tozer, E.T. 1955. Geological reconnaissance, Prince Patrick, Eglington and western Melville Islands, arctic archipelago, N.W.T. Geological survey of Canada Paper 55-5, 32 p.

Trettin, H.P. 1981. Geology of Precambrian to Devonian rocks, M'Clintock Inlet area, District of Franklin (NTS 340 E, H)-preliminary geological map and notes. Geological Survey of Canada, Open File Report 759.

________. 1987a. Pearya: a composite terrane with Caledonian affinities in northern Ellesmere Island. Canadian Journal of Earth Sciences, 24, 224-245.

________. 1987b. Pre-Carboniferous geology, M'Clintock Inlet map-area, northern Ellesmere Island, interim report and map. Geological Survey of Canada, Open File Report 1652.

IV. A PRELIMINARY INVESTIGATION OF DEEP WATER SEDIMENTS FROM DISRAELI FIORD, HIGH ARCTIC CANADA

Introduction

The fiords and channels of the high arctic contain a valuable record of sedimentation which remains uninvestigated (Syvitski 1986). These sediments record depositional processes within a unique environment which have only been speculated upon through inductive modelling (cf. Powell 1984). Furthermore, most fiords in the Canadian high arctic should contain a long depositional record because they remained free of glaciers throughout the last glaciation (England 1983; England et al. 1987). Indeed, some fiords may contain sediments as old as their initial formation, because they do not appear to have ever been occupied by trunk glaciers (England 1987a). Integration of the marine record with the more accessible, but incomplete, terrestrial record provides the opportunity to increase our understanding of former ice extent and the nature and timing of climatic change during the Quaternary (cf. Blake 1970; England 1976, 1987a; Evans and Lemmen 1987). Most previous investigations of fiords and channels in the high arctic have concerned oceanography (Ford and Hattersley-Smith 1965; Hattersley-Smith and Serson 1966; Keys et al. 1969; Lake and Walker 1973, 1976; Keys 1978; Jeffries and Krouse 1984; Jeffries 1985; Horne 1985; Jeffries et al. 1988). Sedimentological investigations of glaciomarine deposits in this region have concerned either; i) sediments that have been isostatically raised above present sea level (Retelle 1986; Bednarski 1988; Stewart 1988); or ii) cores from the Arctic Ocean Basin and shelf areas, (e.g. Herman and Hopkins 1980; Clark et al. 1980; Clark and Hansen 1983; Morris et al. 1985; Mudie and Blasco 1985; Dalrymple and Maass 1987; Hein and Mudie 1987; Mudie et al. 1988). Raised marine deposits are commonly dominated by large volumes of deglacial sediments, frequently ice-proximal facies, which record a complex depositional environment (Stewart 1988). The duration of this record is generally restricted to several hundred to a few thousand years. In contrast, cores from the Arctic Ocean Basin and shelf provide an extensive record dating back to > 5 Ma (Clark et al. 1980). However, owing to their low sedimentation rates they provide limited resolution when examining late Quaternary events. Consequently, fiords represent an important intermediate environment recording fluctuations in sediment input from the adjacent terrestrial environment.

In June, 1987, coring of Disraeli Fiord, northernmost Ellesmere Island (Fig. IV.1), was conducted in conjunction with a detailed investigation of the terrestrial Quaternary record of Marvin Peninsula (Chapters II and III). Previous investigations have been made concerning the oceanography of the fiord (Keys et al. 1969; Keys 1978; Jeffries and Krouse 1984; Jeffries 1985), but this is the first report on the deep water sediments. This paper discusses the processes of physical sedimentation in the fiord and how the observed sediments may relate to these processes, as well as to climatic change. Future analysis of the biological component of the sediment record will likely provide additional insights regarding paleoenvironmental change.

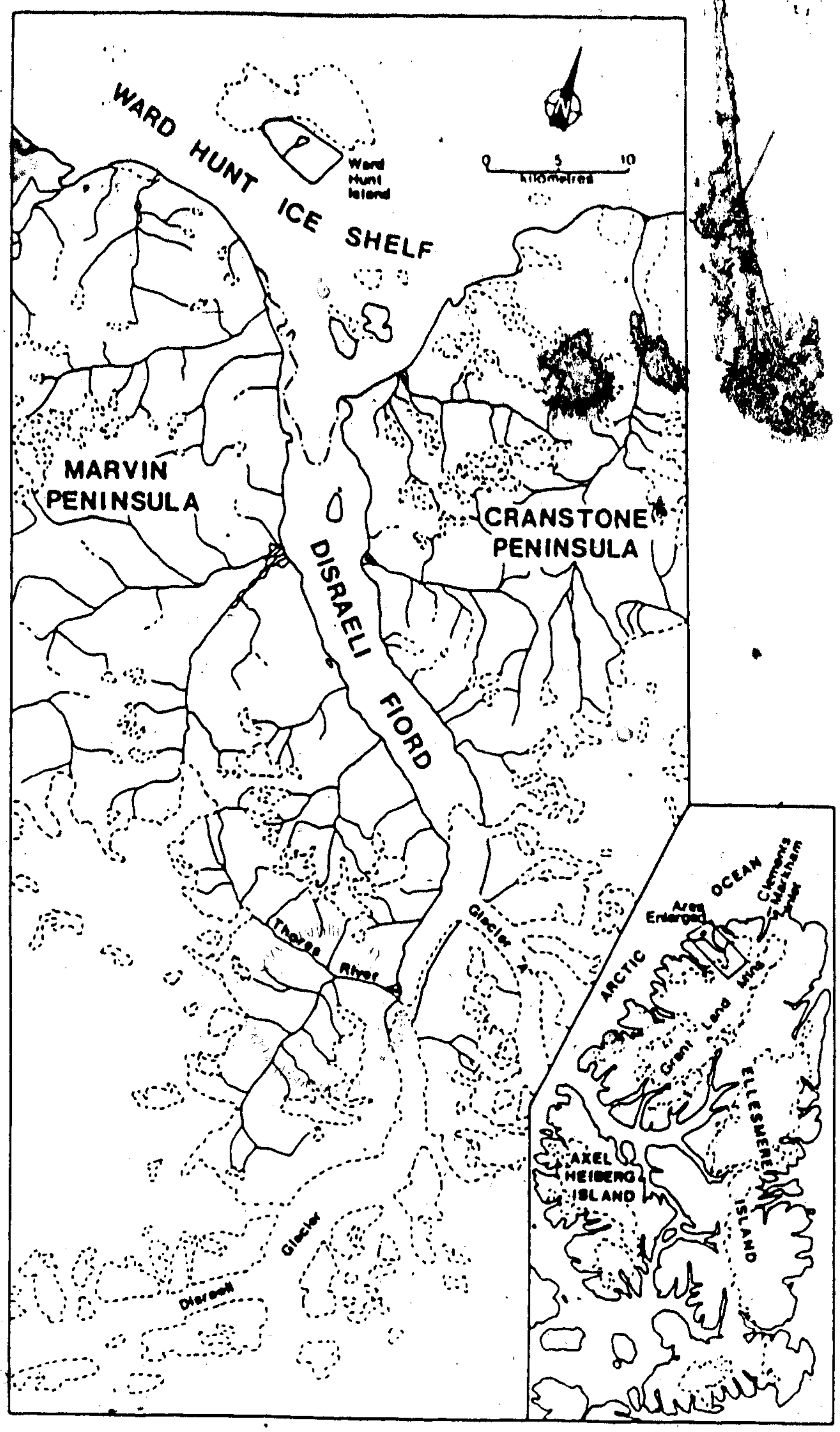


Figure IV.1. Disraeli Fiord and its tributary basins. Glaciers are dashed.

Field area

The north coast of Ellesmere Island features numerous fiords which incise a mountainous landscape. To the south the Grant Land Mountains rise to 2500 m asl and are occupied by extensive ice fields above 1000 m asl (Fig. IV.1). Disraeli Fiord extends about 60 km from these mountains to the Arctic Ocean and is bounded by Marvin and Cranstone Peninsulas to the west and east, respectively (Fig. IV.1). The average width of the fiord is about 5 km. Geologically the area is one of the most complex in the arctic islands, and it encompasses a wide suite of volcanic, metamorphic and sedimentary lithologies (Christie 1964; Trettin 1981, 1987).

Three glaciers contact the sea near the head of Disraeli Fiord. Two of these, Disraeli Glacier and Glacier "A", have filled the inner 16 km of the fiord with floating ice (Figs. IV.1 and IV.5). The Disraeli Glacier is in excess of 600 m thick within 15 km of its terminus (Hattersley-Smith et al. 1969; Narod et al. 1988). The mouth of Disraeli Fiord is occupied by Ward Hunt Ice Shelf, a sea-ice ice shelf (Lemmen et al. in press) which formed ca. 3000 - 4500 BP (Crary 1960, Lyons and Mielke 1973). The ice shelf obstructs the flow of surface water exiting the fiord, creating a 44 m cap of freshwater (Fig. IV.2) corresponding to the average thickness of the ice shelf (Keys et al. 1969; Keys 1978; Jeffries and Krouse 1984; Jeffries 1985). Between the Ward Hunt Ice Shelf and the floating glaciers Disraeli Fiord is covered by multi-year, landfast sea-ice that has attained an equilibrium thickness of ca. 3 m (Keys et al. 1969; Hattersley-Smith 1973; Jeffries 1985). These three features; floating glaciers, multi-year, landfast sea-ice and the enclosure by the Ward Hunt Ice Shelf; combine to

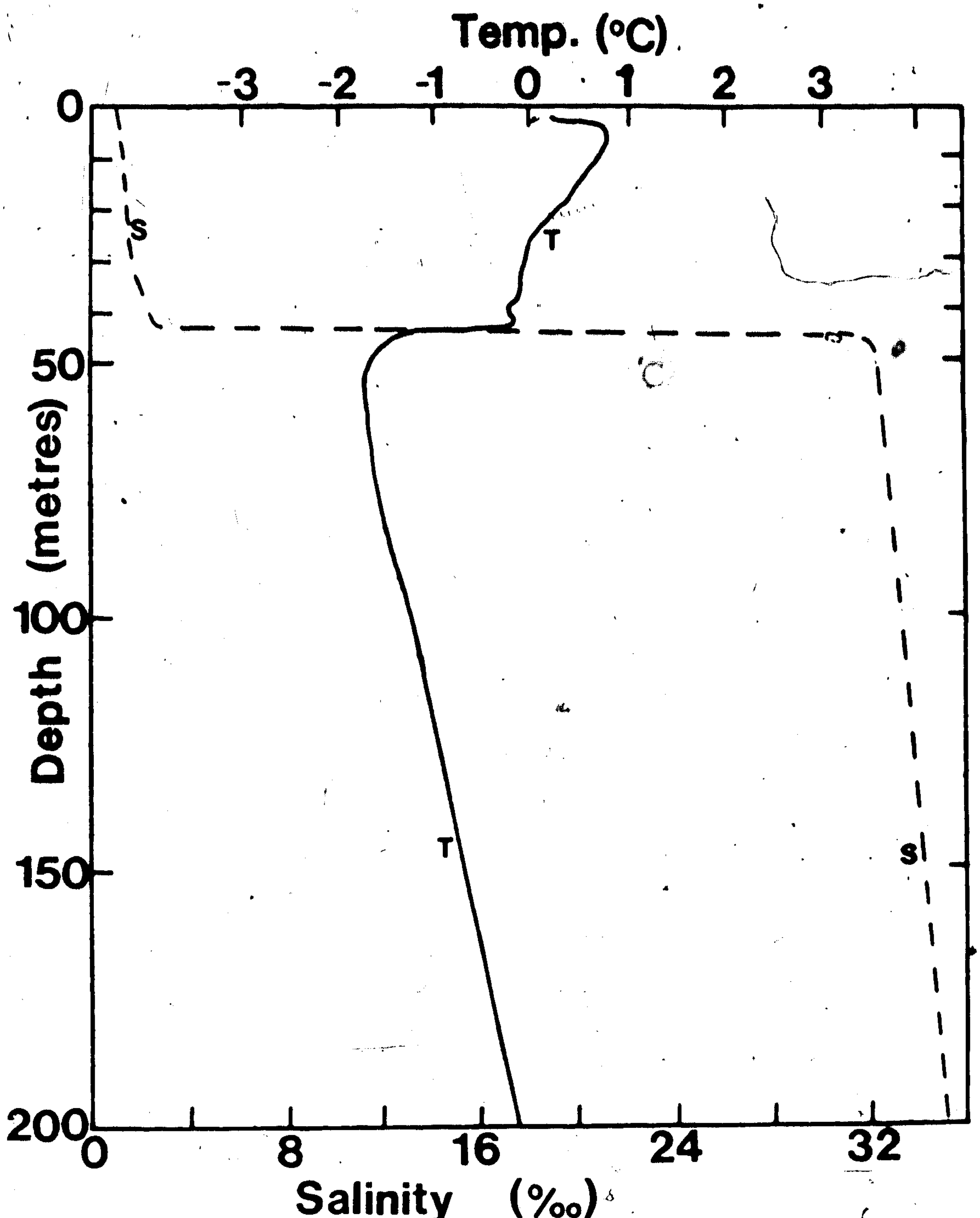


Figure IV.2. Temperature (T) and salinity (S) profiles from Disraeli Fiord (from Keys 1978). Note sharp pycnocline at 44 m depth marking the base of the freshwater cap impounded by the Ward Hunt Ice Shelf.

create a glaciomarine environment unique to the north coast of Ellesmere Island.

Little is known about the bathymetry of the fiord. Conventional echo-sounding of bathymetry is not possible owing to the permanent sea-ice cover. Soundings made by various workers indicate maximum depths of >400 m near the fiord mouth (Crary 1956; Keys et al. 1969; Keys 1978; Jeffries 1985). There is no evidence of a sill, although there are little data available regarding the bathymetry beneath the Ward Hunt Ice Shelf. Speculation regarding the presence of large moraines near the mouth of the fiord (Crary 1956) is based upon a few point soundings, and is not supported by geomorphology on the adjacent land. The tidal range in the fiord is less than 1 m.

No data are available regarding the input of sediment to any of the fiords of the high arctic. Numerous tributary valleys contribute sediment into Disraeli Fiord, but the two largest valleys near its centre are likely the major points of sediment influx (Fig. IV.1). The reworking of glacial, deltaic and marine deposits within these valleys provides most of the sediments. Although sediment input from the glaciers occupying the fiord head is unknown, it is important to note the differences between a tidewater glacier regime (incorrectly reported for Disraeli Fiord by Syvitski 1976) and the floating glacier tongues which presently occupy the fiord (cf. Powell 1984). For example, englacial and basal meltwater, commonly associated with tidewater glaciers, is presumably absent in the cold-based (subpolar) glaciers which feed into the floating tongues. Rather, their meltwater is confined to meandering supraglacial streams that are incapable of transporting large volumes of sediment owing to their gentle gradients and the presence of numerous small ponds, which act as sediment traps.

Quaternary history

Recent studies on the Quaternary history of Marvin Peninsula (Fig. IV.1) have demonstrated that glaciers underwent minor advances during the last glaciation (Chapter II). Ice occupying major valleys terminated rapidly upon contact with the sea, which reached 70 to 120 m asl, often penetrating well inland of the present coastline. The former ice limit at the fiord head is not well recorded, although the Disraeli Glacier did advance beyond its present grounding line. Nonetheless, most of Disraeli Fiord remained ice-free throughout the last glaciation (Fig. IV.3). This reconstruction contradicts a previous report of a pervasive regional ice cover during the last glaciation for this area (Lyons and Mielke 1973). Ice retreat from the last glacial maximum had begun by 9560 BP (TO-488, Chapter II). Sites near the fiord head became ice-free ca. 8000 BP, as evidenced by numerous radiocarbon dates from Thores River (Fig. IV.1). Driftwood collected at Thores River dated 6030±70 (TO-495), indicating that Glacier "A" was less extensive at that time and that it has recently readvanced into the fiord head (Chapter II). Raised marine deltas and fine-grained sediments record the influx of large volumes of deglacial sediment into the fiord. There is no evidence for subsequent impulses of sediment of similar magnitude (Chapter II).

