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LATE QUATERNARY GLACIAL AND SEA LEVEL HISTORY OF WEST-CENTRAL AXEL HEIBERG ISLAND, HIGH ARCTIC CANADA.

by

CATRIONA ANNE MORRISON ©

A thesis submitted to the Faculty of Graduate Studies in partial fulfillment of the requirements for the degree of Master of Science.

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ABSTRACT

This thesis reconstructs the glacial and sea level history of west-central Axel Heiberg Island based upon the interpretation of surficial geology, geomorphology and raised marine deposits. Radiocarbon dates on marine fauna provide a preliminary chronology for ice advance and deglaciation.

The last glaciation was characterised by the northward and westward expansion of the Steacie Ice Cap. Ice inundated the mountains of west-central Axel Heiberg Island and extended an unknown distance offshore. An AMS date of 46.8 ka BP on an individual valve of *Hiatella arctica* from a fossiliferous till may provide a maximum age estimate for ice advance. Deglaciation was underway by 8.8 ka BP and involved the thinning of fiord-based trunk glaciers and the concomitant retreat of terrestrial margins towards the highland interior.

The pattern of differential postglacial emergence is indicated by the 8.5 ka BP isobases across west-central Axel Heiberg Island which reach ~115m asl over the study area. The revised isobases show a NE-SW orientation which is in alignment with a previously documented uplift centre of the Innuitian Ice Sheet to the northwest of Devon Island to the south.

ACKNOWLEDGEMENTS

Financial support for this research was provided by the Canadian Circumpolar Institute (BAR grants) and the Natural Sciences and Engineering Research Council of Canada (Grant A6608 to J. England). Logistical support in the field was provided by the Polar Continental Shelf Project and dedicated and tremendously helpful field assistance was provided by Roy Coultard. Many thanks also to Art Dyke for his guidance in the field.

This thesis has possibly made me sweat more than my regular trips to the gym. However, it has been a rewarding and truly memorable experience. Thank you to my supervisor John for taking me on as a graduate student and giving me the opportunity to work in the beautiful High Arctic. Thank you for the discussion, encouragement and hospitality during the past two years. John deserves special credit for dealing with my endless demands to "just read this". Although, in return, I have to say that he gave exceptionally *detailed* reviews! Thanks also to Martin Sharp and Nigel Atkinson for answering all those "little" questions that clarified so many "large" details. Special thanks also to Dr. Dave Evans (University of Glasgow) for encouraging me to take the path towards graduate studies.

My time at the University of Alberta has been greatly enriched by the many friends that I have made. Thanks to Nigel, Luke and Trudi for the movies and BBQ's. Sarah and Karen, we hobble along in science together! A large part of my life is dedicated to sport and it has been a privilege to train and compete at the Universtiy of Alberta, but most importantly it has been wonderful to meet so many like-minded and inspirational individuals. Kyle, Bill, Ann, Neila and Karen, thank you for all those important stress relieving runs, the socialising and most importantly your friendship.

Finally, thank you to my parents for their love, support and encouragement over the past two years. And thank you Richard for encouraging me to follow my heart. You are with me always no matter how far apart we may be.

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CHAPTER 1: INTRODUCTION AND RATIONALE

1.1 INTRODUCTION

This thesis concerns the late Quaternary glaciation and Holocene sea level history of part of the Queen Elizabeth Islands (QEI), High Arctic Canada (Figure 1.1; tables, plates and figures are located at the end of each chapter). For the past three decades, the style and extent of ice throughout the QEI during the last glacial maximum (LGM) has been debated. Two hypotheses have dominated the literature. Blake (1970, 1972) proposed that the QEI were covered by a pervasive Innuitian Ice Sheet (IIS) that coalesced with the Laurentide Ice Sheet to the south and the Greenland Ice Sheet to the east. In contrast, England (1976) proposed that an array of discontinuous alpine ice caps, the Franklin Ice Complex, occupied the QEI during the LGM. Recent research in the High Arctic has reached a concensus, concluding that the QEI were inundated by the IIS during the LGM (cf. Bednarski 1998; England 1998a, 1998b; Dyke 1998, 1999; Ó Cofaigh 1999).

The fieldwork presented in this thesis examines whether this consensus is applicable to west-central Axel Heiberg Island (Figure 1.1). Other than research in Expedition and Strand fiords (Lemmen *et al.*, 1991, 1994; Figure 1.1), the majority of western Axel Heiberg Island has remained uninvestigated. This area constitutes an important extension of Quaternary research in the alpine sector of the QEI that has previously focussed on Ellesmere and Devon islands.

In this chapter, the characteristics of the study area are presented, followed by an overview of previous regional and local research. The chapter concludes with a statement regarding the objectives of the research.

1.2 GEOLOGY

The oldest bedrock on Axel Heiberg Island is contained within the Franklinian mobile belt (Trettin 1991). It outcrops only on the extreme northwest tip of the island and is composed of volcanic rocks, shallow marine carbonates and deep water sediments, deposited from the lower Cambrian to upper Silurian (Trettin 1991). Superimposed on the Franklinian mobile belt are concordant marine and non-marine sedimentary rocks of the Sverdrup Basin succession, deposited from the lower Carboniferous to late Cretaceous. Bedrock of the Sverdrup Basin was deformed into shallow folds during the late Cretaceous to early Tertiary (Hodgson 1982) and experienced folding and thrust-faulting during the mid Tertiary Eurekan Orogeny (Thorsteinsson & Tozer 1976). Along the outer west coast of Axel Heiberg Island the bedrock is composed predominantly of shale, siltstone, ironstone and sandstone deposited from the Triassic to mid Tertiary (Figures 1.2, 1.3). It is cross-cut by igneous dykes as well as anhydrite and gypsum domes (Thorsteinsson 1970). During the late Tertiary, fluvial sediments ascribed to the Beaufort Formation (Fm) were deposited across the islands bordering the Arctic Ocean (Fyles 1990). Isolated patches of the Beaufort Fm have been identified on eastern Axel Heiberg and western Ellesmere islands (Thorsteinsson & Tozer 1976; Bustin 1982).

1.3 PHYSIOGRAPHY

The study area occurs within the Axel Heiberg-Ellesmere Island geomorphic region (Dawes & Christie 1991). This is dominated by mountains constructed during the mid Tertiary Eurekan Orogeny (Thorsteinsson & Tozer 1976). Deformation created a regionally pervasive north-northwest/south-southeast structural trend. Ice-covered highlands within the study area rise to ~900m asl. Fiords and glacially modified valleys, both accordant and discordant to the regional structure, radiate from these highlands (Dawes & Christie 1991). Thirty percent of Axel Heiberg Island is ice-covered, predominately by the Müller and Steacie ice caps (Figure 1.1).

Three field sites were visited on west-central Axel Heiberg Island. They are discussed in the order of site occupation and extend from north to south: Duck Bay, Mosquito Creek (unofficial name) and Good Friday Bay (Figure 1.4). The Duck Bay field area (79° 07' N, 92° 45' W) occupies ~10km^2 and is dominated by a glacial trough that trends northward from the Steacie Ice Cap, approximately 20km to the southeast (Figure 1.5). Highlands reach ~650m asl and the glacial trough is occupied by the braided, glacially-fed, Amorak River. To the west, surrounding Duck Bay, the topography is dominated by steep, and occasionally cliffed slopes that reach ~150m asl. The Mosquito Creek field site (79° 02' N, 93° 45' W) occupies ~10km^2 and is characterised by a lowland (<150m asl) bordered by small hills (~300m asl). The Good Friday Bay field area (78° 40' N, 92° 45' W) is located on the outer northwest coast of the fiord of the same name. The fiord extends westward from the Steacie Ice Cap to the Arctic Ocean and is ~10km wide and 36km long. The field area investigated covers an area of ~25km^2. It is dominated by a coastal lowland composed of raised marine sediments, bedrock ridges as well as an active sandur and adjacent inactive fluvial terraces. The elevation of the lowland is generally less than 60m asl. Inland the elevation rises to ~360m asl and the topography is characterised by low plateaux dissected by fluvial valleys.

1.4 CLIMATE

The climate of west-central Axel Heiberg is classified as polar desert, dominated by cold, dry, anticyclonic air from the Arctic Ocean (Maxwell 1981). The mean temperature of the region in January is −35°C whereas the July mean temperature is +5°C (Maxwell 1981). Annual precipitation ranges from 100 to 125mm with 35-40% falling as rain (Maxwell 1981). Regional precipitation is controlled by the northeastward movement of low-pressure systems from the Beaufort Sea (Aitken & Bell 1998). This is augmented by local orographic precipitation occasioned by the mountains of Axel Heiberg Island (Edlund & Alt 1988). On the west coast of Axel Heiberg Island the equilibrium line altitude is 400-600m asl and the regional glaciation level is 600-700m asl (Miller *et al.* 1975). In contrast, within the intermontane basin of Eureka Sound which lies within the rain shadow of Axel Heiberg Island, the glaciation level and equilibrium line altitude are 1100 and 900m asl respectively (Miller *et al.* 1975).

1.5 PREVIOUS RESEARCH ON FORMER ICE EXTENT IN THE QUEEN ELIZABETH ISLANDS.

1.5.1 Postglacial emergence

Historically, the magnitude of postglacial emergence has been widely used as a relative indicator of former ice thickness, and therefore an indirect measure of ice extent in the QEI (cf. Dyke 1999). Much of the discussion regarding the configuration of the IIS has centred on the interpretation of postglacial emergence patterns observed over the QEI. Blake (1970, 1975) identified an axis of greatest postglacial emergence (>25m since 5 ka BP; all dates are in radiocarbon years before present unless otherwise indicated) extending from Bathurst Island in the southwest to Eureka Sound in the northeast (Figures 1.1, 1.6). This axis of emergence was termed the Innuitian uplift (Walcott 1972) and was interpreted to record the axis of former maximum ice thickness of the IIS (Blake 1970; Tushingham 1991).

Recent work has shown that the Innuitian uplift originally proposed by Blake (1970, 1975) coincides with the occupation of Eureka Sound by ice flowing from the encircling mountains of Axel Heiberg and Ellesmere islands during the LGM (England 1998b; England & Ó Cofaigh 1998; Ó Cofaigh 1999; 1, Figure 1.7). From Eureka Sound, ice diverged northwards into Nansen Sound and southward into Norwegian Bay (cf. Bednarski 1998; Ó Cofaigh 1999; 2,3 Figure 1.7). The elevation of marine limit in Nansen Sound rises southward towards the entrance of Eureka Sound indicating greatest postglacial emergence and hence greatest change in ice thickness there (Bednarski 1998). Eureka Sound constituted the centre of glacial loading of the alpine sector of the IIS (England & Ó Cofaigh 1998; Ó Cofaigh 1999) and it has consequently experienced the greatest unloading and therefore the greatest amount of glacioisostatic rebound following deglaciation. This is indicated by isobases drawn on the 8.5 ka BP shoreline that form a closed cell over the sound (England & Ó Cofaigh 1998; Ó Cofaigh 1999; Figure 1.6). The ridge of postglacial emergence over Eureka Sound extends westward to a similar ridge within postglacial isobases over the central Arctic that Dyke (1998) attributes to a former loading centre that trended northeast to southwest over the QEI located to the northwest of Devon Island (4, Figure 1.7).

1.5.2 Geological and geomorphological evidence.

Original glacial-geological evidence in support of the IIS was presented by Blake (1977, 1992a) who attributed fresh glacial abrasion of granite in the vicinity of Smith Sound, southern Ellesmere Island (Figure 1.1), to the passage of regionally extensive ice during the LGM. More recently, westward ice flow across Ellesmere Island was recorded by Bell (1992) who identified a granite erratic train extending from the Shield of southeast Ellesmere Island to western Fosheim Peninsula and the east coast of Axel Heiberg Island via the axis of Bay Fiord (5, Figure 1.7). Ó Cofaigh (1999) recognised a second granite erratic train extending from the Prince of Wales Icefield westward through Baumann Fiord. Atkinson (1999) clarified this westward flow through Baumann Fiord showing that it was deflected southwestward across Bjorne Peninsula en route to Norwegian Bay (6, Figure 1.7). Bednarski (1998) recorded glacial flutings, striae and till fabric orientations bordering Nansen Sound that document ice flowing northwards from Eureka Sound (2, Figure 1.7). On Devon Island, striae, ice moulded bedrock and fluted till record coastward ice flow from a divide situated over the central axis of the island (Dyke 1999; 7, Figure 1.7). In Wellington Channel bordering the west coast of Devon Island, southeastward ice flow (8, Figure 1.7). is recorded by landforms of glacial erosion and deposition on Beechy and Ballie Hamilton islands (Figure 1.1). These include ice moulded and striated bedrock and till streaks (Dyke 1999; Hättestrand & Stroeven 1996).