Well developed beaches and abundant sea-ice push ridges at, or just below, marine limit demonstrate the occurrence of significant open water during the early Holocene. Driftwood dating 8850±80 BP (GSC-4559) was collected along the north coast of Marvin Peninsula and indicates that an ice shelf was not present at this time (Chapter II). Evans and Lemmen (1987) use this evidence to suggest an early Holocene climatic optimum in this area, whereas other studies support a mid-Holocene climatic optimum

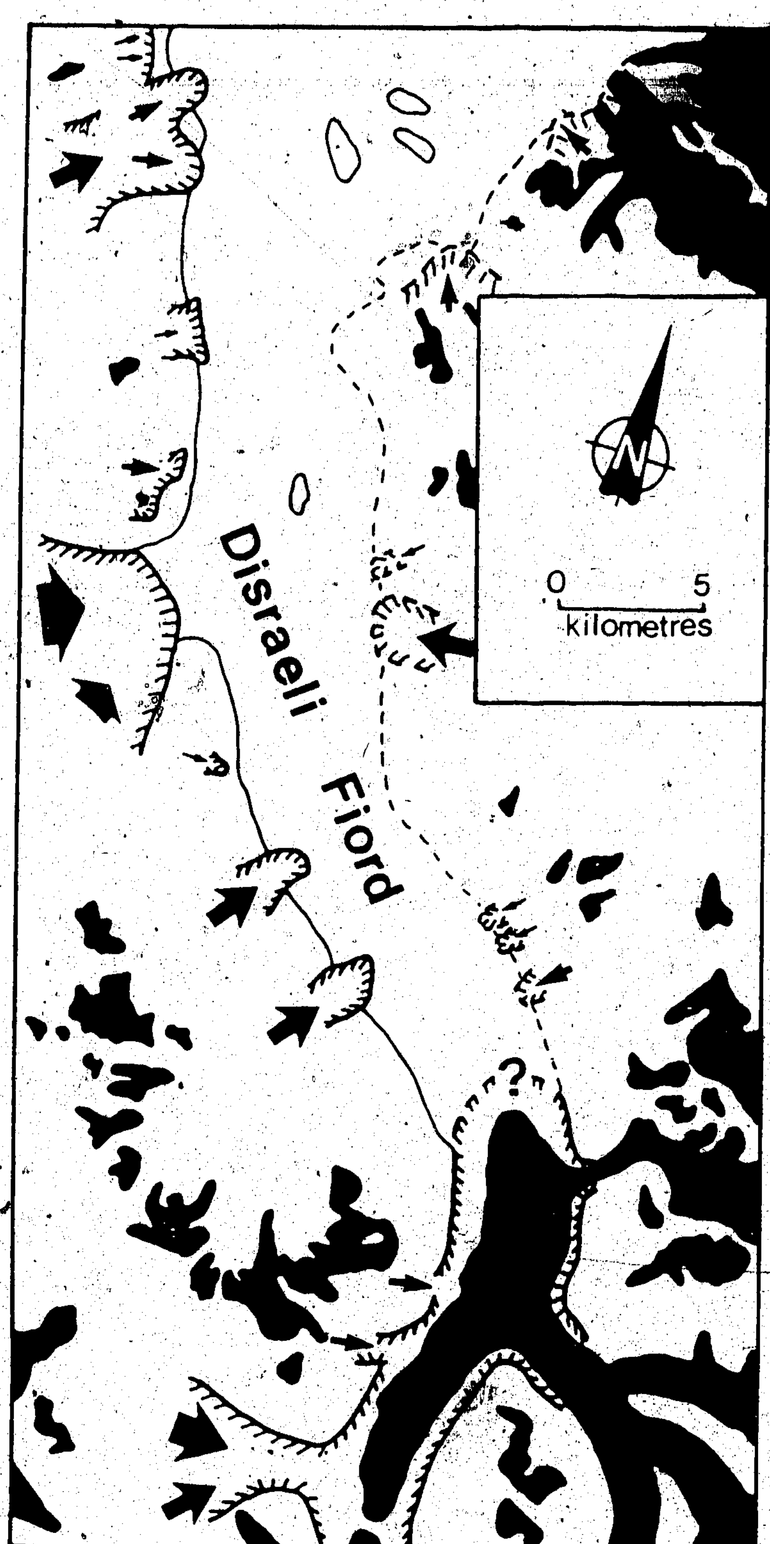


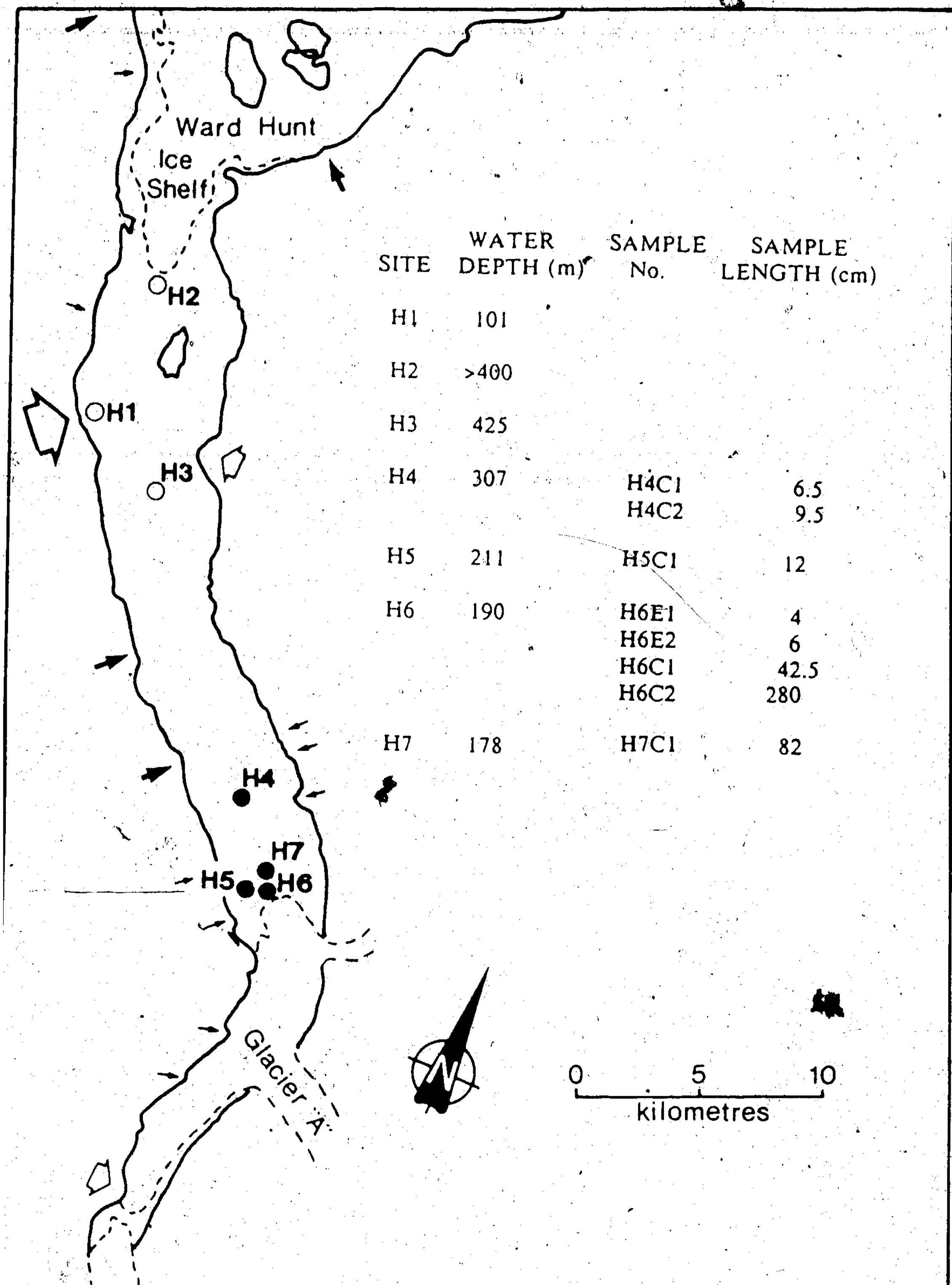
Figure IV.3. Paleogeography of Disraeli Fiord ca. 10,000 BP showing the ice limit of the last glaciation. Arrows indicate the direction of former ice flow. No field work has been conducted along the east coast of the fiord. Present glaciers are indicated by black tone.

(Lyons and Mielke 1973; Stewart and England 1983). Formation of the Ward Hunt Ice Shelf is attributed to a climatic deterioration that began ca. 4500 BP (Lyons and Mielke 1973; Stewart and England 1983; Jeffries 1985), and driftwood younger than 3000±200 BP (L254D, Crary 1960) has not been collected within Disraeli Fiord. Therefore, this date of 3000 BP provides a maximum age for the blockage of Disraeli Fiord by the ice shelf. Isotopic analysis of the Ward Hunt Ice Shelf provides evidence of a subsequent climatic amelioration that lasted about 400 to 900 years, terminating ca. 1600 BP (Jeffries and Krouse 1984; Jeffries 1985).

This record of late Quaternary climatic change provides the basic framework for interpretation of sediment cores from Disraeli Fiord. As most of the fiord remained free of glaciers during the last glaciation, the possibility exists for the recovery of sediments of full glacial age.

Methods

Field work was conducted from June 10 to 23, 1987. At eight sites (Fig. IV.4) a 20 cm diameter hole was drilled through the sea-ice with a power auger. Ice thicknesses ranged from 2.3 to 3.2 m. An Ekman grab sampler was used to obtain an undisturbed record of the uppermost sediments. Coring was conducted with a hand-operated percussion corer (Reasoner 1986), which allowed for a maximum core length of 3 m. A total of eight samples (six cores and two grab samples) were recovered from four sites in the inner fiord, where water depths ranged from 178 to 307 m (Fig. IV.4). Three of these sites were within 1 km of the floating tongue of Glacier "A" (Figs. IV.4 and IV.5). Only three of the samples were >15 cm in length, but the longest core, from site H6, was 280 cm.



SITE	WATER DEPTH (m)	SAMPLE No.	SAMPLE LENGTH (cm)
H1	101		
H2	>400		
H3	425		
H4	307	H4C1	6.5
		H4C2	9.5
H5	211	H5C1	12
H6	190	H6E1	4
		H6E2	6
		H6C1	42.5
		H6C2	280
H7	178	H7C1	82

Figure IV.4. Coring sites on Disraeli Fiord. Black circles note sites of successful coring. No sediments were recovered from the other sites. Arrows denote points of inflow. Dashed lines show the margins of ice shelves; glacial ice shelves at the fiord head and sea-ice ice shelf at the fiord mouth.

Figure IV.5. Oblique air photo looking east across the head of Disraeli Fiord, showing sites where cores were obtained.

Ekman grab samples were dried, split and photographed in the field. Cores were returned to the laboratory in the core tube and stored at 2°C. The cores were x-rayed, then split and photographed. Half of each core was left to dry and then was rephotographed when maximum contrast had been obtained. Samples for particle-size analysis were taken at 10 cm intervals on the longest core (H6C2), and at 5 cm intervals on the shorter samples. Unfortunately, deformation of the sediment during coring precluded the sampling of individual laminae, which in some cases were attenuated more than 1 m down the side of the core. Samples containing a significant sand component were wet-sieved at 62 um. The coarse fraction was then dried and sieved at half Phi intervals. The low percentage of coarse silt found in the sieved samples suggests that some of this fraction was removed along with the sand, introducing additional error to the statistical analysis. The fine-grained fraction was analyzed with a Sedigraph 5000D Particle Size Analyzer (cf. Stein 1985; Andrews 1985), with the results considered to be accurate to within ±1.5%. Statistical parameters were calculated using the inclusive graphic method of Folk (1980). Organic carbon content was calculated by subtracting the percentage of inorganic carbon, determined using the titrimetric method of Bundy and Bremner (1972), from the percentage of total carbon, determined using a Leco Induction Furnace.

Core descriptions

Prior to splitting the cores, x-radiography showed the sediments to be apparently structureless. It was also evident that the sediments within core H6C2 had been stretched (<10 cm?) when the core tube was removed from the fiord floor.

Splitting of the cores revealed marked changes in sediment colour and structure that were correlative between samples. On the basis of these criteria, three units are recognized (Fig. IV.6). The uppermost sediments, unit A, are reddish gray in colour (5YR 5/2) and massive. Five of the samples contain only this unit. In the remaining three cores there is a gradational boundary between 15 and 17 cm depth to reddish brown (5YR 5/4), massive to diffusely-laminated sediments which define unit B (Fig. IV.7). Both units A and B contain abundant evidence of bioturbation. The two longest cores, H6C2 and H7C1, show a sharp change at 42 and 31 cm depth, respectively, to gray-coloured (7.5 YR 5/0) interlaminated sediments (rhythmites) which characterize unit C (Fig. IV.8). This third unit continues to the base of both cores. Most individual rhythmites within zone C are normally graded, although a few are massive. Very fine to fine sand, commonly only a few grains in thickness, is present at the base of some laminae. Rhythmites range in thickness from < 0.1 to 2 cm. However, the significance of these measurements is questionable given the severe attenuation of individual laminae. Nevertheless, the rhythmites do appear to thicken and coarsen with depth. Unequivocal evidence of current flow (i.e. cross-lamination) was not present in any of the samples.

Grain size analysis allows the recognition of two classes of sediment (Fig. IV.9, Table IV.1; complete grain size data are presented in Appendix IV.). The majority of the sediments from all three units (91% of analyzed samples) are considered to be of type 1: fine-grained (mean 8.7ø, clay content 18 - 82%), and moderately to very poorly sorted. Comparison of

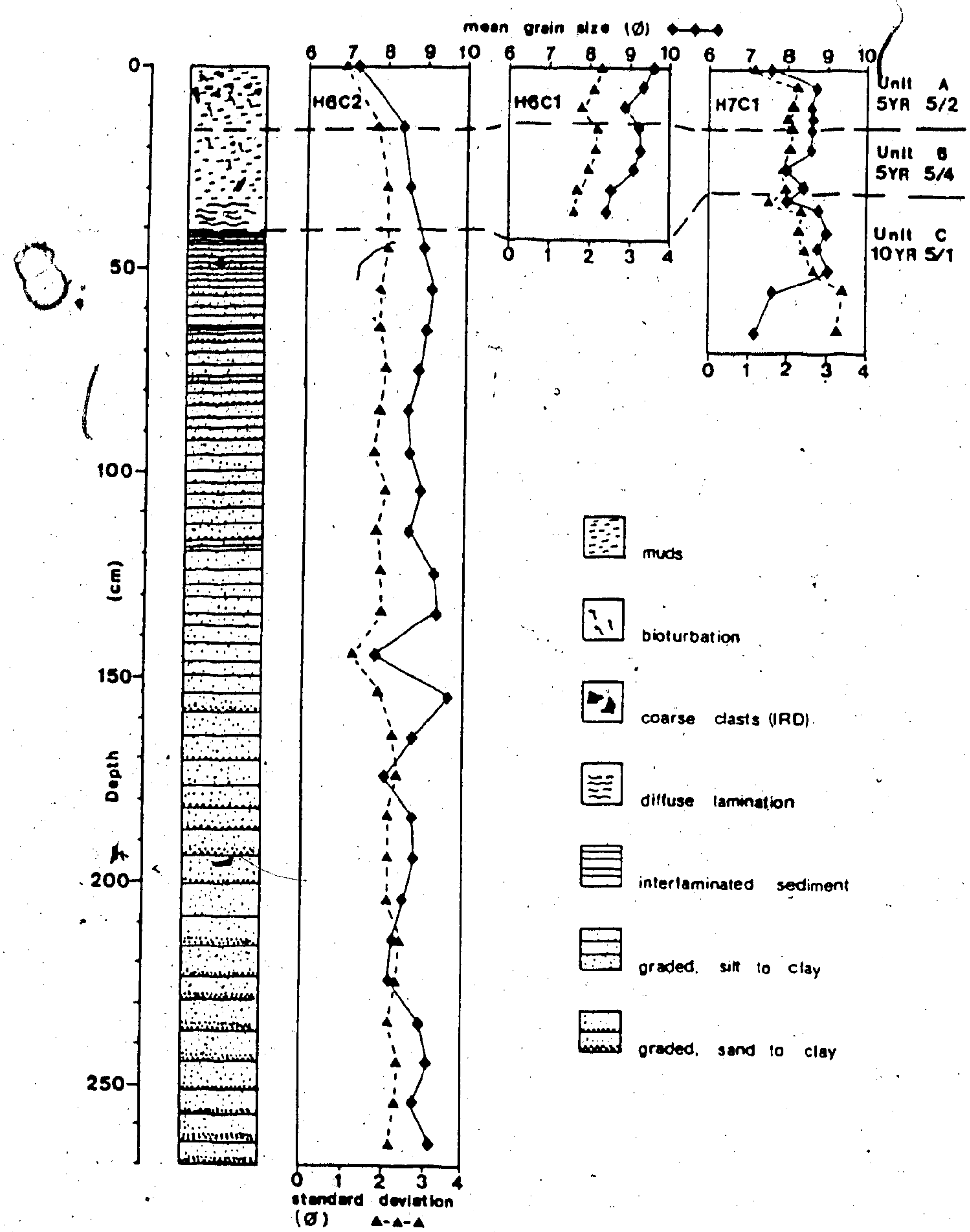


Figure IV.6. Sediment characteristics for the cores longer than 15 cm. Log is schematic and laminations are not shown to scale.

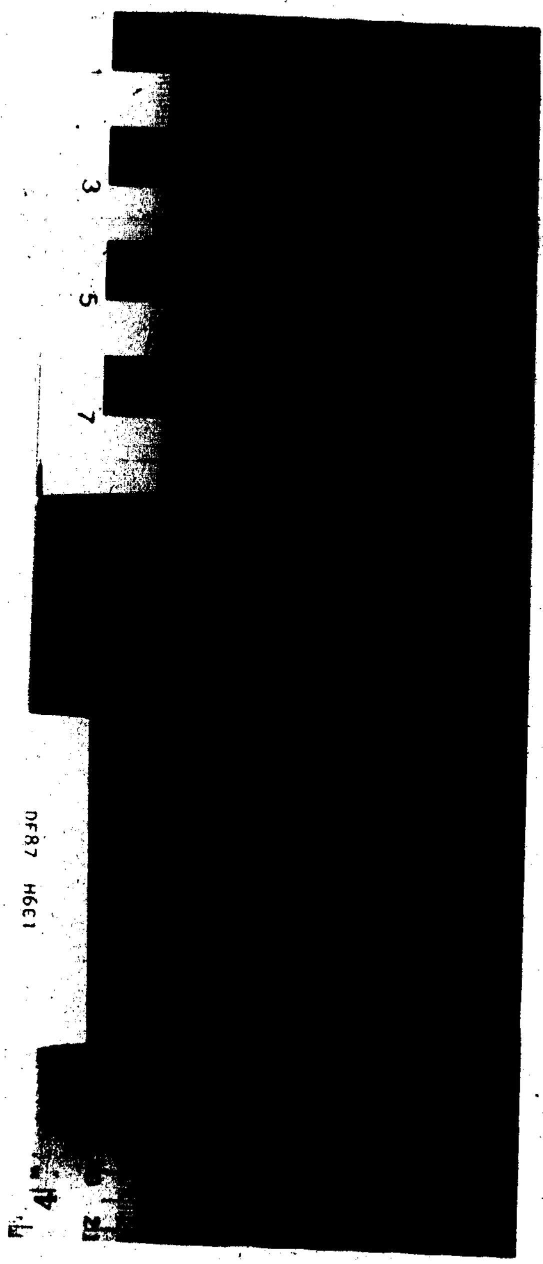


Figure IV.7. Massive to diffusely-laminated sediments of units A and B, core H6C1. Scale is in centimetres. Arrow points to concentration ice-rafted debris between 6 and 9 cm depth.

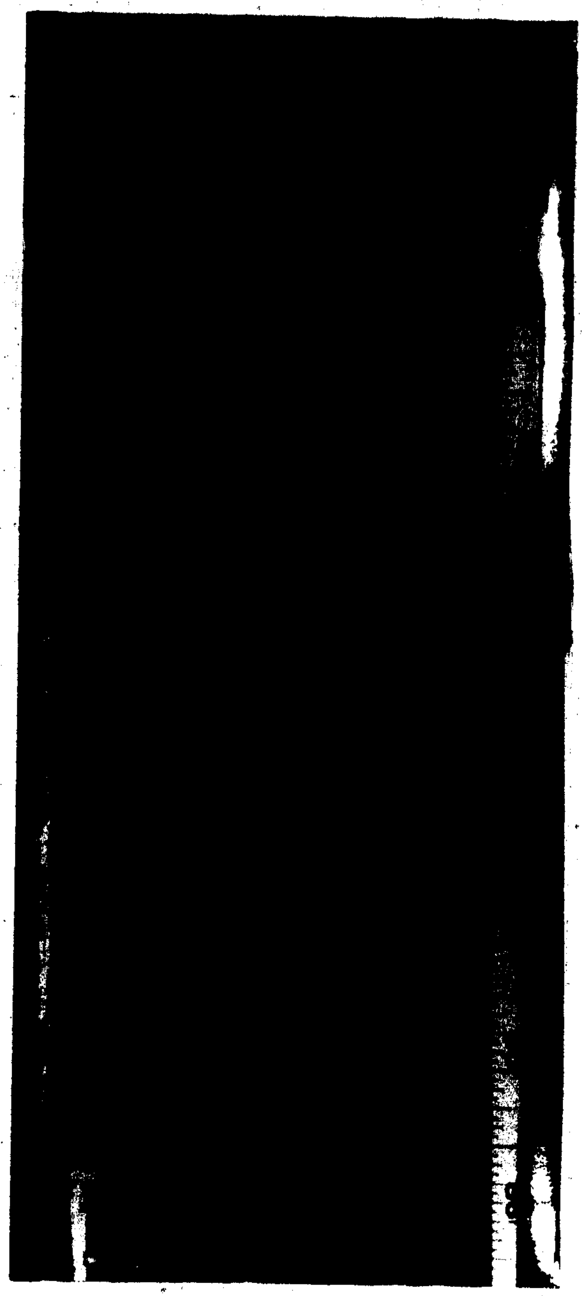


Figure IV.8. Rhythmites of unit C, core H6C2. Inside scale is in inches, outside in cm. Note the severe deformation of the laminations. Arrow points to fine sand at the base of one rhythmite.

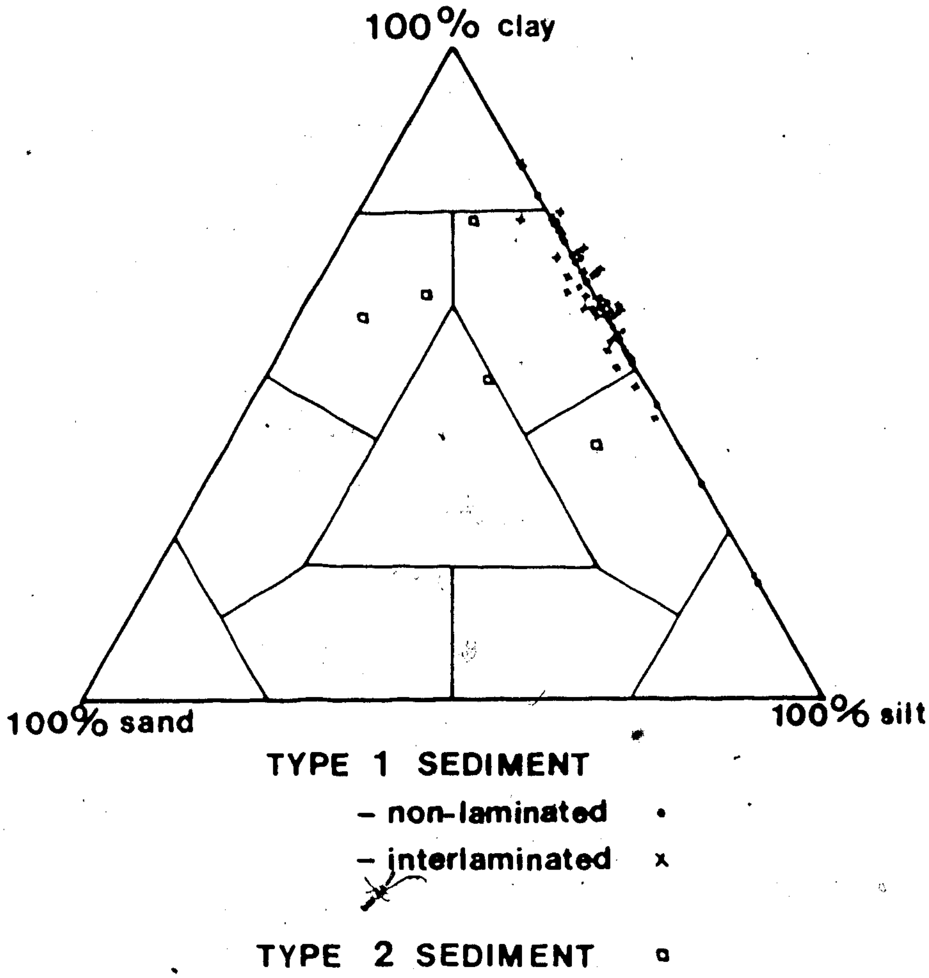


Figure IV.9. Ternary plot for all grain-size samples analyzed. Points plotted outside of the axes contain 0% sand.

TABLE IV.1

Summary Grain Size Statistics and Comparison With Other Studies

Sediment Class	Subclass	Mean (Phi)	SD (Phi)	Skew. (Phi)	Kurt. (Phi)	%Sand	%Silt	%clay
<u>Type 1</u>	All samples							
Mean		8.69	1.95	0.23	1.18	0.5	40.8	58.7
Min.		7.25	0.75	-0.11	0.83	0.0	82.0	18.0
Max.		9.70	2.44	0.35	2.11	4.0	18.0	82.0
n=51								
	<u>Non-laminated</u>							
Mean		8.55	1.82	0.23	1.25	0.0	44.5	55.5
Min.		7.25	0.75	-0.11	0.94	0.0	82.0	77.0
Max.		9.61	2.29	0.35	2.11	0.0	23.0	18.0
n=24								
	<u>Interlaminated</u>							
Mean		8.81	2.06	0.24	1.11	0.9	37.6	61.5
Min.		7.82	1.53	-0.03	0.83	0.0	18.0	42.0
Max.		9.70	2.44	0.33	1.33	4.0	57.0	82.0
n=27								
<u>Type 2</u>								
Mean		7.63	3.00	0.04	1.00	20.6	33.2	46.2
Min.		6.95	2.49	-0.06	0.73	11.0	21.0	67.0
Max.		9.11	3.39	0.19	1.36	34.0	54.0	35.0
n=5								
Stewart (1988)								
<u>Cyclopels</u>								
Mean		7.7	2.0	0.19	0.8	1.7	54.6	43.8
SD		0.8	0.2	0.15	0.1	3.3	12.4	13.1
Min.		5.9	1.6	-0.19	0.7	0.01	32.0	17.0
Max.		8.8	2.7	0.62	1.4	13.0	81.0	64.0
n=34								
Mackiewicz <u>et el.</u> (1984)								
<u>Type I</u>								
Min.		7.17	1.63			0.1		
Max.		8.65	3.15			11.2		

type 1 sediments from units A and B (non-laminated) to those from unit C (interlaminated) reveals that, although the units have a similar mean grain size, the interlaminated sediments show a lower degree of sorting and commonly have a fine sand component (0 - 4%, Table IV.1). Type 2 sediments are characterized by: a relatively high sand content (11 - 34%, Fig.IV.9), the presence of coarse and medium sand (<1%), and a strongly bimodal grain-size distribution (Fig. IV.10). These sediments occur only rarely in the samples (n=5), and are found in both units A and C. Two short cores from site H4 show loading of fine-grained sediments by a surface unit of massive type 2 sediments (Fig. IV.11). Type 2 sediments from near the base of H7C1 are normally graded, and show a sharp basal contact.

Ice-rafted debris (IRD), defined as clasts coarser than -1.0 ø (cf. Mackiewicz et al. 1984), was visually evident in most cores but not present in any of the samples analyzed. A layer containing abundant IRD, with clasts up to 3 cm in length (Fig. IV.12), was found in all three cores from sites H6 and H7. The layer occurs at 4 - 6.5 cm depth in core H7C1 and at 6 - 9 cm depth in the two cores from site H6. The concentration of IRD is considerably higher in core H6C1 than core H6C2. Elsewhere, IRD is present only as isolated granules to small pebbles, and overall is a surprisingly small component of the sediments recovered. The fine-grained fraction of IRD is not distinguishable (cf. Molnia 1983).

Correlation between cores

The correlation on the basis of colour and structure between core H7C1 and cores from site H6 has already been noted (Fig. IV.6). Evaluation of the lateral variability in grain size is possible at site H6,

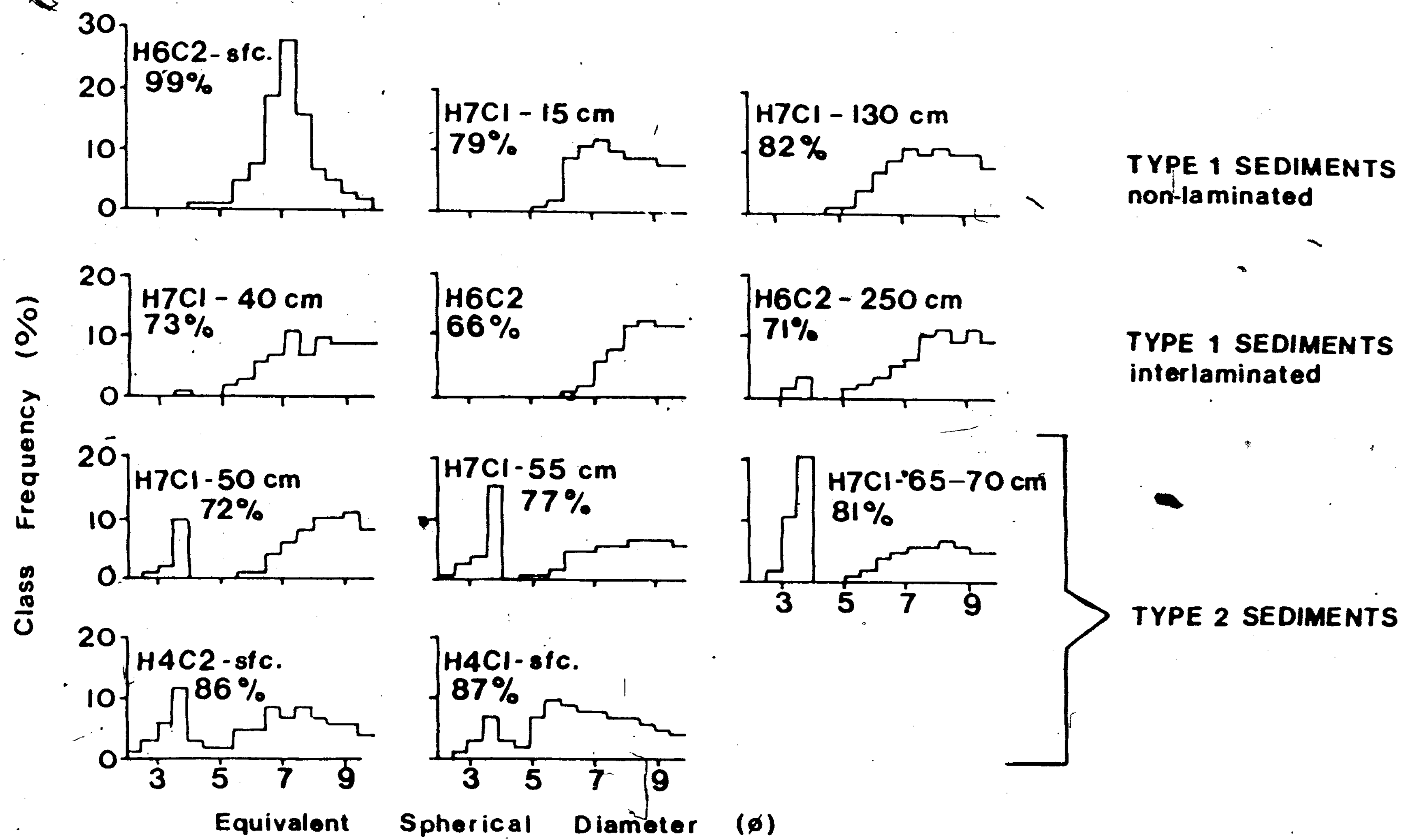


Figure IV.10. Class frequencies from grain-size analysis for all type 2 sediment samples and selected samples of type 1 sediment. Class interval is 1 ø.