1.5.3 Chronology of ice buildup and deglaciation.

The chronology of ice buildup is recorded by radiocarbon dates from shelly till deposited throughout the QEI. Along Nares Strait (Figure 1.1), radiocarbon dates indicate that ice buildup occurred as late as 19 ka BP (England 1998a, 1999). Sub-till organics of a similar age have been reported on southern Ellesmere Island (Blake 1992b). To the north in Nansen Sound, radiocarbon dates of 29.6 ka BP and 28.8 ka BP provide maximum ages for initial ice advance (Bednarski 1998). To the south, on Devon Island, a radiocarbon date from shelly till indicates that ice had advanced to its LGM limit by 23 ka BP Dyke (1998).

Blake (1972) reported that radiocarbon dates on marine molluscs (~11 ka BP) collected on raised marine shorelines along the western extremity of the QEI at Prince Patrick Island became progressively younger towards Eureka Sound (~8-9 ka BP). This age distribution was proposed to record the entry of marine fauna eastward across the

QEI occasioned by the retreating margin of the IIS. Recent field studies support this observation. In Nares Strait, radiocarbon dates on molluscs associated with marine limit indicate re-entry of the sea from the northern and southern ends of the strait at 10 and 9 ka BP respectively (England 1999). In addition, ^{36}Cl dates from islands within Nares Strait show that it was ice-filled until 10 ka BP (Zreda *et al.* 1999). The deglaciation of northern Nansen sound occurred by 10.3 ka BP (Bednarski 1998) and the deglaciation of Fosheim Peninsula between 9.2 and 7.9 ka BP (Bell 1996). Ó Cofaigh (1999) reported that southern Eureka Sound was deglaciated prior to 9.2 ka BP. Deglaciation was proposed to have been facilitated by rising eustatic sea levels during the early Holocene (Fairbanks 1989) coupled to rising temperatures, particularly between 8.5-9.5 k cal BP (Koerner & Fisher 1990; England 1992; Ó Cofaigh 1999). A two-step pattern characterised ice retreat: A scarcity of glacial landforms and sediments in outer fiord areas was attributed to rapid retreat of calving ice fronts whilst, prominent deglacial landform-sediment assemblages at fiord heads were attributed to the on-land stabilisation of ice margins (Hodgson 1985; Ó Cofaigh 1998, 1999; Ó Cofaigh & England 1998; Ó Cofaigh *et al.* 1998).

1.5.4 Previous Research on west-central Axel Heiberg Island.

The resolution of the late Quaternary history of most of Ellesmere Island and the central Arctic has not had its counterpart on western Axel Heiberg Island. Hein *et al.* (1990) and Hein and Mudie (1991) extracted sedimentary cores from the shelf off northwest Axel Heiberg Island. They interpreted the sediments overlying the Tertiary bedrock of the shelf as marine in origin and concluded that grounded ice had never occupied the investigated part of the shelf. Research documenting terrestrial glacial history has focussed on the Expedition Fiord area of west-central Axel Heiberg Island (Figure 1.8). Boesch (1963) identified two weathering zones within the fiord, a lower zone of well-preserved glacial landforms and an upper zone of more weathered landforms and erratics. The lower zone was attributed to the last glaciation whereas the upper zone was regarded as older but of an unknown absolute age. Müller (1963) recorded a marine limit of 80m asl at the head of Expedition Fiord dating to 9.0 ka BP, indicating deglaciation prior to this time.

More recently, Lemmen *et al.* (1991, 1994) investigated the glacial history of Expedition Fiord and inner Strand Fiord (Figure 1.8). In Expedition Fiord, discontinuous moraines are located along the north coast from the fiord head to Middle Islands (~20km from the fiord head; Figure 1.8). However, it was unknown whether the moraines marked the maximum extent of ice during the LGM or whether they marked ice-marginal stabilisation during retreat. Radiocarbon dates indicate that deglaciation of the inner fiord was underway by 8.4 ka BP and proceeded rapidly, facilitated by calving of the trunk glacier's marine margin. By 8.2-8.3 ka BP ice had attained a stable position upvalley of the present terminus of the Thompson Glacier. Lemmen *et al.* (1994) found no evidence to suggest that a trunk glacier had occupied Strand Fiord during the LGM. Rather, meltwater channels record the position of tributary glaciers that terminated at tidewater along the south coast of the fiord. Holocene marine limit was reported to be 124m asl at the head of Strand Fiord and 105m asl towards its centre (Lemmen *et al.* 1994). Radiocarbon dates on marine molluscs contained within raised marine deltas at the head of Strand Fiord indicated that deglaciation had occurred there by 8.4 ka BP (Lemmen *et al.* 1994).

1.6 RATIONALE

The debate concerning the style of glaciation in the QEI during the LGM has now been resolved in several sectors where consensus favours the presence of the IIS. However, a comprehensive overview of the maximum limit, ice surface geometry, flow pattern and chronology of the IIS has yet to be established. The area where the least information exists on the IIS remains the western Arctic islands. The coast of west-central Axel Heiberg Island constitutes one of the uninvestigated sectors of the IIS and warrants study as part of the ongoing regional examination of late Quaternary glacial and sea level history of the QEI.

1.7 RESEARCH OBJECTIVES

This thesis addresses two principal objectives aimed to enhance our knowledge of the late Quaternary glacial and sea level history of west-central Axel Heiberg Island.

1) To determine the configuration, origin and chronology of former glaciation on west-central Axel Heiberg Island

West-central Axel Heiberg Island constitutes the transition from the well-studied alpine sector of the QEI to the open topography of the western islands of the archipelago. Previous regional research has suggested drainage of Innuitian ice into Eureka Sound and subsequent flow westward into Norwegian Bay (England 1998b; Ó Cofaigh 1999). An ice ridge trended westward from Norwegian Bay into the central Arctic islands, from which an ice stream extended southeastward through Wellington Channel (Dyke 1998, 1999; 4,8 Figure 1.7). An implication of this reconstruction is that an opposing northwestward flow may have occupied Massey Sound and interacted with local alpine glaciation on west-central Axel Heiberg Island (J. England pers. comm.1999; 9, Figure 1.7). Previous local research is restricted to the north of the study area where Lemmen *et al.* (1994) concluded that during the LGM ice occupied the tributary valleys that drain northwards into Strand Fiord. The origin and configuration of glaciation on west-central Axel Heiberg Island will be determined by mapping glacial sediments and landforms. This will enable the relative contribution of local and regional ice sources to be determined. At present, radiocarbon dates on ice-buildup are unavailable along western Axel Heiberg Island and dating organic materials contained within glacial sediments will provide a maximum age for ice-buildup in this area.

2) To determine the pattern and chronology of deglaciation and the nature of relative sea level adjustments on west-central Axel Heiberg Island.

Marine transgression along formerly glaciated coastlines is recorded by marine sediments deposited at the margins of retreating ice masses that have formerly isostatically loaded the crust. Subsequent isostatic rebound of the crust in response to glacial unloading results in postglacial emergence of the land surface and its associated sediments. Mapping the distribution of raised marine sediments and landforms and surveying their elevations will document the pattern of postglacial emergence occasioned by deglaciation. Radiocarbon dating of marine fauna located within these sediments will provide a chronological framework of deglaciation. The magnitude of postglacial emergence can be used as a proxy to determine former ice thickness based on the assumption that the amount of postglacial emergence is proportional to the original ice load thickness (Andrews 1970, Tushingham 1991) versus non-glacial controls such as tectonics (cf. England 1987, 1997). Delineating spatial and temporal variations in postglacial emergence over west-central Axel Heiberg Island will document the pattern of differential postglacial emergence. This research augments the regional study of sea level adjustments in the QEI by providing data from the little studied coast of western Axel Heiberg Island.

FIGURE 1.1: QUEEN ELIZABETH ISLANDS

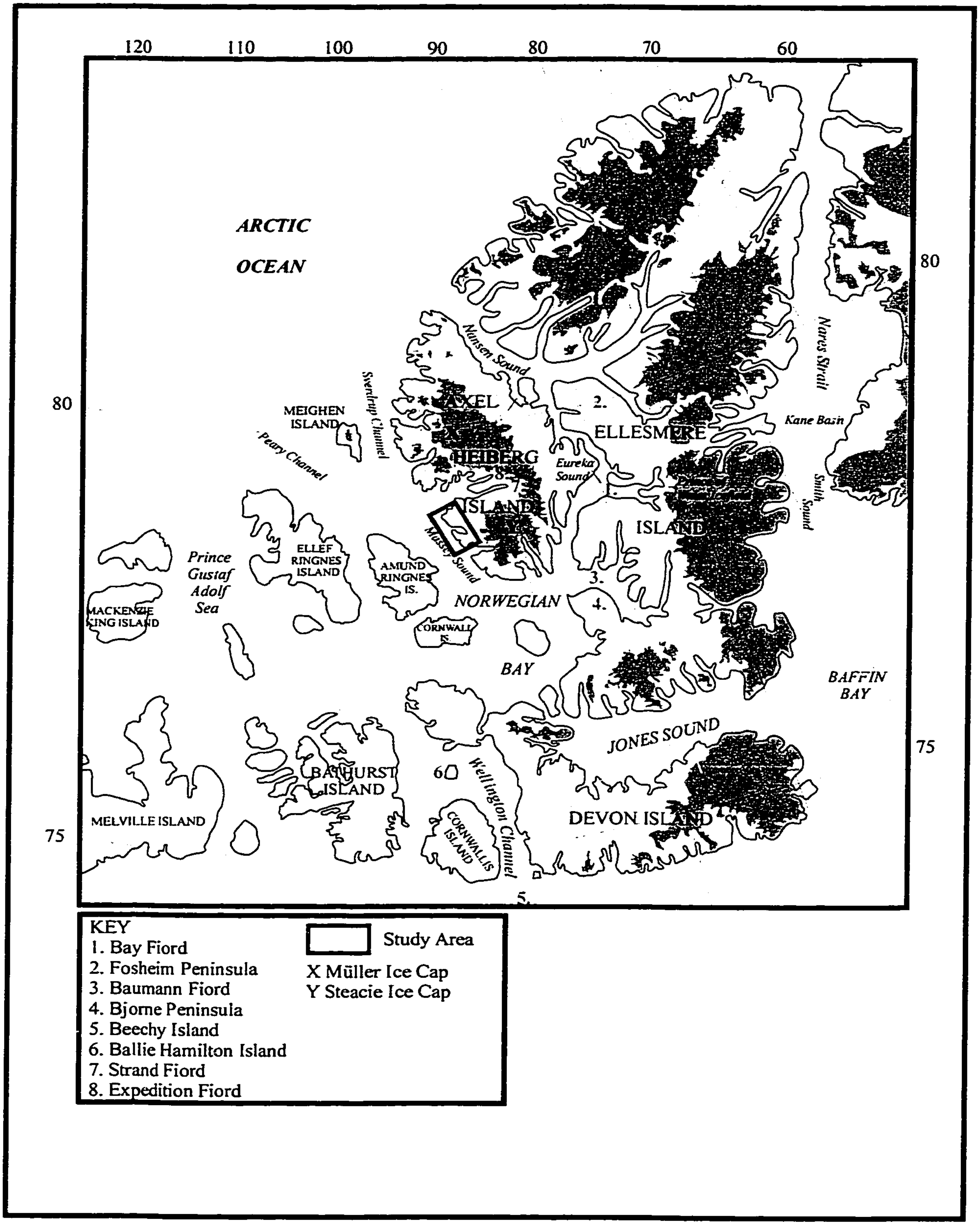
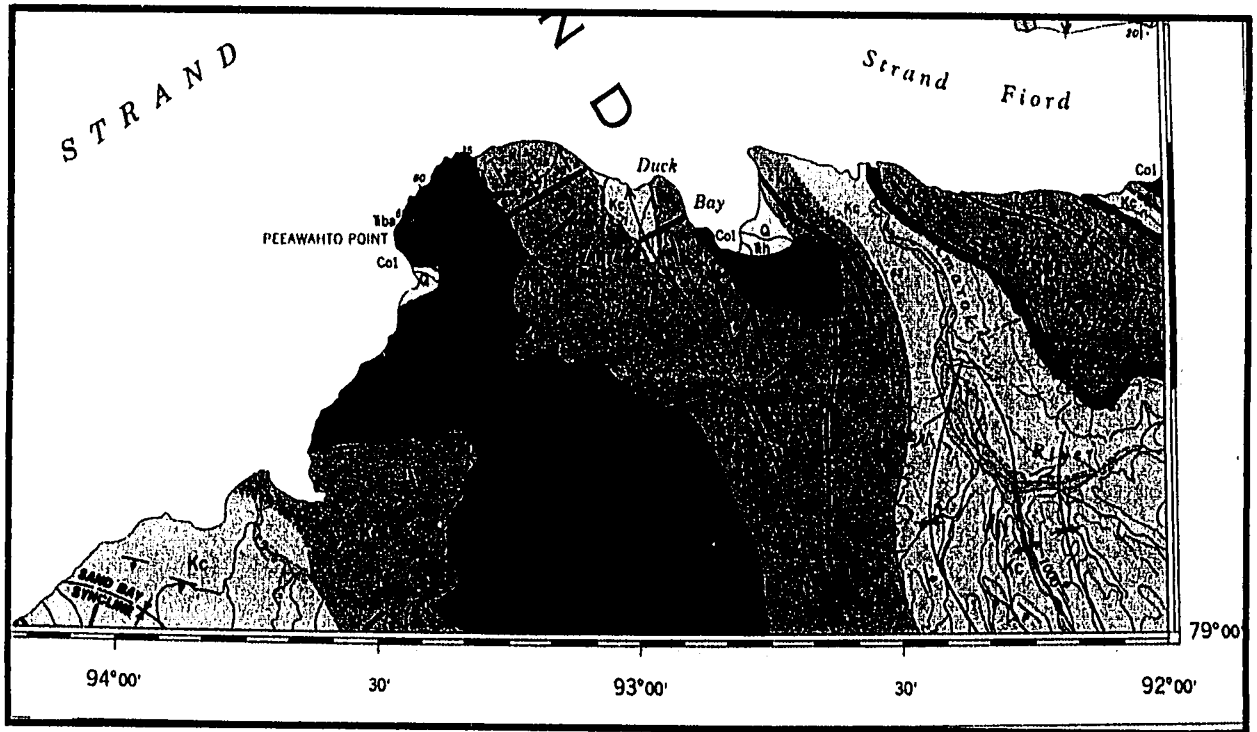