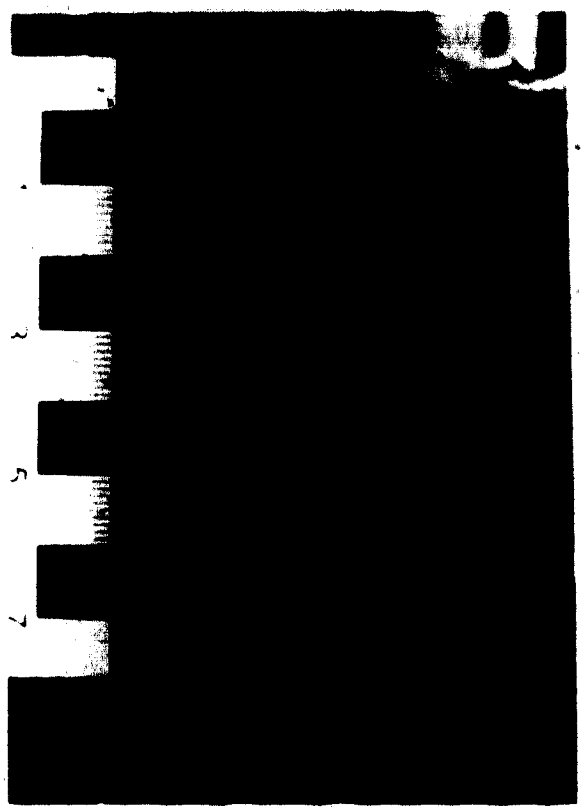


Figure IV.11. Type 2 sediments which form the upper 1.3 - 1.5 cm of core H4C2. Scale is in cm.

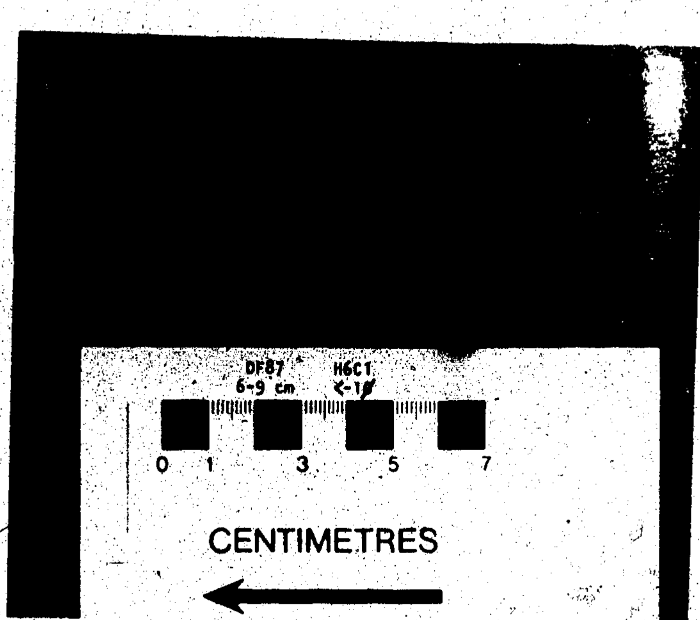


Figure IV.12. Ice-rafted debris (coarser than -1ø) from 6 - 9 cm depth in core H6C1 (see Fig. IV.7).

where two cores and two grab samples were recovered. The grain size envelope for the surface sediments of these four samples accounts for most of the variability observed between the coarsest and finest samples (defined by mean grain size) analyzed in this study (Fig. IV.13). Given this high lateral variability, little significance can be placed on either the down-core or between-core differences in grain size observed in Fig. IV.6. Nonetheless, it is noteworthy that sediments from site H7 were consistently coarser than those from site H6, despite the fact that H7 is located ca. 500 m further downfiord (Fig. IV.4). The mean grain size for units A and B (non-laminated) sediments from core H7C1 is 8.4ø (n=8), whereas for core H6C1 it is 9.0ø (n=8). A similar pattern is observed in the type 1 sediments of unit C, where mean grain size is 8.6ø in core H7C1 (n=4) and 8.9ø over the same depth of H6C2 (n=2). Fine sand is present at the base of laminations < 1 cm below the unit B/C transition in core H7C1, yet not found until 65 cm below the same transition in core H6C2.

Interpretation

The interpretation of the sediment cores is hindered by a lack of absolute dating control. Radiocarbon dating of total organic content (TOC) was rejected given the very low organic carbon content of the sediment (mean 0.76%, maximum 1.00%, n=13). A TOC of at least 6% is considered necessary to produce a meaningful radiocarbon date (R. McNeely, personal communication, 1988). Additionally, the presence of carbonate rocks (Trettin 1981) and Tertiary (?) wood (Blake 1987) in the drainage basin would make any TOC date from the fiord questionable.

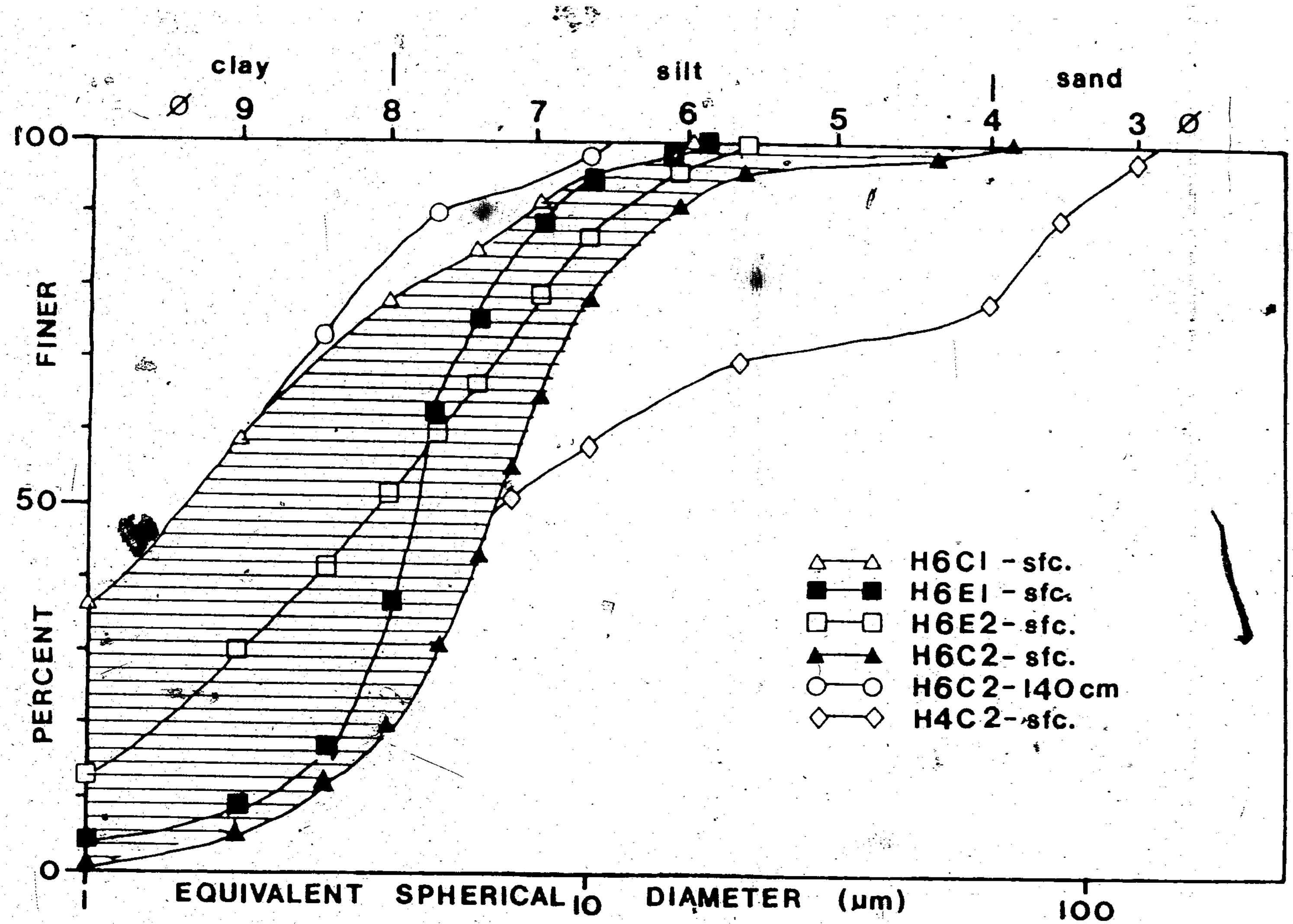


Figure IV.13. Grain-size envelope defined by four surface samples from site H6, compared to the coarsest and finest samples analyzed in the study.

Other techniques such as ^{210}Pb and ^{137}Cs, which are applicable to sediments < ca. 150 years in age (Oldfield 1977), are unlikely to be useful as there is evidence of extensive bioturbation in the upper sediments. The following interpretations are based on the assumption that major changes in colour and structure within the sediment, which are correlative between cores, reflect changes in the oceanography of the fiord and/or the rate of sediment input to the fiord. These changes are ultimately controlled by climate.

Type 2 sediments, those which contain >10% sand, occur only rarely in the samples and are not correlative between cores. Therefore they do not concern the interpretation of units A, B and C. However, these sediments do record infrequent and apparently random depositional processes. All type 2 sediments are interpreted as sediment gravity flow deposits. The massive surface sediments, and associated load structures in the underlying fine-grained deposits, found in the cores from site H4 likely represent the liquified or fluidized flow (Nardin *et al.* 1979, [illegible]ve 1979) of deltaic sediments down the fiord wall. While deposits related to these same processes are commonly described from other fiords, they generally contain much less fine-grained sediment (e.g. Gilbert 1982). The normally graded type 2 sediments near the base of core H7C1 likely represent turbidity current deposits, although unequivocal evidence of current flow is absent. Given the steep slopes which characterize Disraeli Fiord, the paucity of sediment gravity flow deposits in the cores is curious. This likely reflects the low sedimentation rates which presently characterize the fiord (see Discussion). However, it should be noted that samples were not obtained from sites distal to major deltas, where sediment gravity flows would be most common.

Units A and B

The massive to diffusely-laminated type 1 sediments of units A and B (Fig. IV.6) are the product of suspension settling. Such sediments had been intuitively proposed to dominate Disraeli Fiord by Syvitski (1986). Flocculation is likely an important process in promoting the deposition of these silt and clay particles (cf. Kranck 1975; Syvitski and Murray 1981).

The absence of wind-induced turbulence, owing to the presence of multi-year, landfast sea-ice, and significant tidal effects combine to produce a very low-energy contemporary environment in Disraeli Fiord. The blocking of freshwater drainage by the Ward Hunt Ice Shelf has produced a strong density stratification in the upper 50 m of the fiord, with the result that interflow is likely the dominant inflow process (cf. Syvitski 1986). Underflow is all but impossible in the absence of subglacial meltwater (cf. Powell 1980; Gilbert 1983; Mackiewicz et al. 1984), and overflow is also unlikely given the low temperature of the surface water (<1°C, Keys 1978). Prior to the growth of the ice shelf and development of a significant freshwater cap on the fiord, overflow (or shallow interflow) was undoubtedly the dominant inflow type. Unfortunately, the great depth of the fiord makes it impossible to distinguish between deposits related to overflow and interflow. The efficiency of bioturbation in creating an almost homogenous sediment reflects low sedimentation rates, despite the proximity of glacier ice.

Ice-rafted debris is present in both units A and B and may have been transported by sea ice and/or glacier ice. Although neither of these process are significant at present, due to the landfast sea-ice, IRD was found at <1 cm depth in two samples. The only concentrations of IRD are found in unit A of cores from sites H6 and H7. The texture and colour of the fine-grained matrix within this layer, which is 2.5 - 3 cm thick, is the same as both the over and underlying sediments. All the clasts are of the same lithology, a weakly lithified siltstone which outcrops near the fiord head. This suggests that transport occurred by glacier ice. Given the proximity of the cores it is possible that the IRD is related to a single ice berg dump (cf. Thomas and Connell 1985). However,it is equally likely that this layer records an extended interval of more frequent ice rafting.

Interpretation of the colour change between units A (reddish gray) and B (reddish brown) is difficult owing to the complexity of the biogeochemical environment of fiords (cf. Syvitski et al. 1986). The sediments of unit B appear to be more oxidized, and factors which may account for this difference include changes in: i) sedimentation rates, ii) biological productivity, and iii) circulation within the fiord. Although the mixing of surface waters would have been reduced to zero as a result of the growth of the Ward Hunt Ice Shelf and the establishment of a multi-year, landfast sea ice-cover, any attempt to explain these changes in colour in terms of fiord circulation remains speculative.

Unit C

The basal unit (unit C) of cores H6C2 and H7C1 is composed of interlaminated type 1 sediment. The change from non-laminated to interlaminated sediments at the unit B/C boundary suggests that the processes responsible for deposition of the interlaminated sediments (rhythmites) were no longer an important component of fiord sedimentation and/or sedimentation rates were considerably higher (or less bioturbation) at the time the interlaminated sediments were deposited. The presence of very fine sand at the base of many of these normally graded laminations suggests that they are not the product of "normal" suspension settling, such as characterizes the fiord at present. Three alternative processes; turbid plume transport, turbidity currents, and wave transport, will be considered to account for the interlaminated sediments.

Turbid plume transport. Deposits similar to those from the basal unit of the Disraeli Fiord cores have been described within contemporary glaciomarine sediments from SE Alaska. These contemporary sediments are interpreted as having been transported by subglacial meltwater which, upon entering the more dense marine waters, rises towards the surface with subsequent sediment transport occurring as overflow, or interflow, within turbid plumes (Fig. IV.14; Mackiewicz et al. 1984; Powell and Cowan 1987). Deposition of sediments transported within these turbid plumes occurs through suspension settling, producing normally graded laminae termed cyclopels (Mackiewicz et al. 1984). In Alaska, this process is capable of transporting fine sand up to 2.6 km beyond the glacier margin (Powell and Cowan 1986). Stewart (1988) has identified cyclopels within deglacial raised marine sediments from Clements Markham Inlet, ca. 80 km east of Disraeli Fiord (Fig. IV.1).

If the rhythmites in unit C of the Disraeli Fiord cores are cyclopels the apparent absence of other deposits (such as homogenous muds related to "normal" suspension settling) is curious. For example, of 23 Ekman samples obtained by Powell and Cowan (1987) from McBride Inlet, Alaska, only one was found to be composed entirely of cyclopels. Bergstone mud and homogenous mud formed a significant component of the more distal samples from that area. In the Disraeli Fiord cores such muds are not

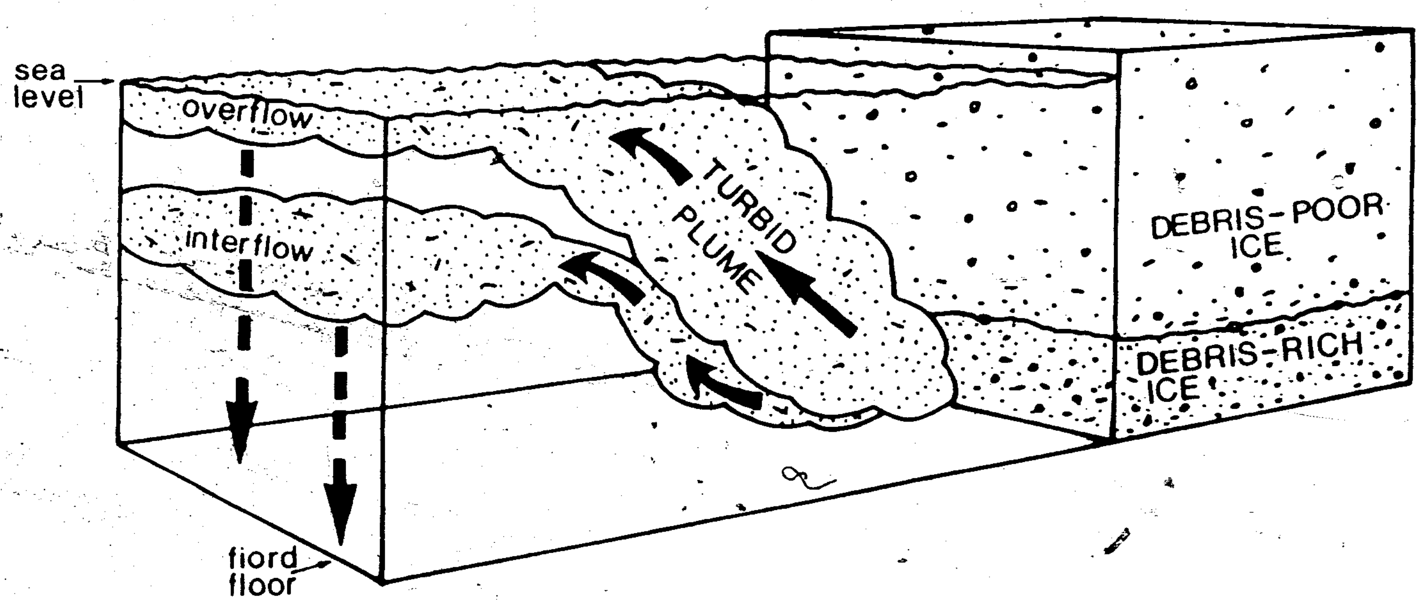


Figure IV.14. Schematic diagram of turbid plume transport at the margin of a tidewater glacier. Suspension settling from overflow and interflow plumes is responsible for the deposition of cyclopels (modified from Stewart 1988).

identifiable within unit C. This may result from muds deposited through "normal" suspension settling being indistinguishable from the fine-grained component of graded laminae related to turbid plume transport, or it may simply indicate the overwhelming dominance of turbid inflow in sediment transport. Grain size distributions of the interlaminated sediments in the Disraeli Fiord cores overlap with, but are consistently finer than, those reported for cyclopels within ice-proximal environments of SE Alaska (Mackiewicz et al. 1984) and northern Ellesmere Island (Stewart 1988; Table IV.1).

Turbidity currents. Individual graded laminae may represent thin-bedded turbidites (cf. Nilsen et al. 1980), but there is no unequivocal evidence of current flow (e.g. cross-laminations) in any of the laminae. Additionally, the absence of deposits which could be associated with suspension settling, which must be assumed to be an active process, makes this alternative less likely. In order to account for all of the rhythmites present in unit C turbidity currents would have to be associated with essentially continuous underflow, which is extremely unlikely in a marine environment (cf. Gilbert 1984). However, as it is difficult to distinguish between cyclopels and thin-bedded turbidites sedimentologically (Stewart 1988), it is possible that the unit C sediments consist of interlaminated turbid plume and turbidity current deposits which have not been differentiated.

Wave activity. Wave erosion and transport are processes which are presently inactive within the fiord. However, they must have been more important prior to the establishment of its landfast sea-ice. Nevertheless, wave action alone is considered insufficient to explain deposition of the rhythmites because: i) the steep walls of the fiord expose very little area to wave erosion; ii) the slopes of the inner fiord are dominated by coarse talus, with little fine grained sediment; and iii) waves are an unlikely mechanism for transporting fine sand several kilometres from coastal areas, especially given the long and narrow configuration of the fiord.

It is therefore concluded that the rhythmites within unit C of the Disraeli Fiord cores represent cyclopels, which by definition are the product of turbid plume transport and suspension settling (Mackiewicz et al. 1984). Critical to this interpretation is the presence of very fine sand at the base of many of the normally graded laminae. This is thought to preclude the possibility that rhythmites represent cyclical variability in "normal" suspension settling processes. While the possibility that some rhythmites represent thin-bedded turbidites cannot be discounted there is no unequivocal evidence to support this interpretation. The cyclopel interpretation suggests that all type 1 sediments are the product of suspension settling, contrasted with type 2 sediments which are the products of sediment gravity flows.

Although turbid plumes are observed in fiords where glaciers do not contact the sea (Burrell and Matthews 1974; Powell 1981; Gilbert 1982), the presence of sand within many of the cyclopels requires the direct inflow of subglacial meltwater and indicates that a glacier was grounded in the fiord head at the time these sediments were deposited. The fact that the cyclopels generally thin and fine upwards to the unit C/B boundary suggests that this glacier was retreating. However, the coarser nature of the sediments at site H7 compared to the more proximal site H6 apparently contradicts this relationship between grain size and the proximity of the glacier. As to why the H7 sediments are coarser remains unexplained. The scarcity of IRD within unit C sediments likely reflects the relatively high sedimentation rates associated with turbid plume transport. The abrupt change from laminated to non-laminated sediments at the boundary between units C and B likely represents a decrease in sedimentation rates (e.g. Lemmen et al. 1988), which in turn reflects changes in glacier extent and volume of meltwater discharge (Andrews 1987). Sedimentation rates could decrease sharply if the continued retreat of glaciers resulted in the formation of new sediment traps up-fiord of the core sites (cf. Andrews 1987). Alternatively, the change in structure at the unit C/B boundary may reflect a decrease in the volume and sediment concentration of glacial meltwater entering the fiord, in response to climatic deterioration and decreased ablation.

Discussion

Chronology and sedimentation rates

The interpretation of the rhythmites in unit C of the sediment cores as cyclopels provides indirect evidence regarding the age of the sediment. The cyclopels are concluded to relate to the retreat of a glacier grounded in the fiord head, and there is no evidence to suggest that such retreat has occurred recently. Conversely, both the Disraeli Glacier and Glacier "A" (Fig. IV.1) are presently advancing, as evidenced by ice-thrust sediments along the grounded margin of the glacier tongues (Fig. IV.15). The absence of similar ice-thrust sediments beyond the present glacier margins suggests that these glaciers are at their maximum positions since deglaciation. Therefore retreat of grounded ice in the fiord head has not

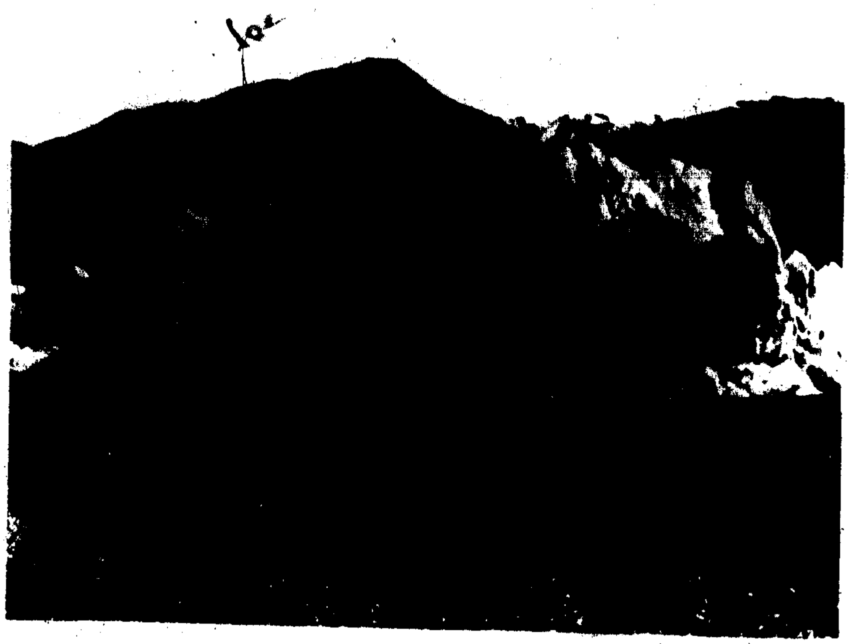


Figure IV.15. Ice-thrust sediments along the grounded margin of the actively advancing Glacier "A". Mark Tushingham is circled for scale.

occurred since deglaciation from the last glacial ice limit, which began locally about 8,000 BP (Chapter II). The cyclopels are believed to have been deposited during this interval.

Knowing the date of the transition from the interlaminated to non-laminated sediments (unit C/B boundary) would allow for a first-order estimation of recent sedimentation rates. If this boundary relates to the retreat of ice out of the fiord and formation of new sediment traps, a conservative estimate of its age would be late deglaciation, ca. 7000 BP. Alternatively, if this boundary relates to decreased sediment input in association with a climatic deterioration (colder and/or drier conditions), it is reasonable to assume that this event would be recorded by other proxy climatic data from the region. Studies of the Ward Hunt Ice Shelf and the Holocene driftwood record from the north coast of Ellesmere Island both suggest a marked regional climatic deterioration ca. 4500 BP (Lyons and Mielke 1973; Stewart and England 1983; Jeffries 1985). This date is tentatively believed to be a minimum estimate on the age of the unit C/B boundary. Assuming that this provisional chronology is correct, and given that the zone C/B boundary occurs at a depth of 31 and 42 cm in cores H7C1 and H6C2, respectively (Fig. IV.6), then mean sedimentation rates at these sites since this boundary have been approximately 0.05 - 0.1 m/ka. These estimates are the lowest reported for a fiord environment anywhere in the world (see Eyles et al. 1985 Table 1). They are also the first reported from a fiord with landfast sea-ice and adjacent cold-based glaciers. Sedimentation rates in the high latitude fiords of Spitzbergen, which are seasonally ice-free and where both warm- and cold-based glaciers are present, are as high as 1000 m/ka (Elverhoi 1984; Elverhoi et al. 1983).

Deglacial sedimentation and paleoclimate

Stewart's (1988) sedimentological investigation of deglacial raised marine sediments at the head of Clements Markham Inlet (Fig. IV.1) documents a high energy, ice-proximal environment characterized by several lithofacies. The abundance of subglacial outwash and turbid plume deposits (cyclopels and cyclopsams) led Stewart (1988) to conclude that deglacial sedimentation was dominated by the inflow of sediment-laden subglacial meltwater from the de[illegible]rich basal ice of a tidewater glacier. This contrasts sharply with th[illegible]sed (subpolar) glaciers which characterize the region at present ([illegible]ller 1976), whose meltwater is restricted to lateral and supraglacial streams. Similar cold-based conditions are believed to have also characterized the region during full glacial times (cf. England 1986, 1987a). Therefore, Stewart (1988) concluded that the processes of sedimentation recorded in the deglacial deposits reflected climatic conditions considerably warmer than those of both the full glacial and the present time.

The cyclopels in unit C of the Disraeli Fiord sediment cores record a rather simple depositional environment in comparison to the deglacial sediments from Clements Markham Inlet. This apparent simplicity suggests that the Disraeli Fiord cores failed to penetrate all of the deglacial sediments present. This is supported by the fact that the cyclopels within unit C are consistently finer grained than those described by Stewart (1988), suggesting that they were deposited in a more ice-distal environment than those from Clements Markham Inlet. Certainly part of the difference between the two studies is due to the limited spatial record which can be obtained through coring. Nonetheless, the paleoclimatic implications of this study echo those of Stewart (1988). Subglacial meltwater is believed essential to account for the deposition of the cyclopels, and such meltwater could only be generated in association with a marked climatic amelioration. Such an amelioration is evidenced by geomorphic data indicating less severe sea-ice conditions during deglaciation (Evans and Lemmen 1987; Chapter II), and is consistent with an episode of very rapid glacier retreat after about ca. 8000 BP along the north coast of Ellesmere Island (Bednarski 1986; Chapter II).

The temporal significance of the cyclopels in unit C of the Disraeli Fiord cores is unknown. In SE Alaska it has been shown that laminations are produced at a rate of greater than twice a day, controlled primarily by tidal activity (Mackiewicz et al. 1984). Given the small amplitude of tides within Disraeli Fiord (<1 m) this is unlikely to be an important factor. As diurnal variability is negligible in this high arctic environment undoubtedly the most pronounced cyclicity in sediment input to the fiord is annual. Although annual laminations have been identified in glaciomarine sediments (e.g. Stevens 1985; Jennings 1986), in the absence of absolute dating control it cannot be concluded that laminations in the Disraeli Fiord sediments are varves. More likely, they represent incursions of subglacial meltwater which could have occurred at any point during the melt season, as controlled by synoptic-scale weather conditions.

<u>Implications for Quaternary studies in the high arctic</u>

This study has shown that it is possible to attain an extensive temporal record of deposition from high arctic fiords using simple coring techniques. This record has not previously been utilized despite its complementary value to the terrestrial record. The low sedimentation rates which have been suggested for the latter part of the Holocene within Disraeli Fiord may also be typical of other fiords in the region. This low sedimentation rate means that even relatively short cores (<1 m) will likely penetrate deglacial sediments. Additionally, the fiord record does provide higher resolution on late Holocene climatic changes than do cores from the Arctic Ocean Basin.

It is likely that most of the high arctic fiords are dominated by deglacial sediments, reflecting sedimentation rates two or more orders of magnitude greater than at present. Deglacial sedimentation rates off the east coast of Baffin Island were significantly higher than during the rest of the Holocene (Andrews et al. 1985), in what Andrews (1987, p.236) terms "the marine expression of 'paraglaciation'". As sedimentation rates tend to decrease rapidly away from the ice front in most glacial marine environments (cf. Powell 1981, 1984) it is possible that a considerable thickness of deglacial sediments, perhaps many tens of metres, are present at sites which bordered the last ice limit (cf. Stewart 1988). These deglacial sediments contain a valuable record of processes which can be integrated with those recorded above present sea level. Interpretations of deep fiord sediments from Disraeli Fiord and raised marine sediments from Clements Markham Inlet lead to similar conclusions, with the former recording (relatively) ice-distal facies which likely relate to ice-proximal facies defined by Stewart (1988). Both studies indicate that a pronounced climatic amelioration occurred during deglaciation, a conclusion that is supported by geomorphic and driftwood data from the adjacent terrestrial environment (Evans and Lemmen 1987; Chapter II). The sediment record, and its attendant glaciological argument favouring warm-based glaciers, implies a magnitude of climatic change that was previously unavailable from these terrestrial records, and strongly supports the idea of an early Holocene climatic optimum in the region (cf. Evans and Lemmen 1987).

The model of the full glacial sea (England 1983) suggests that many fiords in the high arctic remained free of trunk glaciers throughout the last glaciation. However, the identification of marine sediments that relate to full glacial conditions has been difficult (England 1987b, 1987c). If Disraeli Fiord is presently an appropriate analogue for the full glacial sea, it is predicted that full glacial sedimentation would be dominated by massive silts and clays, possibly with extensive bioturbation. Potentially most diagnostic in the sediment record would be the onset of local deglaciation, characterized by higher sedimentation rates, preservation of sedimentary structures and possibly an influx of coarser sediments. The cores from Disraeli Fiord clearly demonstrate that the contemporary ice-shelf environment is not an appropriate analogue for the deglacial environment at the same site. Therefore sedimentological distinction between full glacial and deglacial sediments should be possible. Coring of the fiords, at sites that lie several km beyond the last ice limit, may allow access to sediments of full glacial age. Obtaining such sediments would provide new insights into the paleoenvironmental record of the high arctic, and should become a key objective of future studies.

Conclusions

Access to the marine record in the arctic continues to present technical difficulties, and this challenge is only aggravated in the high arctic where access by ships is made difficult by severe sea-ice conditions. The SAFE project (see Syvitski and Schafer 1985), studying the fiords of eastern Baffin Island, has greatly expanded our knowledge of sedimentary processes in glaciomarine environments and made several contributions to Quaternary studies (e.g. Gilbert 1985; Hein and Longstaffe 1985; Andrews et al. 1985, 1986). Projects of similar scale should be developed for some areas of the high arctic (e.g. Greely and Archer fiords, England 1987a). However, the ice shelves and thick pack ice off the north coast of Ellesmere Island require that future studies here employ new approaches. Priority should be placed on a better understanding of the present environment, especially with regard to sediment budgets. Interpretation of the sediment record without detailed information on bathymetry will remain largely speculative. Regarding glacial history and paleoenvironmental change, the fiords contain an important sediment record which likely predates the last glaciation. Clearly these sediments represent an important complement to both the terrestrial and Arctic Ocean records.

The following conclusions can be made from this study, based upon the assumption that the interlaminated sediments of cores H6C2 and H7C1 are deglacial in age.

1 - Disraeli Fiord is a very low energy depositional environment, even at sites proximal to glaciers. Sediment enters the fiord by interflow and is deposited through suspension settling. Ice rafting is precluded by multi-year, landfast sea-ice.

2 - Sedimentation rates in the fiord since deglaciation have been about 0.05 to 0.1 m/ka. These low rates are the first reported from a fiord environment which features cold-based glaciers.

3 - Changes in sediment structure through time are dominated by changes in the rate of sediment input, which in turn reflect glacioclimatic

changes. Although changes in fiord circulation have occurred during the Holocene it is not clear how those changes are reflected in the sediment record.

4 - During deglaciation sedimentation rates were many times greater than at present. Subglacial meltwater was the dominant source of inflow. Sediment was transported by turbid plumes as over- or interflow, and deposited through suspension settling. This interpretation necessitates that glaciers were warm-based and implies a pronounced climatic amelioration at this time (Stewart 1988).

5 - It should be possible to distinguish between deglacial and full glacial sediments within cores. Such studies are important in testing the model of a full glacial sea occupying many high arctic fiords and channels throughout the last glaciation (England 1983). Sediments spanning the last glaciation would provide new insights concerning paleoenvironmental changes in high latitudes.

Bibliography

Andrews, J.T. 1985. Grain size characteristics of Quaternary sediments, Baffin Island region. In Quaternary Environments: Baffin Island, Baffin Bay, and West Greenland. Edited by, Andrews, J.T. Allen & Unwin, Boston, Mass., 124-153.