FIGURE 1.2: BEDROCK GEOLOGY, WEST-CENTRAL AXEL HEIBERG ISLAND (Thorsteinsson 1970)



KEY

- Q — STREAM, DELTAIC, GLACIAL AND MARINE (Quaternary)
- Kk — KANGUCK FORMATION (upper Cretaceous)
- HASSEL FORMATION (upper Cretaceous)
- Kc — CHRISTOPHER FORMATION (lower Cretaceous)
- Ki — ISACHSEN FORMATION (lower Cretaceous)
- DEER BAY FORMATION (upper Jurassic)
- AWINGAK FORMATION (upper Jurassic)
- SAVIK FORMATION (upperJurassic)
- OTTO FIORD FORMATION (Tertiary)
- // GABBRO, DIABASE AND BASALT DYKES (Cretaceous)

FIGURE 1.3: BEDROCK GEOLOGY, WEST-CENTRAL AXEL HEIBERG ISLAND
(Thorsteinsson 1970)

KEY

Q STREAM, DELTAIC, GLACIAL AND MARINE (Quaternary)

KK KANGUCK FORMATION (upper Cretaceous)

HASSEL FORMATION (upper Cretaceous)

Kc CHRISTOPHER FORMATION (lower Cretaceous)

KI ISACHSEN FORMATION (lower Cretaceous)

DEER BAY FORMATION (upper Jurassic)

AWINGAK FORMATION (upper Jurassic)

OTTO FIORD FORMATION (Tertiary)

// GABBRO, DIABASE AND BASALT DYKES (Cretaceous)

FIGURE 1.4: FIELD AREAS ON WEST-CENTRAL AXEL HEIBERG ISLAND

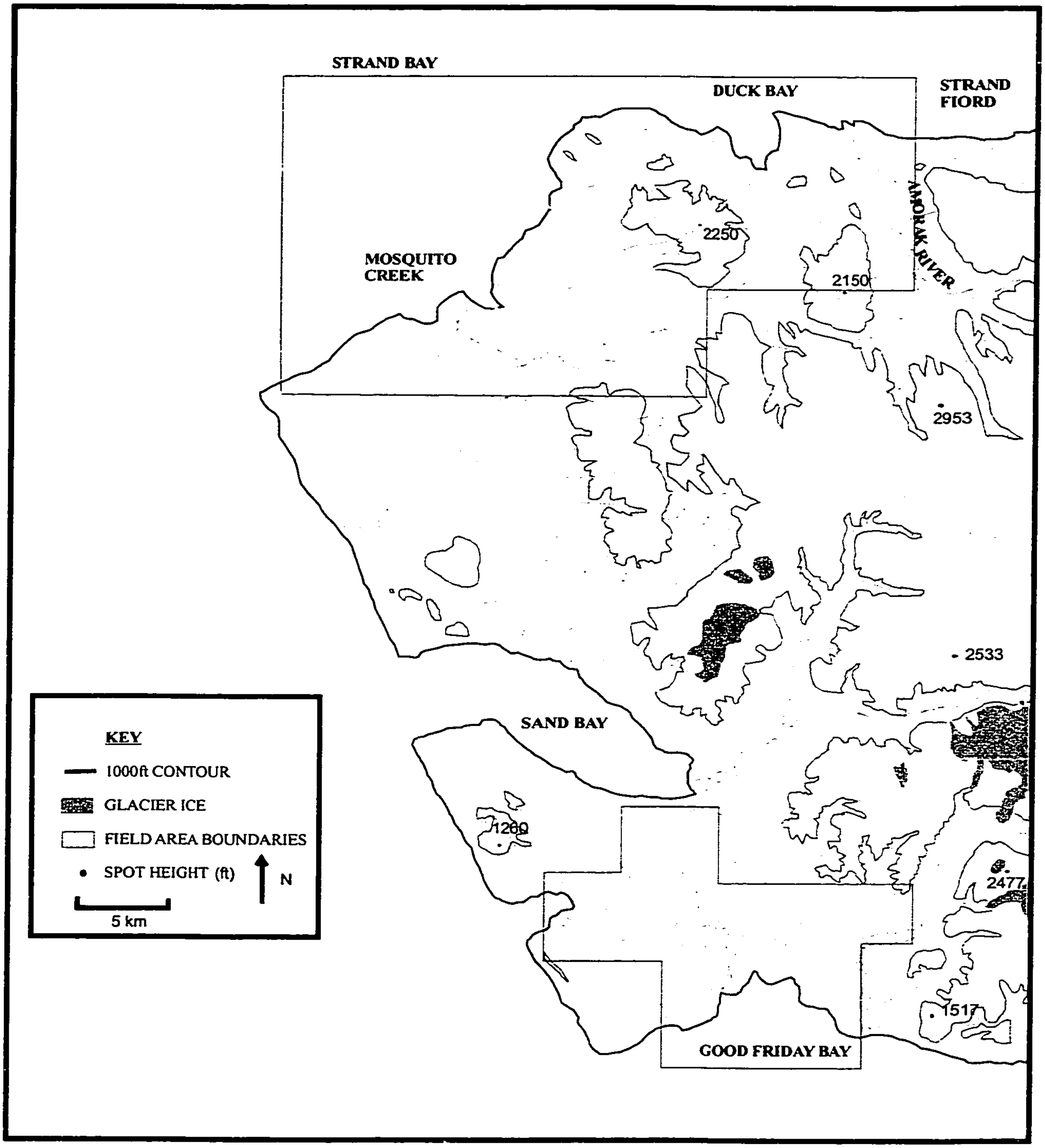


FIGURE 1.5: LANDSAT 7 IMAGE OF STUDY AREA
(Natural Resources Canada 2000)

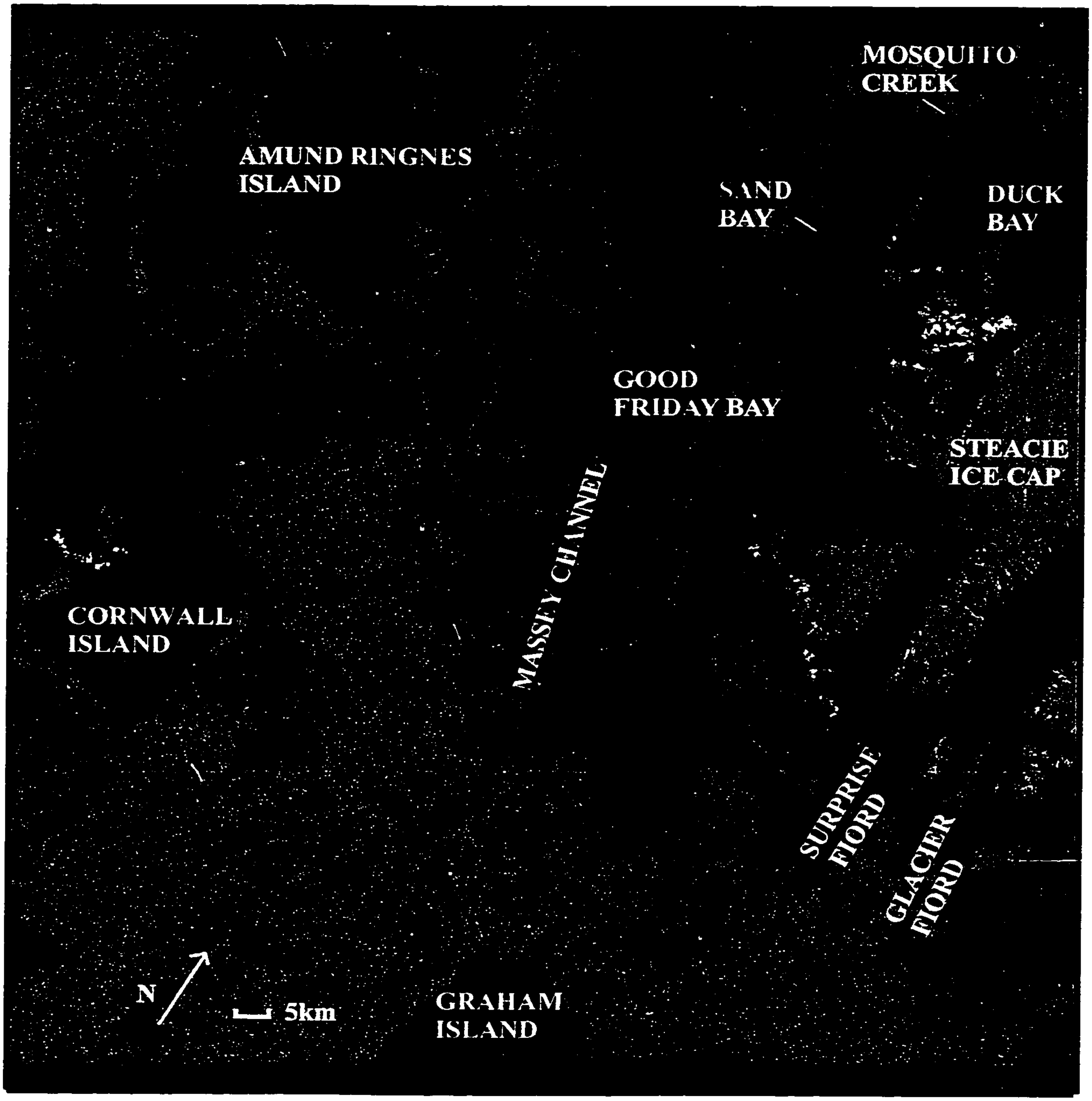


FIGURE 1.6: POSTGLACIAL ISOBASES AT 8.5 ka BP
(England 1999)

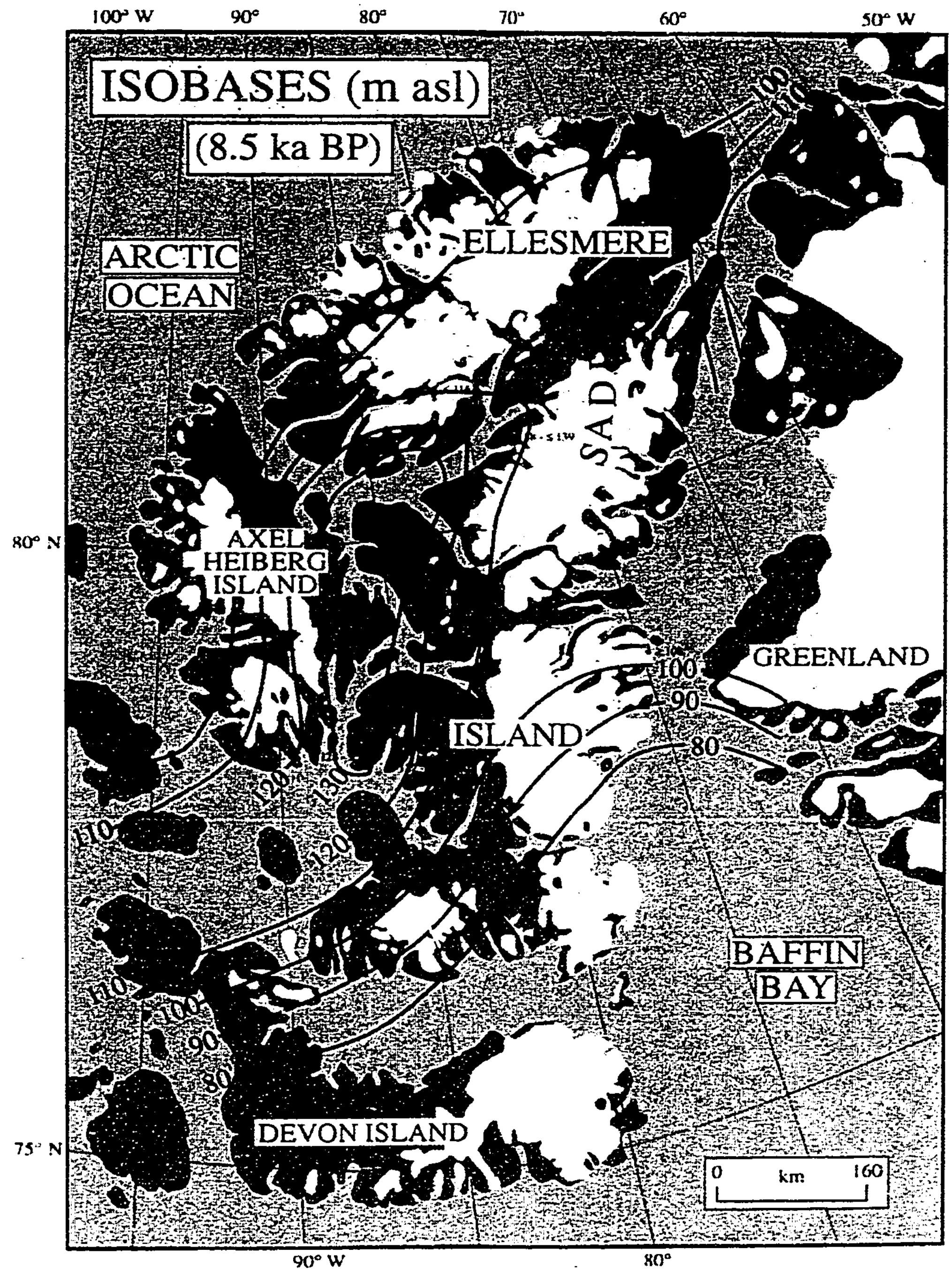
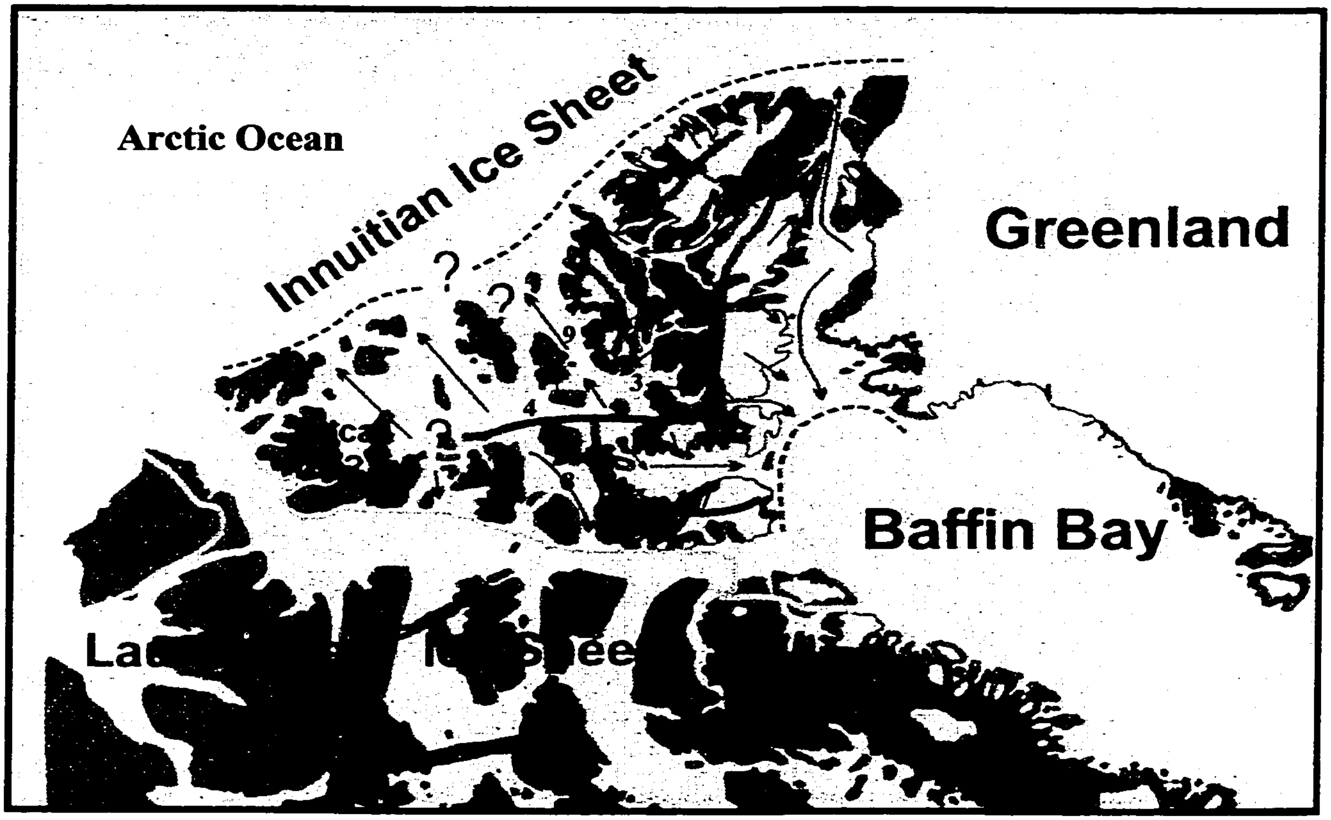
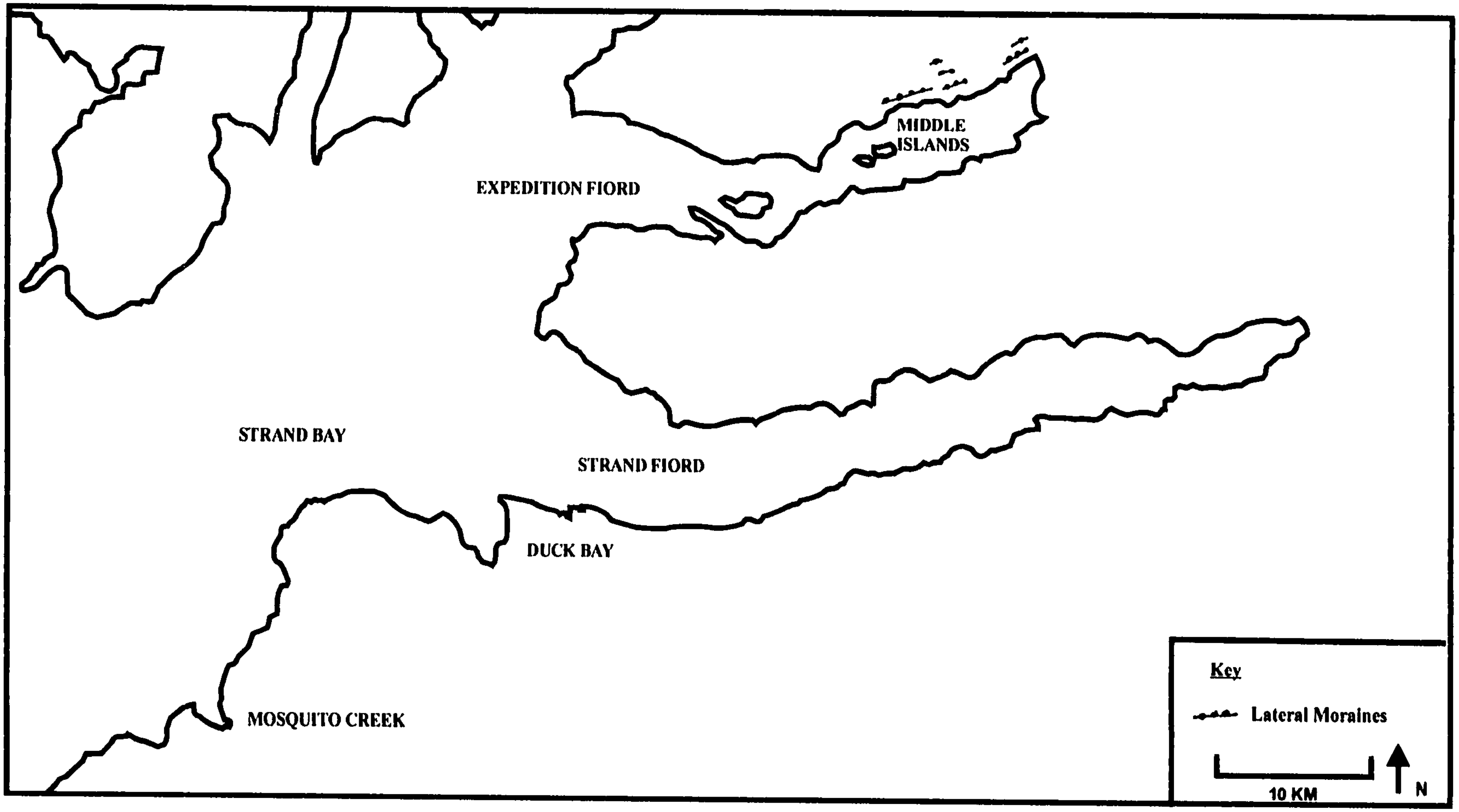


FIGURE 1.7: CONFIGURATION OF THE INNUITIAN ICE SHEET
(Adapted from England 1999)



Key
Margin of Laurentide Ice Sheet
—— Ice Divides
⟶ Ice flow directions
—— Margin of Innuitian Ice Sheet
a, 1 Areas referred to in text

FIGURE 1.8: STRAND AND EXPEDITION FIORDS



CHAPTER 2: METHODS

2.1 INTRODUCTION

The reconstruction of the glacial and sea level history of west-central Axel Heiberg Island involved mapping surficial geology and geomorphology, surveying marine limit and lower relative sea levels and collecting marine fauna related to deglaciation and the pattern of postglacial emergence within the field area.

2.2 METHODS

2.2.1 Surficial Mapping

Prior to initiating fieldwork, aerial photograph interpretation was conducted in order to identify those sites that contained sediments and landforms most relevant to the reconstruction of the glacial, deglacial and sea level history of the field area. In the field, the aerial photos served as traverse maps and original interpretations were evaluated.

Interpretation of surficial geology on aerial photos was based upon morphology and thickness of sediments. Sediments are ascribed to a genetic unit based upon the classification outlined by Dyke (1983). A brief description of the units used in the mapping is presented in Table 2.1. Stratigraphic sections observed in the field were described and recorded using standard stratigraphic and sedimentologic techniques. Particular attention was directed towards identifying sediments and landforms that delimited former ice marginal positions (e.g. moraines, kames and lateral meltwater channels) and towards identifying raised marine sediments that could be investigated for dateable organic material.

Meltwater channels are the most conspicuous ice-marginal landforms on west-central Axel Heiberg Island, particularly at higher elevations. Their distribution delineates former ice marginal positions. These channels are characteristic of cold ice margins where surface meltwater and snowmelt from adjacent slopes are diverted towards ice margins where they erode lateral channels (Dyke 1993). Where such channels grade to dateable marine sediments they enable the chronology of ice retreat to be established. However, caution must be exercised in clearly distinguishing lateral meltwater channels from many larger abandoned channels in the study area which document pre-Quaternary fluvial drainage (A. Dyke pers. comm. 2000). These channels typically exhibit a dendritic pattern and are frequently infilled by till. This suggests that they were carved prior to the LGM and were not significantly altered during postglacial times, otherwise, fluvial incision would have eroded the pre-existing till cover. Furthermore, these fluvial channels are considerably larger than those regarded to be unequivocal meltwater channels (Figure 2.1) and their downstream ends lack deltas that would have formed during Holocene retreat had the channels contained streams that transported outwash. The criteria used to identify meltwater channels include: 1) Channels that crosscut interfluves at an angle that cannot be explained without recourse to a former ice margin and/or 2) Channels that grade to Holocene marine limit. Applying these criteria indicates that meltwater channels are not ubiquitous on west-central Axel Heiberg Island. This may reflect their poor preservation potential in the weak sedimentary bedrock of the study area (Hodgson 1985; Lemmen *et al.* 1994).

2.2.2 Establishing the elevation and age of former relative sea levels.

Surveying raised marine shorelines provides a measure of the amount of postglacial emergence within the field area. The former extent of the sea is determined by the presence of raised marine sediments or landforms (e.g. raised beaches, deltas, washing limits). The maximum elevation attained by the sea along a glacioisostatically depressed shoreline is termed the marine limit. It is recognised as the highest marine landform within the field area. If the establishment of marine limit coincides with a period of low sediment input or marine landforms are reworked by erosion and weathering, marine limit may be misidentified. Where possible, it is determined using several accordant landforms. Where marine landforms are inconspicuous, the maximum elevation of raised marine sediment provides a minimum estimation for marine limit. The elevation of these sites was measured using a Wallace and Tiernan micro-altimeter (accuracy +/- 2m at 100m asl) and the sites were fixed by a Global Positioning System. The altimeter is temperature and pressure dependent. Temperature corrections were made for each measurement using standard temperature correction tables. Transects were closed as frequently as possible using high tide as the datum in order to allow for pressure correction. Corrections were applied to measurements based on the rate of pressure change throughout the transect period. Whenever key sites were affected by changes in atmospheric pressure they were re-surveyed.