________. 1987. Late Quaternary marine sediment accumulation in fiord-shelf-deep-sea transects, Baffin Island to Baffin Bay. Quaternary Science Reviews, 6, 231-243.

Andrews, J.T., Jull, A.J.T., Donahue, D.J., Short, S.K., and Osterman, L.E. 1985. Sedimentation rates in Baffin island fiord cores from comparative radiocarbon dates. Canadian Journal of Earth Sciences, 22, 1827-1834.

Andrews, J.T., Mithersill, J.S., and Tabrez, A.R. 1986. Paleomagnetic record, texture and mineralogy of Late Quaternary sediments, Baffin Island Fiords, N.W.T., Canada. Arctic and Alpine Research, 18, 361-376.

Bednarski, J. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 23, 1343-1355.

________. 1988. The geomorphology of glaciomarine sediments in a high Arctic fiord. Geographie physique et Quaternaire, 42, 65-74.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, 7, 634-664.

Bundy, L.G., and Bremner, J.M. 1972. A simple titrimetric method for determination of inorganic carbon in soils. Proceedings of the Soil Science Society of America, 36, 273-275.

Burrell, D.C., and Matthews, J.B. 1974. Turbid outwash Fiords. In Coastal Ecological Systems of the United States, III. Edited by, Odum, H.T., Copeland B.J., and McMahan, E.A. The Conservation Foundation, Washington, D.C., 11-16.

Christie, R.L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. Geological Survey of Canada Memoir 331, 79 p.

Clark, D.L., Whitman, R.R., Morgan, K.A., and Mackey, S.D. 1980. Stratigraphy and glacial-marine sediments of the Amerasian Basin, central Arctic Ocean. Geological Society of America, Special Paper 181, 57 p.

Clark, D.L., and Hanson, A. 1983. Central Arctic Ocean sediment textures - a key to ice transport mechanisms. In Glacial-marine Sedimentation. Edited by Molnia, B.F. Plenum Publishing, New York, 301-329.

Crary, A.P. 1956. Geophysical studies along northern Ellesmere Island. Arctic, 9, 155-165.

__________. 1960. Arctic ice island and ice shelf studies. Part II. Arctic, 13, 32-50.

Dalrymple, R.W., and Maass, O.C. 1987. Clay mineralogy of late Cenozoic sediments in the CESAR cores, Alpha Ridge, central Arctic Ocean. Canadian Journal of Earth Sciences, 24, 1562-1569.

Elverhoi, A. 1984. Glacigenic and associated marine sediments in the Weddell Sea, fjords of Stitxbergen and the Barents Sea: a review. Marine Geology, 57, 53-88.

Elverhoi, A., Lonne, O., and Seland, R. 1983. Glaciomarine sedimentation in a modern fjord environment, Spitzbergen. Polar Research, 1, 127-149.

England, J. 1976. Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: alternative models. Quaternary Research, 6, 185-202.

__________. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, 20, 895-917.

__________. 1986. Glacial erosion of a High Arctic valley. Journal of Glaciology, 32, 60-64. ?

__________. 1987a. Glaciation and evolution of the Canadian high arctic landscape. Geology, 15, 419-424.

__________. 1987b. In situ shells of full glacial age beneath former ice shelves. 16th Annual Arctic Workshop; Boreal Institute for Northern Studies, Abstracts, 20-22.

__________. 1987c. The late Quaternary history of Hall Land, northwest Greenland: reply. Canadian Journal of Earth Sciences, 24, 374-380.

England, J., Lemmen, D.S., Evans, D.J.A., and Bednarski, J. 1987. The Quaternary of northern Ellesmere Island. International Union for Quaternary Research, XIIth International Congress, Ottawa, Canada. Programme with Abstracts, p. 162.

Evans, D.J.A., and Lemmen, D.S. 1987. Indicators of Holocene climatic change from Wootton Peninsula/Phillips Inlet and Marvin Peninsula/Disraeli Fiord, Northern Ellesmere Island. International Union for Quaternary Research, XIIth International Congress, Ottawa, Canada. Programme with Abstracts, p. 163.

Eyles, C.H., Eyles, N., and Miall, A.D. 1985. Models of glaciomarine sedimentation and their application to the interpretation of ancient glacial sequences. Palaeogeography, Palaeoclimatology, Palaeoecology, 51, 15-84.

Folk, R.L. 1980. Petrology of Sedimentary Rocks. Hemphill Publishing Co., Austin, Texas, 182 p.

Ford, W.L., and Hattersley-Smith, G. 1965. On the oceanography of the Nansen Sound fiord system. Arctic, 18, 158-171.

Gilbert, R. 1982. Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: glaciomarine processes in fiords of eastern Cumberland Peninsula. Arctic and Alpine Research, 14, 1-12.

________. 1984. Sedimentary processes of Canadian arctic fiords. Sedimentary Geology, 36, 147-175.

________. 1985. Quaternary glaciomarine sedimentation interpreted from seismic surveys of fiords on Baffin Island, N.W.T. Arctic, 38, 271-280.

Hattersley-Smith, G. 1973. Ice shelf and Fiord ice problems in Disraeli Fiord, northern Ellesmere Island, N.W.T. Defense Research Establishment Ottawa, Technical Note no. 72-34, 12 p.

Hattersley-Smith, G., and Serson, H. 1966. Reconnaissance oceanography over the ice of the Nansen Sound fiord system. Defence Research Board of Canada, Ottawa, Hazen 28, 13 p.

Hattersley-Smith, G., Fuzesy, A., and Evans, S. 1969. Glacier depths in northern Ellesmere Island: airborne radio echo sounding in 1966. Defence Research Board, Canada, Defence Research Establishment Ottawa Technical Note No. 69-6, 23 p.

Hein, F.J., and Longstaffe, F.J. 1985. Sedimentologic, mineralogic and geotechnical descriptions of fine-grained slope and basin deposits, Baffin Island Fiords. Geo-Marine Letters, 5, 11-16.

Hein, F.J., and Mudie, P.J. 1987. Facies description and sedimentary history, Canadian polar margin, offshore Axel Heiberg Island. International Union for Quaternary Research, XII[th] International Congress, Ottawa, Canada. Programme with Abstracts, p. 185.

Herman, Y., and Hopkins, D.M. 1980. Arctic oceanic climate in late Cenozoic time. Science, 209, 557-562.

Horne, E.P.W. 1985. Ice-induced vertical circulation in an arctic fiord. Journal of Geophysical Research, 90 (C1), 1078-1086.

Jeffries, M.O. 1985. Physical, chemical and isotopic investigations of Ward Hunt Ice Shelf and Milne Ice Shelf, Ellesmere Island, N.W.T. Ph.D. thesis, University of Calgary, 310 p.

Jeffries, M.O., and Krouse, H.R. 1984. Arctic ice shelf growth, fiord oceanography and climate. Zeitschrift für Gletscherkunde und Glazialgeologie, 20, 147-153.

Jeffries, M.O., Sackinger, W.M., Krouse, H.R., and Serson, H.V. 1988. Water circulation and ice accretion beneath Ward Hunt Ice Shelf (northern Ellesmere Island), deduced from salinity and isotope analysis of ice cores. Annals of Glaciology, 10, 68-72.

Jennings, A.E. 1986. Late Quaternary marine sediments from a transect of fiord and continental shelf environments: a study of piston cores from Clark Fiord and Scott Trough, Baffin Island, Canada. M.Sc. thesis, University of Colorado, Boulder, Colorado, 252 pp.

Keys, J.E. 1978. Water regime of Disraeli Fiord, Ellesmere Island, Canada. Department of National Defence Research Establishment, Ottawa (DREO), Report 792, 58 p.

Keys, J., Johannessen, O.M., and Long, A. 1969. The oceanography of Disraeli Fiord, northern Ellesmere Island. Defence Research Board, Ottawa, Geophysics Hazen 34, 7 p.

Kranck, K. 1975. Sediment deposition from flocculated suspension. Sedimentology, 22, 111-123.

Lake, R.A., and Walker, E.R. 1973. Notes on the oceanography of d'Iberville Fiord. Arctic, 26, 222-229.

________. 1976. A Canadian arctic fjord with some comparisons to fjords of the western Americas. Journal of the Fisheries Research Board of Canada, 22, 2272-2285.

Lemmen, D.S., Gilbert, R., Smol, J., and Hall, R.I. 1988. Holocene sedimentation in glacial Tasikutaaq Lake, Baffin Island. Canadian Journal of Earth Sciences, 25, 810-823.

Lemmen, D.S., Evans, D.J.A., and England, J. in press. Glaciers and the morphology and structure of Milne Ice Shelf, Ellesmere Island, N.W.T., Canada: discussion. Arctic and Alpine Research.

Lowe, D.R. 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. In Geology of Continental Slopes. Edited by Doyle, L.J., and Pilkey, O.H. Society of Economic Paleontologists and Mineralogists, Special Publication, 27, 75-82.

Lyons, J.B., and Mielke, J.E. 1973. Holocene history of a portion of northernmost Ellesmere Island, Arctic, 26, 314-323.

Mackiewicz, N.E., Powell, R.D., Carlson, P.R., and Molnia, B.F. 1984. Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. Marine Geology, 57, 113-147.

Morris, T.H., Clark, D.L., and Blasco S.M. 1985. Sediments of the Lomonosov Ridge and Makarov Basin: a Pleistocene stratigraphy for the North Pole. Geological Society of America Bulletin, 96, 901-910.

Molnia, B.F. 1983. Distal glacial marine sedimentation: abundance, composition and distribution of North Atlantic Pleistocene ice rafted sediment. In Glacial Marine Sedimentation. Edited by Molnia, B.F. Plenum Press, New York, 593-626.

Mudie, P.J., and Blasco, S.M. 1985. Core recovery and lithostratigraphy of the CESAR cores. In Initial Geological Report on CESAR - Canadian Experiment to Study the Alpha Ridge. Edited by Jackson, H.R., Mudie, P.J., and Blasco, S.M. Geological Survey of Canada, Paper 84-22, 59-99.

Mudie, P.J., Hein, F.J., and Van Wagoner, N.A. 1988. Marine geology studies on the Canadian polar margin. Geological Association of Canada, Joint Annual Meeting, Program with Abstracts, p. A88.

Muller, F. 1976. On the thermal regime of a high-arctic valley glacier. Journal of Glaciology, 16, 119-133.

Nardin, T.R., Hein, F.J., Gorsline, D.S., and Edwards, B.D. 1979. A review of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of slope systems versus canyon-fan-basin floor systems. In Geology of Continental Slopes. Edited by Doyle, L.J., and Pilkey, O.H. Society of Economic Paleontologists and Mineralogists, Special Publication, 27, 61-73.

Narod, B.B., Clarke, G.K.C., and Prager, B.T. 1988. Airborne UHF radar sounding of glaciers and ice shelves, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 25, 95-105.

Nilsen, T.H., Walker, R.G., and Normack, W.R. 1980. Modern and ancient submarine fans: discussion and replies. Bulletin of the American Association of Petroleum Geologists, 64, 1094-1113.

Oldfield, F. 1977. Lakes and their drainage basins as units of sediment-based ecological study. Progress in Physical Geography, 1, 460-504.

Powell, R.D. 1981. A model for sedimentation by tidewater glaciers. Annals of Glaciology, 2, 129-134.

__________. 1984. Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. Marine Geology, 57, 1-52.

Powell, R.D., and Cowan, E.A. 1987. Depositional processes at McBride Inlet and Riggs Glacier. In Observed Processes of Glacial Deposition in Glacier Bay, Alaska. Edited by, Anderson, P.J., Goldthwaite, R.P., and McKenzie, G.D. Byrd Polar Research Center. Miscellaneous Publication No. 236, 140-164.

Reasoner, M.A. 1986. An inexpensive, lightweight percussion core sampling system. Geographie physique et Quaternaire, 40, 217-219.

Retelle, M.J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. Canadian Journal of Earth Sciences, 23, 1001 -1012.

Stein, R. 1985. Rapid grain-size analysis of clay and silt fraction by Sedigraph 5000 D: comparison with Coulter Counter and Atterberg methods. Journal of Sedimentary Petrology, 55, 590-593.

Stevens, R. 1985. Glaciomarine varves in late-Pliestocene clays near Goteborg, southwest Sweden. Boreas, 14, 127-132.

Stewart, T.G. 1988. Deglacial-marine sediments from Clements Markham Inlet, Ellesmere Island, N.W.T., Canada. Ph.D. thesis, University of Alberta, Edmonton, Alberta, 229 p.

Stewart, T.G., and England, J. 1983. Holocene sea-ice variations and paleoenvironmental change, northernmost Ellesmere Island, N.W.T., Canada. Arctic and Alpine Research, 15, 1-17.

Syvitski, J.P.M. 1986. DNAG #2. Estuaries, deltas and fjords of eastern Canada. Geoscience Canada, 13, 91-100.

Syvitski, J.P.M., and Murray, J.W. 1981. Particle interaction in fjord suspended sediment. Marine Geology, 39, 215-242.

Syvitski, J.P.M., and Schafer, C.T. 1985. Sedimentology of arctic fiords experiment (SAFE): project introduction. Arctic, 38, 264-270.

Syvitski, J.P.M., Burrell, D.C., and Skei, J.M. 1986. Fjords: Processes and Products. Springer-Verlag, New York, 520 p.

Thomas, G.S.P., and Connell, R.J. 1985. Iceberg drop, dump, and grounding structures from Pliestocene glacio-lacustrine sediments, Scotland. Journal of Sedimentary Petrology, 55, 243-249.

Trettin, H.P. 1981. Geology of Precambrian to Devonian rocks, M'Clintock Inlet area, District of Franklin (NTS 340 E, H) - preliminary geological map and notes. Geological Survey of Canada, Open File Report 759.

________. 1987. Pearya: a composite terrane with Caledonian affinities in northern Ellesmere Island. Canadian Journal of Earth Sciences, 24, 224-245.

V. CONCLUSIONS

Evidence collected over the past decade in support of a limited ice cover, and the adjoining full glacial sea, during the last glaciation of northern Ellesmere Island is overwhelming. In addition to geomorphic/stratigraphic data (e.g. England 1978, 1983, 1987; England et al. 1978, 1981; Bednarski 1986; Retelle 1986; Evans 1988), recent ice core studies (Koerner et al. 1987) and the offshore sediment record (Mudie et al. 1988) also support the concept of limited ice extent throughout the Wisconsinan. Where models of more extensive ice cover are discussed there is a strong dependence upon undatable field evidence such as erratics and striae, and/or the assumptions that: i) ice extent can be predicted on the basis of postglacial emergence; and ii) retreat of cold-based glaciers can occur without leaving any geomorphic evidence of its former presence (e.g. Blake 1970, 1977; Lyons and Mielke 1973; Denton and Hughes 1981; Hodgson 1985; Tushingham and Peltier 1988). That these assumptions are invalid in many cases has already been demonstrated (e.g. England 1976, 1987; Dyke 1983; Stewart 1988).

During the last glaciation glaciers were largely restricted to the major valleys on northernmost Ellesmere Island. The reconstructed ice margins are explicable from a climatic /meteorologic perspective which suggests that weather patterns were not significantly different than at present, and that the Arctic Ocean did not serve as an efficient moisture source. Of particular importance are the radiocarbon dates on subfossil bryophytes (ca. 11 - 15 ka BP) which demonstrate that ice-free areas remained biologically viable throughout the last glaciation, even adjacent to glaciers.

Much of the field evidence used by Lyons and Mielke (1973) to support the concept of an extensive ice cover during the last glaciation has been shown to relate to older glacial events. Although chronological data relating to these earlier events are limited, it is suggested that the record from the field area shows regional correlations with northeastern Ellesmere Island (England et al. 1978 1981; Retelle 1986). AMS dates of ca. 27 - 32 ka on marine shells associated with this older glaciation are not considered to represent accurate, finite dates. To the contrary, sea levels and ice margins identified beyond the last ice limit may be pre-Wisconsinan in age. While it is clear that features relating to the last glaciation dominate both the geomorphic and stratigraphic record on northern Ellesmere Island the same is true for most arctic, and many temperate, regions. Despite the limited extent of the last glaciation in the area it is possibled that if other glacial advances occurred during the Wisconsinan, they were even less extensive. It is particularly appropriate to question the applicability of terms such as glacial/interglacial and stadial/interstadial in high latitudes.