In order to determine the age of a paleoshoreline, radiocarbon dating was performed upon marine fauna affiliated with marine landforms. To provide accurate age determinations, dateable organic deposits must be closely associated with specific paleoshorelines (Andrews 1970). For example, paired mollusc valves located in growth position can be unambiguously related to former sea levels when their stratigraphic position occurs within topset, foreset or bottomset beds of surveyed marine deltas. Mollusc collections from undifferentiated marine sediments may also be radiocarbon dated. However, as molluscs live at varying water depths it is difficult to relate such samples to specific former sea levels. Samples lacking a clear stratigraphic context provide a minimum age estimate for the establishment of a higher paleoshoreline. Accurate radiocarbon dating is also dependent upon the quality of the material submitted for analysis. Submission of paired valves is preferable. Collections of fragmented molluscs are more problematic as they may contain a range of ages. In those cases where samples consisting of fragmentary material were considered relevant, individual fragments were submitted for AMS dating. Molluscs are also subject to contamination, most commonly from the encrustation of secondary calcite but also by lichen or algal growth (Dyke *et al.* 1991). Surface contaminants may have a different age than the original shell and provide incorrect ages if dated. Careful selection of molluscs in the field can minimise this potential source of error. In order to remove any visible surface contaminants, mollusc samples submitted for radiocarbon dating were cleaned in an ultrasonic bath containing deionised water for 20 minutes. They were subsequently air-dried and cleaned using an electric sander and a hand pick. Remaining abraded material was removed by resubmersion in the ultrasonic bath for a further 20 minutes. Samples were finally weighed and packaged into sterile bags. Larger samples (~40g) of whole valves were submitted to Beta Analytic (Miami) for conventional radiocarbon analysis (beta counting). Individual mollusc fragments were submitted to Lawrence Livermore National Laboratory (California) for AMS dating.

TABLE 2.1: SURFICIAL MAPPING SCHEME

UNIT	CHARACTERISTICS	NOTES
Rock.	Bedrock.	Conspicuous bedding exposed.
Residuum.	In situ weathered bedrock.	Overburden conforms to geologic structure.
Till veneer.	<2m thick. Masks underlying topography and bedrock structure but does not completely obscure it.	Till tends to be locally well-vegetated due to increased moisture content. Well-vegetated tills produce darker tones on aerial photos. Poorly vegetated tills produce lighter tones.
Till blanket.	> 2m thick. Obscures underlying bedrock structure.	Ice-wedge polygons are useful in distinguishing till blankets.
Glacifluvial	Fluvial facies deposited subglacially or proglacially.	Gravel and sand deposits located beyond existing ice margins and outside modern drainage basins.
Fluvial (active).	Fluvial deposits within the range of modern flooding. Exhibit channel scarred surfaces that are unvegetated.	Located within active floodplains.
Fluvial (inactive).	Fluvial deposits generally separated from the active floodplain by a bluff ≥ 1 m high. Tend to be vegetated.	Form prominent terraces.
Colluvium.	Weathered materials transported downslope by mass movement.	Widespread at higher elevations also commonly override valley walls.
Raised Marine.	a) Deltaic: often characterised topset, foreset and bottomset beds. b) Nearshore: prodeltaic, lagoonal or foreshore deposits. c) Beach: raised beach (littoral) deposits.	Fine sand, silt and clay characterised by ice-wedge polygons and earth hummocks.

FIGURE 2.1: MELTWATER CHANNELS AND PRE-GLACIAL FLUVIAL CHANNELS, WEST-CENTRAL AXEL HEIBERG ISLAND

KEY

Meltwater channels (barb on up-ice side)

Pre-glacial fluvial drainage

CHAPTER 3: SURFICIAL GEOLOGY AND GEOMORPHOLOGY

3.1. INTRODUCTION

This chapter presents the surficial geology and geomorphology of west-central Axel Heiberg Island. Mapping followed the scheme outlined by Dyke (1983; Section 2.2). The Duck Bay-Mosquito Creek and Good Friday Bay field areas were ground-truthed in the field whilst the remainder of the study area was mapped by aerial photo interpretation. The relevance of the surficial geology and geomorphology to the reconstruction of the late Quaternary glacial and sea level history of west-central Axel Heiberg Island is addressed in Chapter 4.

3.2 ROCK AND RESIDUUM

3.2.1 Duck Bay-Mosquito Creek

Bedrock and residuum dominate the surficial geology of Duck Bay and Mosquito Creek (Figure 3.1). Three geological formations underlie the field area, outcropping as northwest-southeast trending synclines and anticlines (Figure 1.2). They consist of the upper Jurassic and lower Cretaceous Deer Bay Fm (shale, siltstone, sandstone and mudstone); the lower Cretaceous Isachsen Fm (sandstone, shale, siltstone, igneous intrusions and pyroclastic rocks) and the lower Cretaceous Christopher Fm (shale, sandstone, siltstone, mudstone and pyroclastic rocks). The weathering characteristics of the formations were documented by Hodgson (1982) on the islands to the west of Axel Heiberg Island and are similar within the field area. Shale of the Christopher Fm weathers to silty clay with a cobble-granule lag of mudstone and ironstone. Weathering of the Isachsen Fm produces silt, clay and coarse sand with a cobble-granule lag of siltstone, mudstone and ironstone. Around Duck Bay, well-cemented quartzose sandstone of the Isachsen Fm outcrops in strike-aligned ridges that trend northwest-southeast. On more resistant lithologies such as the quartzose sandstone of the Isachsen Fm, frost shattering has produced coarse residuum and felsenmeer that conform to the underlying bedrock structure. Shale of the Deer Bay Fm weathers to clayey silt with a lag of siltstone and ironstone granules.

Small-scale landforms of glacial erosion (e.g. striae, ice-moulded bedrock) were not observed on the bedrock of the field area. This is probably due to its poor lithification that promotes rapid weathering and inhibits the preservation of such landforms (Lemmen *et al.* 1994). Their absence may also be due to the presence of non-erosive, cold-based ice within the field area (cf. Dyke 1993).

3.2.2 Good Friday Bay

Bedrock and residuum are also common at Good Friday Bay (Figure 3.2). The Deer Bay, Isachsen and Christopher Fms outcrop along the north-central shore of Good Friday Bay in addition to the late Jurassic Awingak Fm (quartzose sandstone, siltstone and shale) and Cretaceous gabbro, diabase and basalt dykes (Figure 1.3). Weathering of the Awingak Fm creates fine to coarse-grained sand and silt with a discontinuous lag of flaggy sandstone clasts (Hodgson 1982). Dykes generally weather to angular rock fragments of varying size and often outcrop as linear, castellated features (Hodgson 1982). Frost shattering in the west of the field area has produced northwest-southeast strike-aligned bands of coarse, angular debris that conform to the underlying bedrock (Plate 3.1).

As at Duck Bay-Mosquito Creek, small-scale landforms of glacial erosion were not observed. However, occasional diabase erratics located on residuum attest to former glaciation within the field area (Plate 3.1)

3.3 BEAUFORT FORMATION

The Beaufort Formation consists of unlithified, flat-lying clay, silt, sand and gravel commonly containing woody debris (Fyles 1990; Miall 1991). Beaufort Fm sediments extend from Meighen Island to Banks Island (Miall 1991) and are regarded to be fluvial in origin, deposited in the late Tertiary by rivers flowing northwestward across a contiguous Arctic landmass to the Arctic Ocean (England 1987).

A north-south trending sinuous ridge extends for ~4km between Good Friday Bay and Sand Bay (Plate 3.2; X, Figure 3.2). The ridge is composed of unconsolidated, quartz-rich fine to medium sand with a thin (<50cm) pebble-boulder cap. The sand is typically planar or cross-bedded (Plate 3.3) and individual beds range in thickness from 5mm-20cm. The northern end of the ridge consists of alternating beds of laminated (5mm-3cm) fine sand and mud (Plate 3.4). The aforementioned sediments are similar to those documented by Fyles (1990; p.395) who interpreted them as characteristic of a braided river environment and ascribed them to the Beaufort Fm. The sediments within the ridge at Good Friday Bay are similarly interpreted and represent the first documentation of the Beaufort Fm on western Axel Heiberg Island.

The pebble-boulder cap is composed of variable local lithologies (sandstone, siltstone and diabase) which are striated and faceted. The sediment is interpreted as till.

3.4 TILL

3.4.1 Duck Bay-Mosquito Creek

Till is differentiated from the underlying weathered bedrock by the presence of erratics and striated and faceted clasts. Outcrops of till around Duck Bay occur as a discontinuous veneer that masks but does not obscure the underlying bedrock (<2m thick, Dyke 1983; Figure 3.1). Average thickness of the till is ~50cm and the maximum observed thickness was 2m. Till composition is influenced by the underlying bedrock such that till overlying shale has a clay-rich matrix and till overlying sandstone has a sand-rich matrix. Erratics of local provenance (e.g. sandstone, siltstone, mudstone, diabase) are commonly exposed on the till surface. Till veneer tends to be poorly to moderately vegetated (~50% groundcover). Tundra hummocks ranging from ~0.5-1m in diameter are prevalent in more densely vegetated areas. Till veneer is ubiquitous at Mosquito Creek (Figure 3.1). Although vertical sections are rare, sub-surface excavations indicate that the till has a fined-grained matrix (clay-silt). This is also evident from the abundant earthflows on the till surface. Erratics of local provenance are widespread within the till.

Till is mapped as a blanket where the form, composition and structure of underlying bedrock is completely obscured, typically at a thickness of >2m (Dyke 1983). Till blankets tend to be more common on valley floors throughout the field area. Aerial photo analysis reveals conspicuous networks of ice-wedge polygons within the till in the main river valley at Mosquito Creek. The polygons cover an area of ~9km^2 (adjacent to A, Figure 3.1). Dyke *et al.* (1992) state that polygon size is inversely proportional to the coefficient of thermal expansion of the containing sediment, in particular, the material lying below the permafrost table. For example, fine grained till (typical in the study area) has a higher thermal expansion coefficient than fluvial gravel and should consequently exhibit smaller polygons (Dyke *et al.* 1992). However, the tundra polygons mapped on the aerial photos indicate that till blankets actually display large features (~50-100m). One possible explanation is that the till blankets are underlain by massive ice. Ice has a low coefficient of thermal expansion and may account for the creation of large polygons (Dyke *et al.* 1992). Throughout west-central Axel Heiberg Island, large polygons are located above marine limit and adjacent to ice marginal landforms. This suggests that the underlying ice may be of glacial origin. The identification of polygonised terrain from aerial photo analysis may indicate the presence of ice-cored till within the study area.

3.4.2 Good Friday Bay

Till at Good Friday Bay commonly obscures underlying bedrock and is consequently mapped as a blanket (Figure 3.2). Ice-wedge polygons are widespread at the mouth of the main valley that enters the east side of the field area and on the plateau bordering its northern side (Figure 3.2). Till veneer is also common throughout the field area (Figure 3.2). Erratics of local provenance are prevalent within the tills. A morainal bank and an end moraine encircle the mouth of the main valley that enters the east side of the field area and a lateral moraine flanks the valley wall to the north (A, B, C; Figure 3.2). The till blanket on the west side of the valley mouth also exhibits northeast to southwest oriented flutings (D; Figure 3.2). Together, these landforms document ice flow from the mountainous interior into the lowlands of Good Friday Bay.

3.5 GLACIFLUVIAL SEDIMENTS AND LANDFORMS

3.5.1 Duck Bay-Mosquito Creek

Lateral meltwater channels are the most conspicuous ice marginal landforms within the Duck Bay-Mosquito Creek field area (Figure 3.1). They dissect bedrock, residuum and till and are particularly prominent within the more resistant sandstone of the Deer Bay and Isachsen Fms. Their distribution delineates the configuration of former ice margins during successive stages of ice retreat and their chronology is determined where they contact dateable marine sediments. Around Duck Bay-Mosquito Creek, meltwater channels extend from ~650m asl to marine limit (73m asl) which is marked by an ice-contact delta at Mosquito Creek (A, Figure 3.1). These channels record the thinning of ice within Strand Fiord and its retreat from the coast into the main valleys of the field area. As meltwater channels are important to the reconstruction of the Quaternary history of the field area they are discussed further in Chapter 4.

Glacifluvial sediments constitute a small percentage of the surficial cover within the field area. At the ice-contact delta on the south shore of the main (unnamed) river at Mosquito Creek (A, Figure 3.1), hummocky gravels comprised of sub-angular to sub-rounded sandstone (Plate 3.5) grade to proglacial outwash. These, in turn, extend to foreset bedded sand of the delta. The hummocky gravel may record the uneven meltout of buried ice which would have been determined by the relative thickness of the overlying debris cover and the depth of the active layer.