It is believed that pervasive glaciation of the field area occurred at some time in the distant past, possibly during the early Quaternary or even late Tertiary. At that time the Arctic Ocean must have served as a more efficient moisture source. England (1987) has recently argued that the changing extent and style of glaciation through time is a product of tectonic changes in the landscape rather than climate. Evidence to directly support or reject England's ideas has not been found in this study, although it is thought that Disraeli Fiord is primarily tectonic in origin and not the product of selective linear erosion. It is argued that the role of the Arctic Ocean in controlling regional climate, and therefore glaciation, must also receive consideration in explaining changes in glacial style through time. An important test would be to demonstrate whether the earliest glaciations occurred prior to the establishment of a permanent sea-ice cover on the Arctic Ocean Basin.

Finally, this study represents the first attempt in the Queen Elizabeth Islands to investigate the valuable sediment record from the deep fiords. This preliminary work has demonstrated that such studies are both feasible and useful. It has been shown that the processes of deglacial sedimentation were unlike those of the contemporary glaciomarine environment, and this work supports Stewart's (1988) conclusion of warmer climatic conditions and the presence of temperate tidewater glaciers at that time. Sedimentation rates within the fiords of northern Ellesmere Island are likely lower than any in the world outside of Antarctica, and therefore the potential exists to obtain an extensive temporal record of sedimentation. Such information could contribute significantly to determining the age of the high arctic fiords. Using the contemporary environment of Disraeli Fiord as an analogue for full glacial conditions it should be possible to distinguish between full-glacial and deglacial sediments on the basis of sedimentological analysis. Priority in future Quaternary studies in the high arctic must be placed upon the integration of the marine and terrestrial records.

BIBLIOGRAPHY

Bednarski, J. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, 23, 1343-1355.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, 7, 634-664.

________. 1977. Glacial sculpture along the east-central coast of Ellesmere Island, Arctic Archipelago. Geological Survey of Canada, Paper 77-1C, 107-115.

Denton, G.H., and Hughes, T.J. 1981. The Arctic Ice Sheet: An outrageous hypothesis. In The Last Great Ice Sheets. Edited by Denton, G.H., and Hughes, T.J. Wiley-Interscience, New York, 437-467.

Dyke, A.S. 1983. Quaternary geology of Somerset Island, District of Franklin. Geological Survey of Canada Memoir 404, 32 p.

England, J. 1976. Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: alternative models. Quaternary Research, 6, 185-202.

________. 1978. The glacial geology of northeastern Ellesmere Island, N.W.T., Canada. Canadian Journal of Earth Sciences, 15, 603-617.

________. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, 20, 895-917.

________. 1987. Glaciation and evolution of the Canadian high arctic landscape. Geology, 15, 419-424.

England, J., Bradley, R.S., and Miller, G.H. 1978. Former ice shelves in the Canadian high Arctic. Journal of Glaciology, 20, 393-404.

England, J., Bradley, R.S., and Stuckenrath, R. 1981. Multiple glaciations and marine transgressions, western Kennedy Channel, Northwest Territories, Canada. Boreas, 10, 71-89.

Evans, D.J.A. 1988. Glacial geomorphology and late Quaternary of Phillips Inlet and the Wootton Peninsula, northwest Ellesmere Island, Canada. Ph.D. thesis, The University of Alberta, Edmonton.

Hodgson, D.A. 1985. The last glaciation of west-central Ellesmere Island, Arctic Archipelago, Canada. Canadian Journal of Earth Sciences, 22, 347-368.

Koerner, R.M., Fisher, D.A. and Paterson, W.S.B. 1987. Wisconsinan and pre-Wisconsinan ice thicknesses on Ellesmere Island, Canada: inferences from ice cores. Canadian Journal of Earth Sciences, 24, 296-301.

Lyons, J.B., and Mielke, J.E. 1973. Holocene histo[illegible] a portion of northernmost Ellesmere Island, Arctic, 26, 314-[illegible]

Mudie, P.J., Hein, F.J., and Van Wagoner, N.A. 19[illegible]. Marine geology studies on the Canadian polar margin. Geological Association of Canada, Joint Annual Meeting, Program with Abstracts, p. A88.

Retelle, M.J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, Northwest Territories. Canadian Journal of Earth Sciences, 23, 1001 -1012.

Stewart, T.G. 1988. Deglacial-marine sediments from Clements Markham Inlet, Ellesmere Island, N.W.T., Canada. Ph.D. thesis, The University of Alberta, Edmonton, Alberta.

Tushingham, M., and Peltier, W.R. 1988. A new global model of Late Pleistocene deglacition inferred from relative sea-level data. 17th Annual Arctic Workshop, Institute of Arctic and Alpine Research, Boulder, Colorado. Program with abstracts, 57-58.

APPENDIX I

RADIOCARBON DATES (NEW AND PREVIOUSLY PUBLISHED) FROM MARVIN PENINSULA, WARD HUNT ISLAND AND ICE SHELF

RADIOCARBON DATES FROM MARVIN PENINSULA, WARD HUNT ISLAND AND ICE SHELF

TOTAL NUMBER OF DATES = 83

RADIOCARBON DATES FROM MARVIN PENINSULA, WARD HUNT ISLAND AND ICE SHELF

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATIGRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
HOLOCENE MARINE SHELLS							
Ward Hunt Island							
Ward Hunt Island - N side (Crary 1960)	L248A	Marine shells	7200±200	beach, surface sample	38 m (?)		
Ward Hunt Island - N side (Lyons and Mielke, 1973)	SI-718	Marine shells, Hiatella	7755±150	beach, surface sample	38 m (38 m?)		
Ward Hunt Island - E side (Lyons and Mielke, 1973)	SI-720	Marine shells, Hiatella	5950±155	beach? surface sample?	5 m (?)		
North Coast (east to west)							
Camp Creek E side (Lyons and Mielke, 1973)	SI-725	Marine shells, Hiatella, Astarte, Vermetus	7045±190	??	30 m (?)		
Camp Creek E side (Lemmen, this study)	TO-490	Marine shells, Hiatella	8630±100	marine silts	46 m (<75 m)	83°01'	74°51'
Camp Creek Ice Rise - S margin (Lyons and Mielke, 1973)	SI-723	Marine shells, Hiatella	4775±120	surface sample, shells washing out beneath ice rise	5 m. (?)		

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)

HOLOCENE MARINE SHELLS (cont.)

Rambow Delta (Lyons and Mielke, 1973)	SI-724	Marine shells, Hiatella	5735±110	??	10 m (?)		
Sentinal River Valley (Lemmen, this study)	TO-487	Marine shells, Portlandia	9080±110	marine diamicton	37 m (?)	83°00'	75°39'
Sentinal River Valley (Lemmen, this study)	TO-488	Marine shells, Portlandia	9560±70	marine silts	39 m (?)	83°00'	75°39'
Paradise Lost - C.D. Ice Rise (Lemmen, this study)	TO-861	Marine shells, Hiatella	8630±70	silty-sand	42 m (67± m)	83°03'	76°08'

M'Clintock Inlet (north to south)

M'Clintock Inlet - W side, outer (Crary, 1960) (Christie, 1967)	L-248B	Marine shells	7200±250	near terminus of glacier	40± m		
M'Clintock Inlet - Ootah Bay (Lemmen, this study)	TO-265	Marine shells, Hiatella	7320±90	delta foresets	32 m (68 m?)	82°47'	75°54'
M'Clintock Inlet - Ootah Bay (Lemmen, this study)	TO-263	Marine shells, Hiatella	7800±90	delta foresets	46 m (81 m?)	82°46'	75°42'
M'Clintock Inlet - Central Valley (Lemmen, this study)	TO-499	Marine shells, Hiatella	8140±90	marine silt	73 m (<123 m)	82°39'	75°48'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
HOLOCENE MARINE SHELLS (cont.)							
M'Clintock Inlet - Central Valley (Lemmen, this study)	TO-267	Marine shells, Hiatella	8870±110	marine silt	84 m (<110 m)	82°38'	75°37'
M'Clintock Inlet - Central Valley (Lemmen, this study)	TO-262	Marine shells, Hiatella	7770±70	silty-sand	36 m (?)	82°38'	75°16'
<u>Outer Disraeli Fiord</u>							
Disraeli Fiord - Ice Shelf Creek (Lemmen, this study)	TO-862	Marine shells, Portlandia	9250±80	marine sand	62 m (78 m)	82°57'	74°02'
<u>Disraeli Fiord - Central Valley</u>							
Disraeli Fiord - S Central Valley (Blake, 1987)	GSC-1850	Marine shells, Mya truncata	8130±120	marine clay, surface sample	51 m	82°47'	73°56'
Disraeli Fiord - S Central Valley (Lemmen, this study)	TO-268	Marine shells, Hiatella arctica	5450±90	marine silts	7 m (?)	82°50'	73°42'
Disraeli Fiord - S Central Valley (Lemmen, this study)	TO-270	Marine shells, Portlandia	8860±60	marine silts over diamicton	46 m (85 - 97 m ?)	82°48'	73°51'
Disraeli Fiord - Central Valley (Lemmen, this study)	TO-269	Marine shells, Mya truncata	8150±60	marine silts, proximal bottomsets	45 m (92 m)	82°49'	73°54'
Disraeli Fiord - Central Valley (Lemmen, this study)	TO-264	Marine shells, Mya truncata	4920±70	marine silts	15 m (?)	82°50'	73°48'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
HOLOCENE MARINE SHELLS (cont.)							
Disraeli Fiord - Central Valley (Lemmen, this study)	TO-266	Marine shells, Mya truncata	4080±60	marine sandy-silt	3 m (?)	82°51'	73°46'
Inner Disraeli Fiord (north to south)							
Disraeli Fiord - W side, inner (Blake, 1987)	GSC-2637	Marine shells, Mya truncata	4390±60		<1 m	82°40'	72°36'
Disraeli Fiord - Collision Delta (Lemmen, this study)	TO-496	Marine shells, Hiatella	8150±80	deltaic sands	64 m (85 m)	82°39'	72°39'
Disraeli Fiord - Collision Delta (Tushingham, unpublished)	TO-519	Marine shells, Hiatella	7950±60	deltaic sands	64 m (85 m)	82°39'	72°39'
Disraeli Fiord - Collision Delta (Tushingham, unpublished)	TO-520	Marine shells, Hiatella	7800±60	deltaic sands	64 m (85 m)	82°39'	72°39'
Disraeli Fiord - Collision Delta (Tushingham, unpublished)	TO-521	Marine shells, Hiatella	10,090±70	deltaic sands	64 m (85 m)	82°39'	72°39'
Disraeli Fiord - Collision Delta (Tushingham, unpublished)	TO-522	Marine shells, Hiatella	6460±60	marine silts	23 m (?)	82°38'	72°35'
Disraeli Fiord - Thores River (Lemmen, this study)	TO-494	Marine shells, Hiatella	7260±80	marine silt	33 m (<68 m)	82°35'	72°44'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
HOLOCENE MARINE SHELLS (cont.)							
Disraeli Fiord - Thores River (Tushingham, unpublished)	TO-526	Marine shells, Hiatella	7110+60	marine silt	32 m (<<68 m)	82°35'	72°44'
Disraeli Fiord - Thores River (Lemmen, this study)	TO-491	Marine shells, Portlandia	8010±100	marine silts, disturbed	31 m (?)	82°36'	72°45'
Disraeli Fiord - Thores River (Tushingham, unpublished)	TO-523	Marine shells, Hiatella	4040±50	deltaic gravels	2 m (14 m?)	82°35'	72°42'
Disraeli Fiord - Thores River (Tushingham, unpublished)	TO-524	Marine shells, Hiatella	5680±90	marine silts	21 m (?)	82°35'	72°45'
Disraeli Fiord - Thores River (Tushingham, unpublished)	TO-525	Marine shells, Hiatella	6890±60	marine silts	6 m (?)	82°36'	72°42'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
PRE-HOLOCENE MARINE SHELLS							
<u>Ward Hunt Island</u>							
Ward Hunt Island - E side Walker Hill (Lemmen, this study)	TO-858	Shell fragments	29,790±200	colluvium	97 - 115 m (?)	83°05'	74°12'
Ward Hunt Island - S. side (Lemmen, this study)	TO-860	Shell fragments	27,560±220	silty-sand, surface sample	52 m (62± m)	83°04'	74°07'
<u>North coast</u>							
Sentinal River Valley (Lemmen, this study)	TO-486	Shell fragments	32,010±300	silty-sand, surface sample	100 m (>100 m)	83°02'	75°43'
Sentinel River Valley (Lemmen, this study)	TO-860	Marine shells, Astarte	27,170±210	silty-sand	117 m (117+ m)	83°02'	75°46'
<u>Disraeli Fiord</u>							
Disraeli Fiord - Central Valley (Lemmen, this study)	TO-500	Shell fragments	30,440±330	glaciomarine diamicton (grounding line deposit?)	81 m	82°50'	73°49'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
ORGANICS							
North coast							
Sentinal River Valley (Lemmen, this study)	TO-489	Subfossil bryophytes	23,340±430	marine sands	41 m (?)	83°00'	75°39'
M'Clintock Inlet							
M'Clintock Inlet - Central Valley (Lemmen, this study)	TO-498	Subfossil bryophytes	14,730±120	marine sand	98 m (98± m)	82°38'	75°38'
M'Clintock Inlet - Central Valley (Lemmen, this study)	TO-497	Subfossil bryophytes	15,710±180	marine sand	89 m (?)	82°38'	75°37'
Disraeli Fiord							
Disraeli Fiord - Ice Shelf Creek (Lemmen, this study)	TO-857	Organics, sandy peat	11,340±70	marine sand	62 m (78 m)	82°57'	74°02'
Disraeli Fiord - Thores River (Lemmen, this study)	TO-492	Subfossil bryophytes	31,360±400	lacustrine silty sand, ice-dammed lake	72 m (NR)	82°36'	54°??'
Disraeli Fiord - Thores River (Lemmen, this study)	TO-493	organics, sandy peat	7730±70	deltaic sands	39 m (68± m)	82°36'	72°45'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATIGRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
DRIFTWOOD							
North coast							
Paradise Lost - C.D. Ice Rise (Lemmen, this study)	TO-863	Driftwood	5800±50	beach berm	0.4 - 0.6 m	83°03'	75°56'
Paradise Lost - C.D. Ice Rise (Lemmen, this study)	TO-864	Driftwood	5730±60	beach berm	0.4 - 0.6 m	83°03'	75°56'
Paradise Lost - C.D. Ice Rise (Lemmen, this study)	GSC-4559	Driftwood *Picea*	8850±80	beach berm	0.4 - 0.6 m	83°03'	75°56'
Disraeli Fiord							
Disraeli Fiord - W side, mouth (Crary, 1960)	L254A	Driftwood	3400±150		<3 m	83°00'	74°13'
Disraeli Fiord - W side, mouth (Crary, 1960)	L254B	Driftwood	5740±200		<3 m	83°00'	74°13'
Disraeli Fiord - W side, mouth (Crary, 1960)	L254C	Driftwood	6120±150		<3 m	83°00'	74°13'
Disraeli Fiord - W side, mouth (Crary, 1960)	L254D	Driftwood	3000±200		<3 m	83°00'	74°13'
Disraeli Fiord - W side, mouth (Mielke and Long 1969)	SI-568	Driftwood *Larix*	6280±140			83°00'	74°13'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATIGRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
DRIFTWOOD (cont.)							
Disraeli Fiord (Mielke and Long 1969)	SI-566	Driftwood	>35,000				
Disraeli Fiord (Mielke and Long 1969)	SI-567	Driftwood	>35,000				
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2512	Driftwood *Picea*	4570±70		0.5 - 1.0 m	82°50'	73°15'
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2226	Driftwood *Larix*	4390±60		<.7 m	82°45'	72°45'
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2292	Driftwood *Picea*	3940±60		<.7 m	82°45'	72°45'
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2357	Driftwood *Picea*	3980±60		<.7 m	82°45'	72°45'
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2398	Driftwood *Picea*	3830±140		<.7 m	82°45'	72°45'
Disraeli Fiord - E side, central (Blake, 1987)	GSC-2733	Driftwood *Larix*	4550±60		0.5 - 1.0 m	82°50'	73°15'
Disraeli Fiord - E side, central (Blake, 1987)	GSG-2784	Driftwood *Larix*	4680±60		0.5 - 1.0 m	82°50'	73°15'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATIGRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
DRIFTWOOD (cont.)							
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1831	Driftwood <u>Picea</u>	> 42,000		0.3 - 0.5 m	82°50'	73°42'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1885	Driftwood <u>Picea</u>	> 42,000		0.3 - 0.5 m	82°50'	73°42'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1862	Driftwood <u>Picea</u>	> 40,000		0.3 - 0.5 m	82°50'	73°42'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-2154	Driftwood <u>Larix</u>	> 41,000		0.3 - 1.0 m	82°50'	73°45'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1794	Driftwood <u>Larix</u>	5930±60		23 m	82°49'	73°45'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-2047	Driftwood <u>Larix</u>	5300±60		8 m	82°49'	73°45'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1823	Driftwood <u>Picea</u>	6220±80		0.3 - 0.5 m	82°50'	73°42'
Disraeli Fiord - Central Valley (Blake, 1987)	GSC-1806	Driftwood <u>Picea</u>	5180±70		0.3 - 0.5 m	82°50'	73°42'
Disraeli Fiord - S Central Valley (Lemmen, this study)	TO-261	Driftwood	6510±70	surface of silts	19 m	82°49'	73°44'
Disraeli Fiord - Thores River (Lemmen, this study)	TO-495	Driftwood	5030±70	delta surface	27 m (?)	82°35'	72°44'