3.5.2 Good Friday Bay

Lateral meltwater channels are prominent around Good Friday Bay where they also dissect bedrock, residuum and till (Figure 3.2). They extend from the margins of existing ice caps (~760m asl) to marine limit (~120m asl) and record the occupation of Good Friday Bay and Sand Bay by ice that overtopped the lowlands separating the bays (Figures 3.2, 3.3). Lateral meltwater channels record the thinning and retreat of ice into the main valleys surrounding the field area.

Glacifluvial deposits constitute a small proportion of the surficial sediments at Good Friday Bay. Glacifluvial gravels are located on a plateau to the east of the field area where a low (< 2m high), sinuous ridge composed of rounded pebbles and cobbles extends for ~250m (Figure 3.2; Plate 3.6). The ridge is interpreted as a kame based upon its morphology and sedimentology.

3.6 RAISED MARINE DEPOSITS

During glaciation the west-central coast of Axel Heiberg Island was subjected to glacioisostatic loading and depression (Lemmen *et al.* 1994). Consequently, marine transgression across the depressed land surface accompanied ice retreat. Submerged coastlines temporarily received large volumes of sediment and subsequent postglacial emergence has exposed these. Raised marine sediments are divided into deltaic, nearshore and beach facies (cf. Dyke 1983; Section 2.2.1).

3.6.1 Duck Bay-Mosquito Creek

Well-preserved deltas are rare within the field area. This is largely a consequence of fluvial incision and lateral migration of rivers during postglacial emergence that has eroded away the delta bedding. Deltas occur above the contemporary shoreline at Duck Bay where they range in elevation from 68 to 10m asl (Figure 3.1). The 68m asl delta provides a minimum estimate of marine limit at Duck Bay (B, Figure 3.1). At Mosquito Creek, deltas occur from marine limit at 73m asl (marked by the ice contact delta: A, Figure 3.1) to 10m asl. A prominent delta forms a terrace at 38m asl with well-defined foreset beds and extends downvalley for 4km flanking the southern margin of the unnamed river (C, Figure 3.1). Deltas within the field area are only rarely fossiliferous.

Fine-grained nearshore sediments mantle the coastlines of the field area below marine limit. These sediments are typical in proglacial environments where the discharge of sediment-rich subglacial meltwater from nearby ice margins and/or the oversteepening and failure of delta slopes creates turbid overflows (cf. Powell 1984; Aitken & Bell 1998). Suspension settling from these currents commonly produces lithofacies consisting of massive or laminated sand, silt, and mud (cf. Stewart 1991). Clast-rich muds are common in ice-proximal settlings where they indicate the deposition of ice-rafted sediments (cf. Powell 1984). In the study area these muds are often highly fossiliferous.

Nearshore deposits are often poorly drained and develop wet-meadow vegetation that is visible as dark patches on aerial photographs (cf. Dyke 1983). At Duck Bay, an extensive blanket of marine fines at 31m asl has been fluvially incised to produce a terrace that flanks the west bank of the Amorak River for ~7km (D, Figure 3.1). The sediments are composed of thin (5mm-1cm) rhythmically bedded fine to medium sand and silty mud that indicate suspension settling from turbidity currents in a prodeltaic environment (cf. Aitken & Bell 1998).

Within the field area, raised beaches are rare. This may be attributed to a lack of fetch caused by pervasive summer sea ice, destruction by mass movement or a combination of the above. Beaches occur as isolated patches of sand with a sub-angular to sub-rounded gravel lag. These sediments have a limited extent and are not represented on Figure 3.1.

3.6.2 Good Friday Bay

Raised marine sediments are prolific around Good Friday Bay (Figure 3.2) where marine limit is marked by a delta at 123m asl (E, Figure 3.2). The delta is composed of foreset fine-medium sand with sporadic granule beds (~2cm in depth). The foresets are occasionally disturbed by cobble-sized dropstones and they grade upward into clast-supported gravel that has a matrix of medium sand. The gravel forms a cap of ~1m in depth and contains rounded to sub-rounded clasts that often exhibit faceting and striations. The foresets of the delta are of a similar composition to the sands observed in the Beaufort Fm (quartz-rich fine to medium sand) to the north, the erosion of which may have provided a sediment source for the delta. Erosional channel cuts in the gravel cap have been infilled with imbricated clasts that indicate a paleo-flow direction to the southwest. The channelised nature of the gravels may indicate that they were deposited in a subaerial environment close to high tide whilst the dropstones in the forest beds suggest proximity to a nearby ice margin. Farther east, meltwater channels descend to a delta composed of medium sand foresets at 119m asl (F, Figure 3.2). On the east side of the main river entering Good Friday Bay, a lateral meltwater channel descends to an ice-contact delta at 120m asl (G, Figure 3.2).

Deltaic sediments dominate the lowland of Good Friday Bay. This is likely due to the weak bedrock of the study area that would have been easily eroded by meltwater during deglaciation. Deltaic sediments extend from 10 to 45m asl and cover an area of 3km^2 (Plate 3.7). The deltas record progradation into a falling sea level following deglaciation (Andrews 1970). They consist of fine-medium grained foreset sand that is occasionally fossiliferous. In one case, foreset beds can be traced to topset beds at 45m asl. The bottomset beds are composed of laminated to massive silty mud that is highly fossiliferous.

Elsewhere, nearshore sediments consist of finely laminated sand, silt and mud deposited by suspension settling. These sediments are commonly fossiliferous and contain abundant ice-rafted pebbles, cobbles and boulders.

Raised beaches consist of medium to coarse sand with a pebble-cobble lag of local lithologies. They occur as isolated patches and are unmapped due to their restricted extent.

3.7 ALLUVIUM

3.7.1 Duck bay-Mosquito Creek

Active alluvium at Duck Bay is confined to the floodplain of the Amorak River. The floodplain is ~200m wide and exhibits sand-dominated channels and bars. Inactive sandy alluvium forms vegetated terraces flanking the river ~1-2m above the modern sandur Thermal erosion of permafrost within marine sediments on the west bank has created prominent thermo-erosional scarps that are ~20m high (Plate 3.8). On the southeastern shore of Duck Bay a well vegetated terrace consisting of inactive gravel alluvium lies ~2-3m above the modern channel. At Mosquito Creek the active floodplain of the main river is composed of gravel channels and bars. Inactive gravel alluvium forms terraces bordering the river ~1-3m above the active channel.

3.7.2 Good Friday Bay

The lowland terrain of Good Friday Bay contains widespread alluvium (Figure 3.2). Active alluvium within the main river to the east is sand-dominated and is confined to the floodplain that is up to 2.5km wide. Inactive sand-dominated alluvium borders the river, forming terraces ~1-2m above the active channel. To the west, active alluvium is sand-dominated and inactive sandy alluvium forms terraces ~1m above the active channels. Farther upstream active and inactive alluvium becomes gravel dominated.

3.8 COLLUVIUM

3.8.1 Duck Bay-Mosquito Creek

Coalescent talus cones mantle the base of the glacial trough occupied by the Amorak River (Figure 3.1). Earthflows are common on fine-grained bedrock, especially shale which is prevalent in all bedrock formations in the field area. Gelifluction is ubiquitous on gently to moderately inclined slopes composed of unconsolidated sediment such as colluvium, till or raised marine deposits.

3.8.2 Good Friday Bay

Talus cones and aprons are common in the mountainous terrain in the northeast of the field area (Figure 3.3). Earthflows are widespread within fine-grained bedrock, especially the shale of the Deer Bay Fm that outcrops in the west of the field area. Gelifluction is also ubiquitous within the Good Friday Bay field area.

3.9 SURFICIAL GEOLOGY AND GEOMORPHOLOGY OUTSIDE DESIGNATED FIELD AREAS

Logistical constraints prevented ground-truthing between the Duck Bay-Mosquito Creek and Good Friday Bay field areas. The surficial geology and geomorphology of this area is documented by aerial photo interpretation and is presented on Figure 3.3. Bedrock and residuum form the dominant surficial cover. Till veneer drapes the low plateau separating Mosquito Creek and Sand Bay. The lowlands surrounding Sand Bay are dominated by alluvium, raised marine sediments and occasional till blankets. A delta within these lowlands marks marine limit at 123m asl (A, Figure 3.3). To the east, the floor of the main valley entering Sand Bay is dominated by alluvium and raised marine sediments. The head of this valley bifurcates into northeast and east trending arms. Lateral moraines flank the walls of the eastern arm, demarcating the margins of a former tributary glacier. Meltwater channels extend from existing ice caps (~800m asl) to marine limit (123m asl) and show that ice extended offshore beyond the mouth of Sand Bay before retreating into adjacent tributary valleys (Figure 3.3).

3.10 LICHEN-KILL ZONES

"Lichen-kill" zones are areas of the ground surface where vegetation cover is absent or diminished relative to that on adjacent terrain. These zones are common in the High Arctic (cf. Ives 1962; Locke & Locke 1977; Falconer 1966) where they document the expansion of snow and ice cover during the Little Ice Age (LIA; ~100-400 cal BP). For example, dead moss collected from a lichen-kill zone exposed by the retreat of the Tiger Ice Cap on northern Baffin Island was radiocarbon dated at 330±75 yrs BP, placing the chronology of the lichen-kill zone within the LIA (Falconer 1966). During the LIA, the High Arctic is regarded to have been both colder (cf. Bradley 1990) and wetter (cf. Lamoureux 2000) therefore lowering the regional equilibrium line altitude (ELA). Ice core records from Agassiz Ice Cap on Ellesmere Island indicate a cooling of summer temperatures during the LIA (cf. Bradley 1990). Consequently, this would have facilitated glacial advance and the growth of permanent snow patches that would have destroyed pre-existing lichen cover and produced areas of lichen-kill. These areas can be mapped from aerial photos as they typically appear as lighter-toned patches due to their higher reflectance. As the mean elevation of permanent snow cover is an approximate measure of the equilibrium line altitude (ELA) of an area (cf. Locke & Locke 1977), demarcating former snow cover from the distribution of lichen-kill zones provides a measure of paleo-ELA. This information is relevant to reconstructing the climatic regime of west-central Axel Heiberg Island during the LIA. A conservative approach was adopted to mapping lichen-kill zones due to the widespread poorly lithified bedrock of the study area which precludes lichen growth and produces lichen free areas due to geological rather than climatic factors (cf. Beschel 1961). In addition, sandstones of the study area often have a high reflectance and a similar appearance to lichen-kill zones, especially where they outcrop along ridges.

Evidence of expanded glacial cover on west-central Axel Heiberg Island is limited to terminal moraines that show restricted local ice advance during the LIA (B, C; Figure 3.3). However, lichen-kill zones demarcating areas of permanent snow cover are widespread. They are commonly located a few hundreds of meters beyond modern ice margins and on high plateaux where snow can readily accumulate (Figure 3.4). These surfaces often appear mottled on aerial photos where vegetation is interspersed with lichen-free patches. Such areas depict the former position of individual snowbanks (Figure 3.4). Throughout the field area, lichen-kill zones are consistently located at ~610m asl. This provides an approximation of the ELA for west-central Axel Heiberg during the LIA.

Small cirque glaciers are common within the highlands of the study area and are preferentially located on north and northeast facing slopes (A,B,C; Figure 3.4). Most of these glaciers appear completely exposed (snow-free) on the aerial photos (A,B,C; Figure 3.4; Aerial photos taken in July 1958 and 1959). Their persistence may be attributed to aspect as cirques on south and west facing slopes, that receive greater summer insolation, are presently ice-free. The exposed cirque glaciers are considered remnants from the LIA that currently lie below the modern ELA and are probably experiencing a negative net mass balance. They are located consistently at ~610m asl suggesting that the modern ELA lies at a higher elevation. On the same aerial photos, permanent snow cover (indicated by a high albedo on aerial photographs) is identified at ~850m asl. Due to the high annual variability of High Arctic ELA's (Bradley & England 1978) and the possibility that snowmelt may have occurred after the aerial photos were taken, 850m asl is best regarded as a minimum estimate of modern ELA. Miller *et al.* (1975) proposed that the modern ELA for western Axel Heiberg Island was 400-600m asl. The modern ELA proposed in this study (~850m asl) suggests that Miller *et al.'s* ELA estimate should be revised, at least locally. Indeed, the ELA of White Glacier at the head of Expedition Fiord to the north has been measured at ~970m asl (Cogley *et. al* 1996).

Given the ELA estimate of ~610m asl for the LIA and the modern ELA estimate of 850m asl, it appears that the ELA on west-central Axel Heiberg Island during the LIA was depressed by ~240m. This was likely due to a combination of decreased temperatures and increased precipitation over the High Arctic during this period (cf. Bradley 1990; Lamoureux 2000).

KEY FOR FIGURES 3.1, 3.2, 3.3.

ROCK AND RESIDUUM	CONTOURS
BEAUFORT FORMATION	RIVERS
TILL BLANKET	LATERAL MELTWATER CHANNELS (barb points up-ice)
TILL VENEER	MORAINES (end; lateral)
GLACIFLUVIAL	FLUTINGS
RAISED MARINE	KAMES
ALLUVIUM (active)	DELTAS
ALLUVIUM (inactive)	a,b... LANDFORMS REFERRED TO IN TEXT
COLLUVIUM	ROCK GLACIERS
ICE	

FIGURE 3.1: SURFICIAL GEOLOGY AND GEOMORPHOLOGY, DUCK BAY-MOSQUITO CREEK (STRAND BAY)

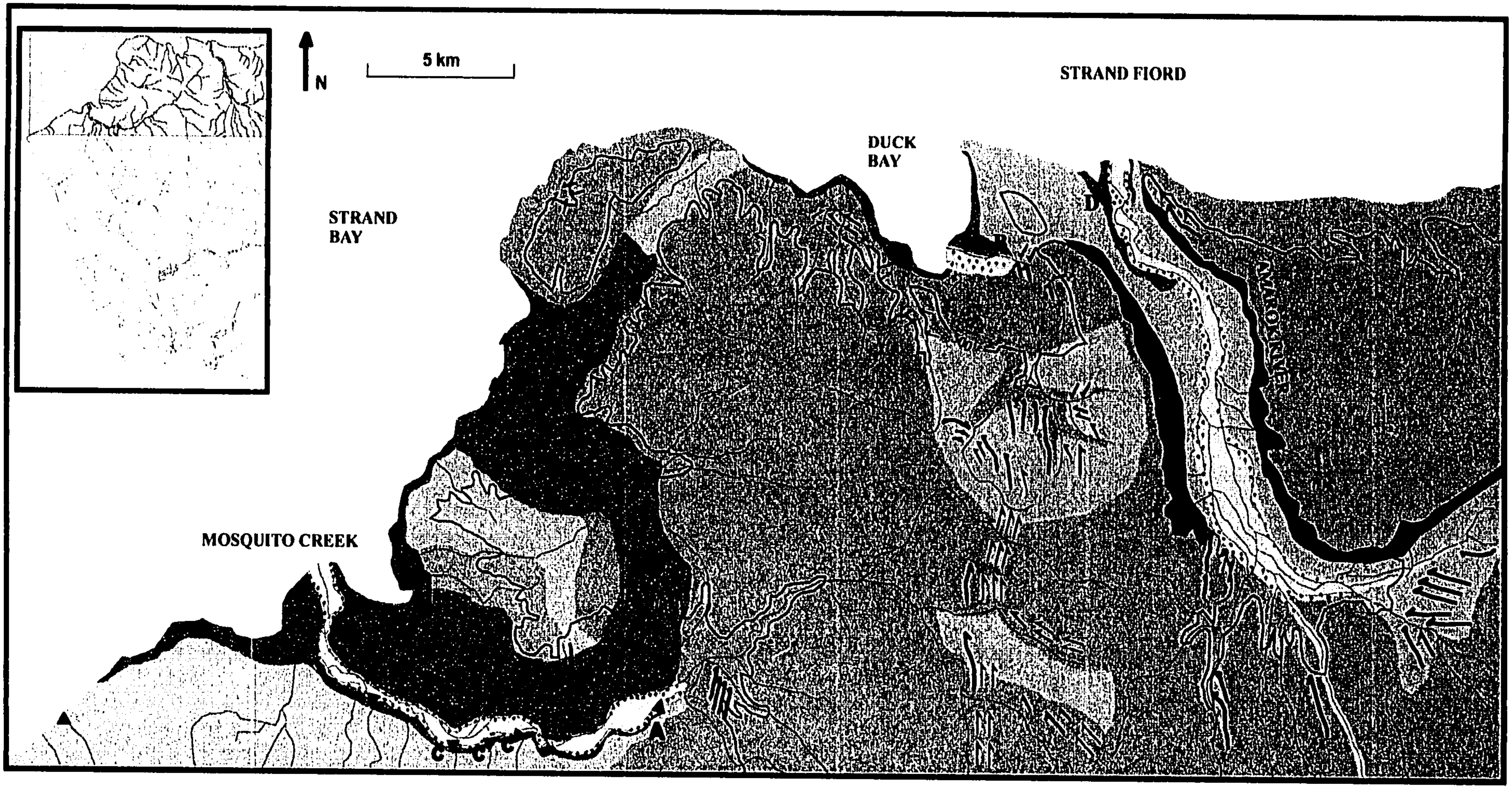


FIGURE 3.2: SURFICIAL GEOLOGY AND GEOMORPHOLOGY OF GOOD FRIDAY BAY

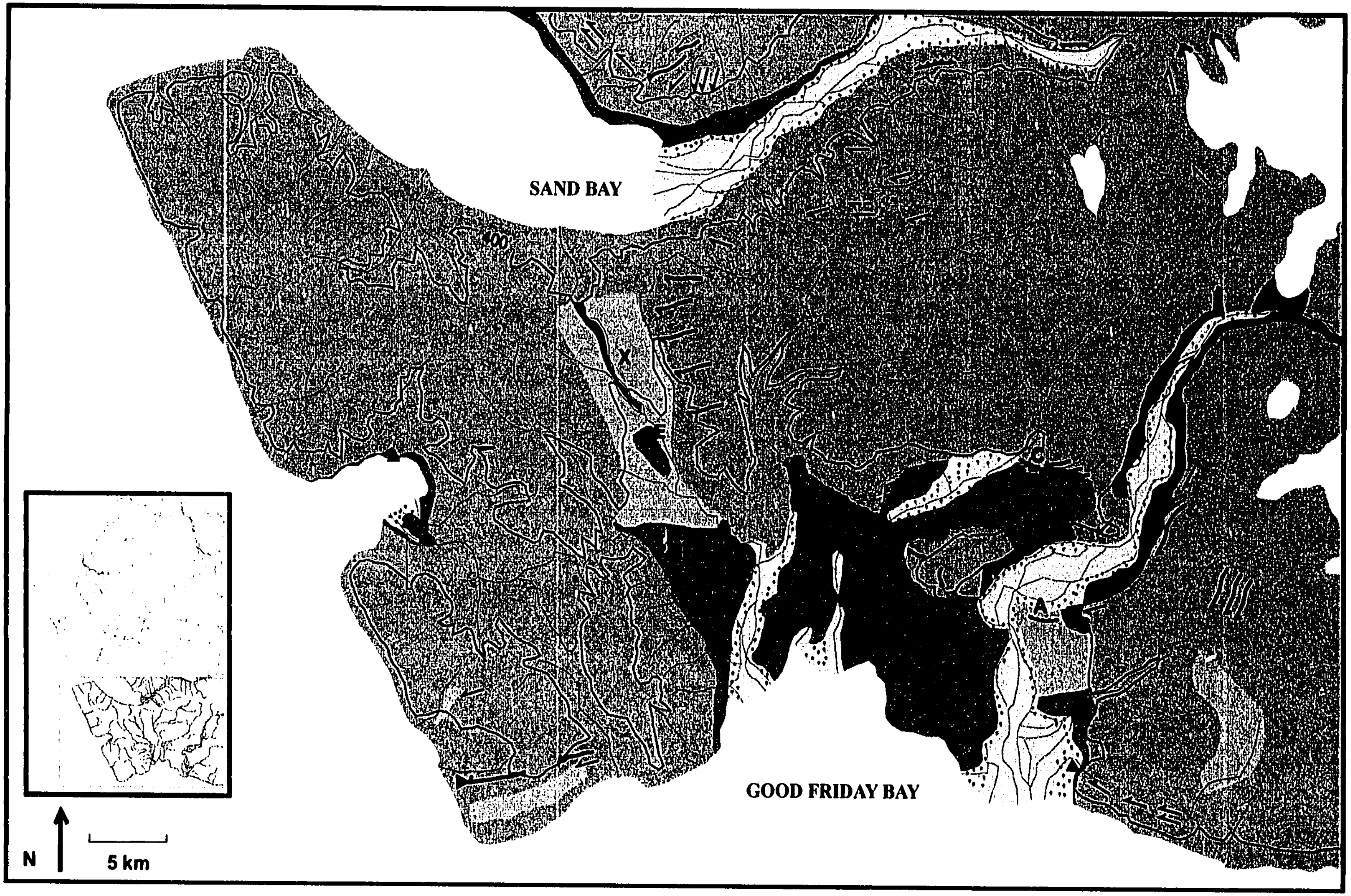


FIGURE 3.3: SURFICIAL GEOLOGY AND GEOMORPHOLOGY OF INTERVENING LANDSCAPE

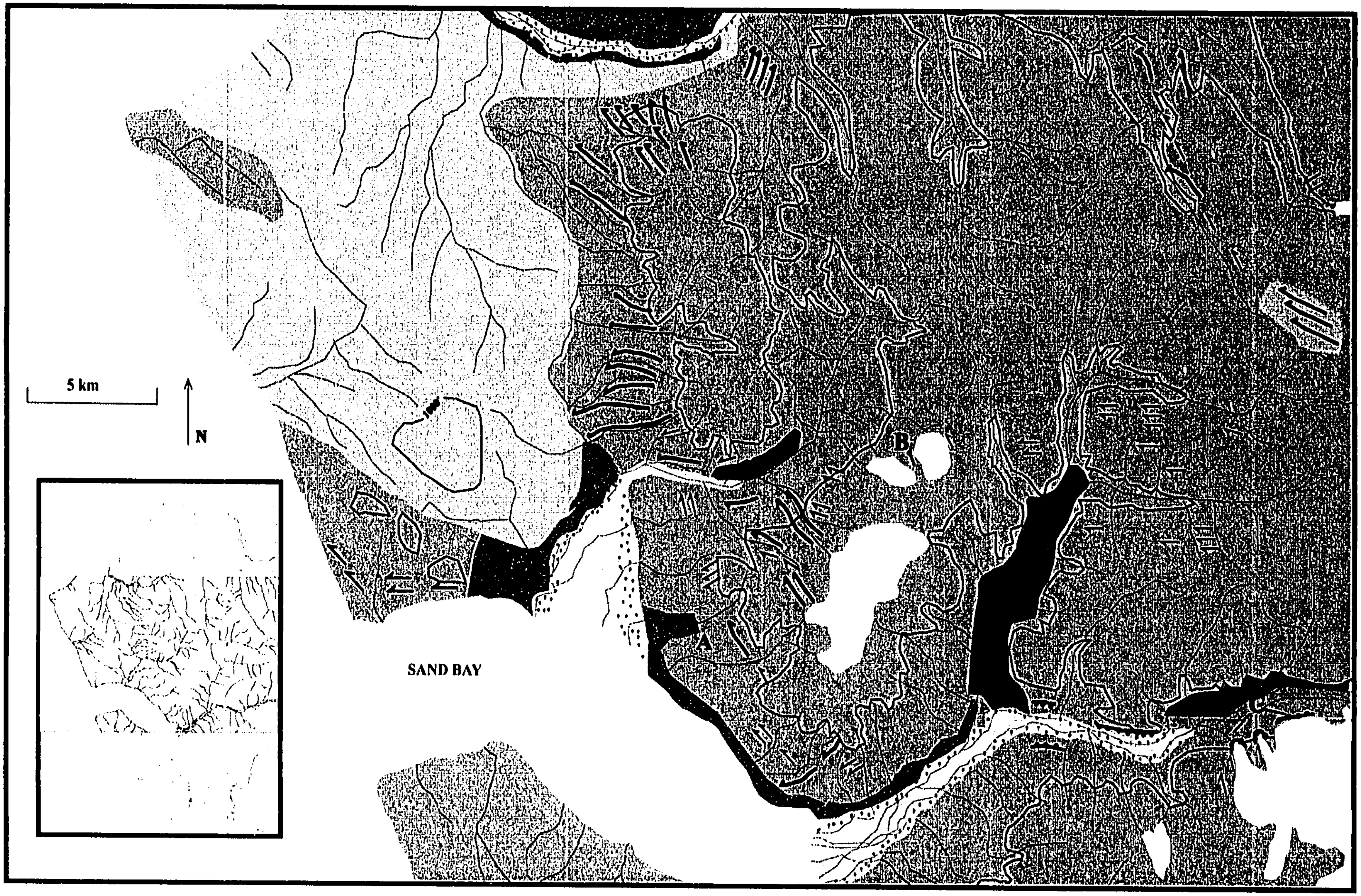


FIGURE 3.4: LICHEN-KILL ZONES, WEST-CENTRAL AXEL HEIBERG ISLAND

PLATE 3.1

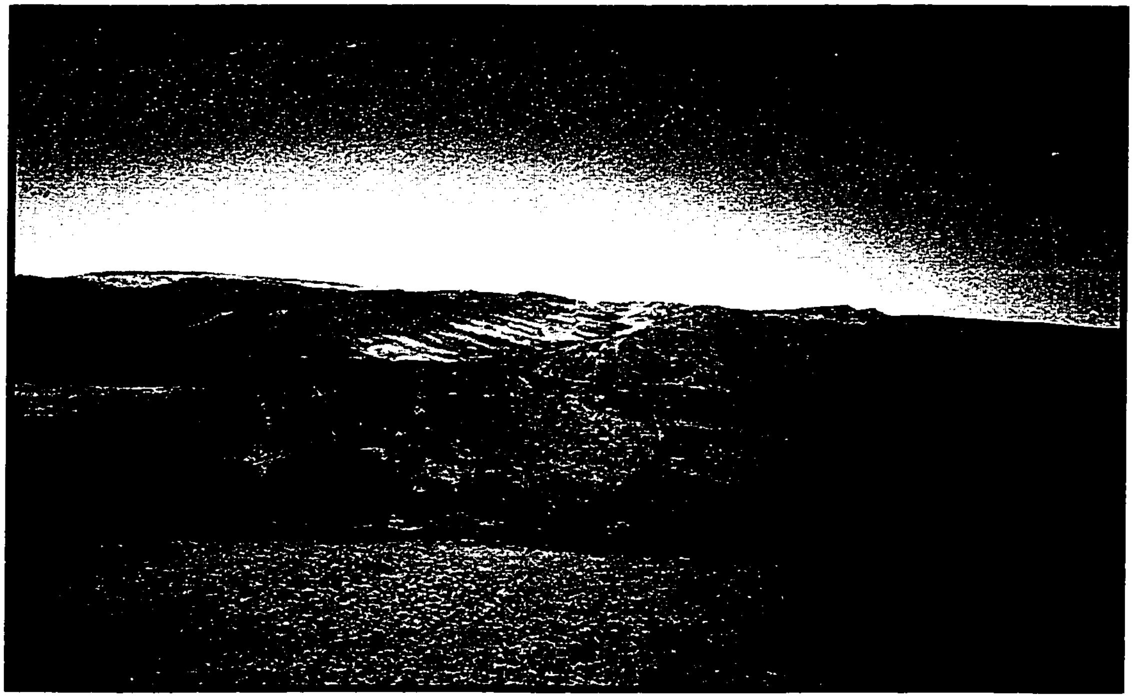
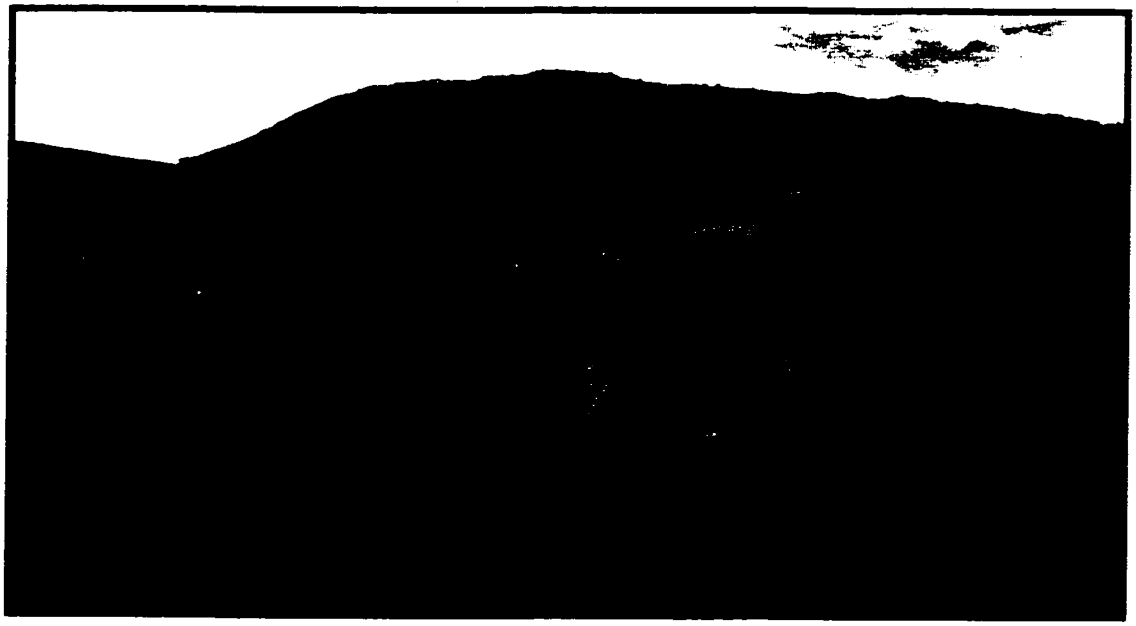