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATIGRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
WARD HUNT ICE SHELF							
Ice Shelf- W end of Ward Hunt Island (Broecker et al. 1956)	L-284	sponges	400±150	Ice Shelf surface			
Ice Shelf-SE corner Ward Hunt Island (Lowdon and Blake, 1970)	GSC-1025	Marine shells	4510±150	Ice Shelf	<1 m	83°05′	73°52′
Ice Shelf-SE corner Ward Hunt Island (Lyons and Mielke, 1973)	SI-719A	Carbonate, calcilutite material, siliceous sponges	13,200±440	Ice Shelf, ice-thrust structures		83°04′	74°05′
Ice Shelf-SE corner Ward Hunt Island (Lyons and Mielke, 1973)	SI-719B	Total organics in sponge debris	3400±140	Ice Shelf		83°04′	74°05′
Ice Shelf-SE corner Ward Hunt Island (Lyons and Mielke, 1973)	SI-638	Marine shells	3645±120	Ice Shelf basement ice		83°04′	74°05′
Ice Shelf-SE corner Ward Hunt Island (Lyons and Mielke, 1973)	SI-722	Marine shells	3390±130	Ice Shelf basement ice		83°04′	74°05′
Ice Shelf- 1.6 km NNE of NE tip of Ward Hunt Island (Lyons and Mielke, 1973)	SI-721	Marine shells, Astarte, Vermetus	6815±190	Ice Shelf junction basement ice with moat ice			

SITE/LOCATION REFERENCE	LAB NO.	MATERIAL	AGE	STRATI-GRAPHY	SAMPLE ELEV. (RELATED SEA LEVEL)	LAT. (°N)	LONG. (°W)
WARD HUNT ICE SHELF (cont.)							
Ice Shelf-Cape Alexandra (Lyons and Mielke, 1973)	SI-727	Marine shells, Hiatella, Astarte	3680±100	Ice Shelf basement ice			
LACUSTRINE							
Lake A (Stuckenrath and Mielke 1973)	SI-730	Saline bottom water	4590±150		(5 m±)	83°00'	75°20'
Lake A (Stuckenrath and Mielke 1973)	SI-730	Saline bottom water	4315±150		(5 m±)	83°00'	75°20'

APPENDIX II

SURFICIAL GEOLOGY MAP OF MARVIN PENINSULA AND WARD HUNT ISLAND

Surficial geology map

A map of the surficial geology of Marvin Peninsula and Ward Hunt Island has been compiled from air photographs and ground-truthing. This map (Fig. A-II.1) presents only a preliminary interpretation due to the limited extent of ground-truthing, particularly at sites located inland and at high elevations. Additional problems with the air photography were encountered owing to heavy shadows in mountainous areas as well as fog along some coastal areas.

Differentiation between the major map units (defined below) is often difficult. Distinction between bedrock, colluvium and till veneer can be arbitrary because their boundaries are often gradational. Given the mountainous nature of the study area slope processes tend to be the dominant geomorphic agents. These are enhanced by the presence of continuous permafrost, such that gelifluction is an efficient process on even gentle slopes. Confusion is encountered in distinguishing between weathered bedrock and glacial sediments when both have been reworked by slope processes.

In an effort to maintain objectivity in the mapping process the following criteria were used to distinguish between these three units. Till veneer was mapped in areas where there has been some obscuring of the geologic structure by surficial materials which are associated with unequivocal glacial features (moraines and/or meltwater channels). Colluvium was mapped on valley sides and floors in areas of gentle relief where unequivocal geomorphic evidence of glaciation is not present. Bedrock was mapped on steeper slopes even when a thin cover of weathered debris was present, unless a textural break unrelated to

Figure AII.1. Surficial geology map of Marvin Peninsula and Ward Hunt Island.

WARD HUNT ICE SHELF
CRANSTONE
PENINSULA
DISRAELI
FIORD
LEGEND
active alluvium
alluvium
till veneer
till blanket
glaciolacustrine
marine - fines
marine - deltaic
colluvium
residuum
bedrock
glaciers
moraine
abandoned channel
striae
kettle hole
rock glacier
beach
tor
SCALE
5 0 5 10
kilometres

lithological changes could be detected. In general the amount of colluvium indicated on the surficial geology map greatly under-represents the amount of such material that occurs in the field area. Conversely, the amount of till indicated on Fig. A-II.1 exceeds that which could sedimentologically be identified as till through ground-truthing.

Bedrock

Bedrock is the predominant surficial material in the field area and occurs as clean exposures in numerous ridges and cliffs. A thin mantle of felsenmeer may occur on steep to intermediate slopes. Sparse erratics may also be present.

Colluvium

Poorly sorted slope materials, primarily derived from weathered bedrock, are collectively mapped as colluvium. This includes: coarse angular talus deposited at the base of steep bedrock slopes; as well as weathered bedrock and sparse erratics on lower slopes and valley floors which have been transported by gelifluction and slope wash.

Residuum

Residuum refers to thick accumulations of felsenmeer formed <u>in situ</u> through the weathering of the underlying bedrock. It is found on upland plateaus and broad summits, and commonly occurs in association with tors. Sparse erratics may occur within the felsenmeer.

Till

Sediment which has been deposited directly by glaciers is mapped as till, although it has often been subject to reworking by nonglacial processes. This unit has been subdivided into:

till veneer; a thin, often discontinuous till cover less than ca. 2 m thick, and frequently <0.5 m thick. The ground surface mimics the underlying topography, and often does not obscure the bedrock structure;

and till blanket; a till cover which is greater than ca. 2 m in thickness. Till blankets occur only rarely in study area.

Alluvium

All fluvial and glaciofluvial sediments are mapped as alluvium. These sediments are composed primarily of bedded gravel and sand which have been deposited as sandurs and fans, as well as some small kame deposits. This unit has been subdivided into:

active alluvium; which includes all seasonally active depositional environments;

and inactive alluvium; fluvial and glaciofluvial sediments which occur above the contemporary floodplain. Inactive alluvium is most common in the form of terraces which are related to deposition into a sea level higher than at present.

Raised marine sediments

This unit includes all sediments which have been deposited into a marine environment, at or below sea level, and have subsequently emerged. Raised marine sediments have been subdivided into:

deltaic sediments: composed of coarse sands and gravels which form the foreset and topset beds of Gilbert-type deltas. At many sites these sediments are important for defining the local marine limit;

and marine fines: sand, silt and clay, commonly fossiliferous, which represents littoral or deep water sedimentation. Locally this unit may include beach deposits.

Glaciolacustrine sediments

All sediments deposited into former ice-dammed lakes are included in this unit. These sediments are of minor importance in the study area.

APPENDIX III

Bryophyte Species Present in Sample MIGB86-O1

Radiocarbon dated 15,710±180 BP (TO-497)

<u>Ditrichum flexicaule</u>

<u>Orthothecium chryseum</u>

<u>Scorpidium turgescens</u>

<u>Tomenthypnum nitens</u>

<u>Calliergon giganteum</u>

<u>Bryum pseudotriquetrum</u>

<u>Distichium</u> sp.

<u>Drepanocladus revolvens</u>

Identification by C. LaFarge-England, Dept. of Botany, University of Alberta.

APPENDIX IV

GRAIN-SIZE ANALYSIS DATA FOR DISRAELI FIORD SEDIMENT SAMPLES

DEPTH	5phi	16phi	25phi	50phi	75phi	84phi	95phi	MEAN	SD	SKEW.	KURT.	%SAND	%SILT	%CLAY
H7C1														
0.00	5.8	6.6	7.1	7.6	7.9	8.1	10.3	7.46	1.04	-0.11	2.11	0.00	81.00	19.00
5.00	5.9	6.7	7.2	8.4	9.8	11.2	13.2	8.75	2.20	0.28	1.12	0.00	43.00	57.00
10.00	6.0	6.7	7.1	8.2	9.7	10.9	13.0	8.59	2.12	0.35	1.11	0.00	47.00	53.00
13.00	6.3	6.9	7.3	8.3	9.6	10.7	13.0	8.64	1.98	0.32	1.23	0.00	42.00	58.00
15.00	6.1	6.7	7.2	8.3	9.7	10.9	13.0	8.64	2.10	0.31	1.12	0.00	45.00	55.00
20.00	6.0	6.7	7.1	8.3	9.7	10.8	13.0	8.57	2.08	0.28	1.12	0.00	45.00	55.00
25.00	5.6	6.3	6.7	7.8	9.0	9.8	12.5	7.98	1.92	0.27	1.23	0.00	55.00	45.00
30.00	5.9	6.7	7.2	8.3	9.6	10.4	12.8	8.47	1.98	0.23	1.16	0.00	44.00	56.00
33.00	5.6	6.4	6.9	8.1	9.0	9.3	11.0	7.94	1.53	-0.03	1.08	0.00	48.00	52.00
35.00	5.6	6.6	7.2	8.4	10.0	11.4	13.2	8.79	2.33	0.24	1.10	0.00	42.00	58.00
40.00	6.1	6.9	7.4	8.6	10.3	11.6	13.3	9.04	2.28	0.27	1.00	1.00	37.00	62.00
45.00	5.6	6.5	7.2	8.4	10.2	11.5	13.2	8.81	2.38	0.24	1.04	2.00	39.00	59.00
50.00	3.8	6.8	7.5	8.8	10.4	11.7	13.3	9.11	2.67	0.05	1.36	12.00	21.00	67.00
55.00	3.1	3.8	4.6	8.0	9.8	11.3	13.2	7.66	3.39	-0.04	0.79	24.00	26.00	50.00
65.00	3.2	3.6	3.8	7.3	9.3	10.8	13.0	7.20	3.29	0.07	0.73	34.00	24.00	42.00
H6C1														
0.00	6.2	7.4	8.1	9.3	11.2	12.2	13.4	9.61	2.29	0.17	0.96	0.00	22.00	76.00
5.00	6.4	7.3	7.8	9.0	10.4	11.7	13.3	9.31	2.16	0.24	1.06	0.00	30.00	70.00
10.00	6.6	7.3	7.7	8.6	9.6	10.5	12.9	8.79	1.75	0.28	1.36	0.00	33.00	67.00
15.00	6.2	7.1	7.6	8.9	10.4	11.7	13.3	9.23	2.23	0.23	1.02	0.00	33.00	67.00
20.00	6.6	7.2	7.6	8.9	10.6	11.8	13.3	9.31	2.16	0.29	0.94	0.00	32.00	68.00
25.00	6.5	7.3	7.7	8.7	9.8	11.3	13.2	9.07	2.00	0.31	1.26	0.00	32.00	68.00
30.00	6.5	7.2	7.5	8.4	9.4	10.0	12.7	8.51	1.64	0.27	1.35	0.00	32.00	68.00
35.00	6.5	7.1	7.4	8.3	9.3	9.8	12.5	8.42	1.58	0.28	1.31	0.00	42.00	58.00
H6C2														
0.00	5.7	6.4	6.7	7.3	7.8	8.0	9.0	7.25	0.90	0.02	1.32	0.00	82.00	17.00
10.00	6.2	6.9	7.4	8.4	9.5	10.0	12.6	8.42	1.74	0.15	1.24	0.00	40.00	60.00
20.00	6.0	6.8	7.3	8.4	9.6	10.6	12.9	8.58	2.01	0.23	1.24	0.00	41.00	59.00
30.00	6.4	7.1	7.5	8.6	9.8	11.1	13.0	8.93	2.01	0.27	1.19	0.00	36.00	64.00
40.00	6.7	7.5	7.9	9.0	9.8	11.1	13.0	9.17	1.85	0.22	1.32	0.00	27.00	73.00
50.00	6.7	7.4	7.9	8.8	9.8	11.0	13.0	9.08	1.85	0.27	1.31	0.00	29.00	71.00
60.00	6.2	7.0	7.4	8.6	9.8	11.0	13.0	8.90	2.03	0.24	1.20	0.00	36.00	64.00
70.00	6.2	6.9	7.4	8.4	9.6	10.4	12.8	8.58	1.86	0.26	1.21	0.00	41.00	59.00
80.00	6.3	7.2	7.6	8.5	9.6	10.2	12.7	8.63	1.73	0.20	1.33	0.00	35.00	65.00
90.00	6.1	7.0	7.6	8.8	9.8	11.1	13.0	8.96	2.08	0.19	1.25	0.00	32.00	68.00
100.00	6.3	7.0	7.4	8.4	9.6	10.3	12.8	8.58	1.80	0.24	1.20	0.00	40.00	60.00
110.00	6.8	7.5	7.9	8.9	10.1	11.5	13.2	9.29	1.95	0.33	1.19	0.00	28.00	72.00
120.00	6.9	7.5	7.9	9.0	10.3	11.5	13.2	9.33	1.95	0.30	1.09	0.00	27.00	73.00
130.00	6.2	6.5	6.9	7.8	8.7	9.2	9.8	7.82	1.22	0.10	0.83	1.00	57.00	42.00
140.00	7.3	8.0	8.3	9.3	11.0	11.9	13.3	9.70	1.88	0.32	0.91	0.00	18.00	82.00
150.00	6.3	6.6	7.2	8.5	10.2	11.4	13.2	8.85	2.24	0.30	0.94	1.00	41.00	58.00
160.00	5.2	6.0	6.4	7.8	9.6	10.7	12.9	8.17	2.35	0.27	1.00	1.00	53.00	46.00
170.00	6.1	6.8	7.3	8.5	9.8	11.2	13.0	8.82	2.14	0.27	1.11	1.00	40.00	59.00
180.00	6.1	6.8	7.3	8.5	9.8	11.2	13.2	8.84	2.17	0.26	1.16	1.00	38.00	61.00
190.00	5.8	6.6	7.1	8.2	9.7	11.0	13.0	8.60	2.19	0.29	1.12	1.00	45.00	54.00
200.00	5.1	6.0	6.6	8.1	9.6	10.9	13.0	8.34	2.44	0.20	1.09	2.00	46.00	52.00
210.00	5.2	6.1	6.8	8.0	9.6	10.8	12.9	8.28	2.34	0.24	1.05	2.00	50.00	48.00
220.00	6.0	7.1	7.5	8.6	10.1	11.4	13.2	9.02	2.16	0.26	1.11	2.00	34.00	64.00
230.00	5.8	6.7	7.3	8.9	10.7	11.9	13.3	9.17	[illegible]	0.17	0.91	2.00	33.00	65.00
240.00	5.3	6.5	7.1	8.5	10.1	11.5	13.2	8.84	2.45	0.18	1.06	3.00	37.00	60.00
250.00	6.1	7.2	7.8	8.9	10.5	11.8	13.3	9.31	2.23	0.25	1.08	4.00	27.00	69.00
H4C1														
0.00	3.6	5.1	5.6	7.0	8.6	9.6	12.5	7.22	2.49	[illegible]	1.21	11.00	54.00	35.00
H4C2														
0.00	3.1	3.8	4.6	7.4	8.9	9.7	12.6	6.95	2.92	-0.06	[illegible]	22.00	41.00	37.00
5.00	6.3	6.9	7.3	8.3	9.5	10.3	12.8	8.52	1.84	0.27	1.2[illegible]	[illegible]	41.00	59.00
H5C1														
0.00	6.4	7.0	7.4	8.2	9.0	9.5	12.1	8.24	1.48	0.24	1.37	0.00	[illegible]	[illegible]
5.00	6.3	7.1	7.5	8.6	9.7	10.9	12.9	8.83	1.94	0.27	1.23	0.00	36.00	64.00
H6E1														
0.00	6.7	7.2	7.4	7.8	8.1	8.4	9.6	7.80	0.75	0.08	1.72	0.00	67.00	33.00
H6E2														
0.00	6.2	6.8	7.2	8.0	9.2	9.8	12.3	8.19	1.68	0.27	1.23	0.00	49.00	51.00