Photo looking northwards towards Sand Bay. Note that residuum conforms to the strike of the bedrock. Diabase erratic in foreground (the erratic is ~1.5m in diameter)

PLATE 3.2

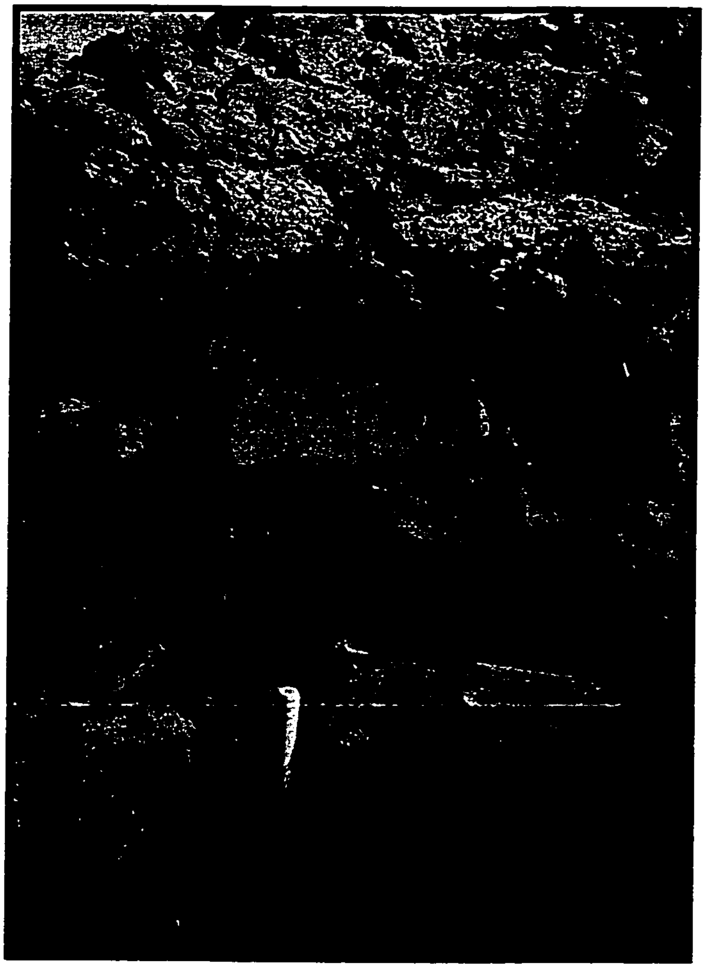
The photo is taken on the surface of the 123m marine limit delta at Good Friday Bay looking northwards towards Sand Bay(Section 3.5.2). Note the rounded to sub-rounded gravel cap of the delta in the foreground and the winding ridge of the Beaufort Formation in the background.

PLATE 3.3



Exposure of the Beaufort formation showing planar bedded sands.

PLATE 3.4



Exposure of the Beaufort Formation showing alternating fine sand and mud.

PLATE 3.5

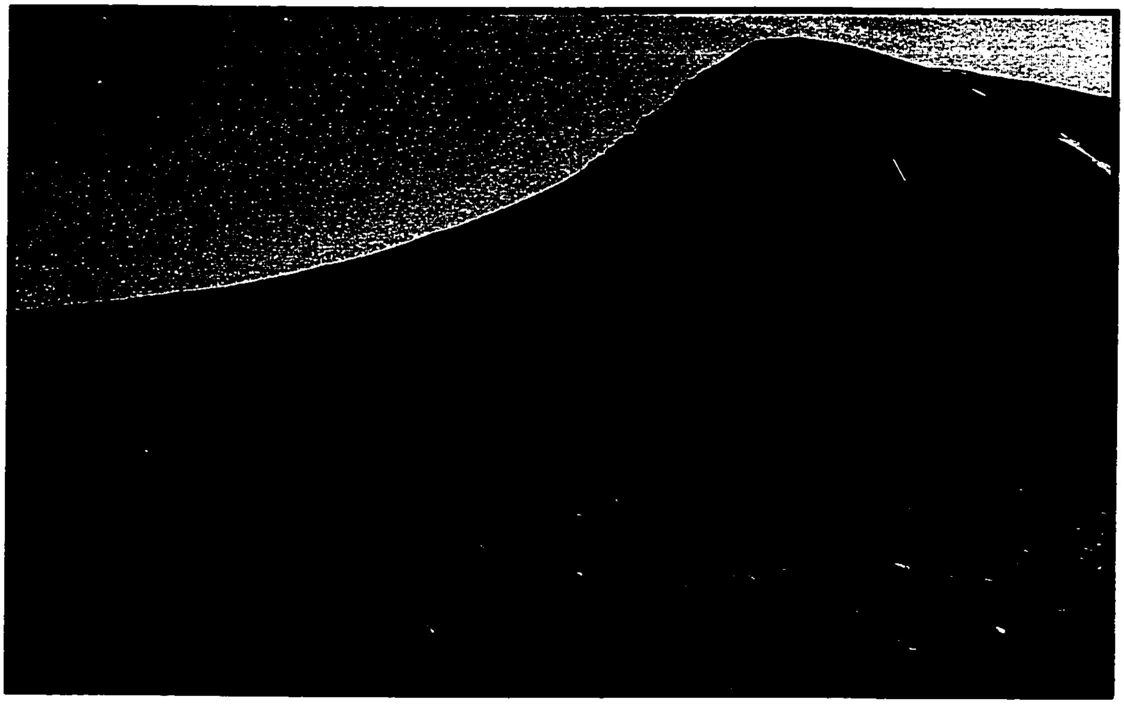
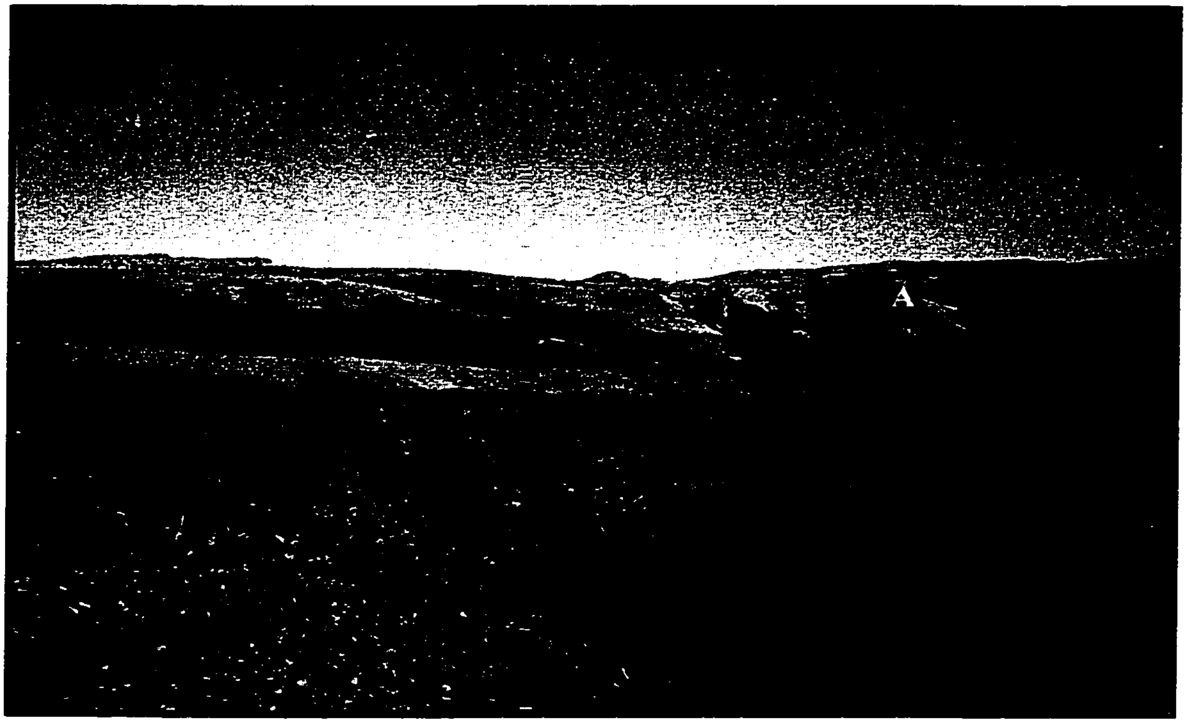


Photo taken southwards showing from the 73m ice contact delta at Mosquito Creek. Note glacifluvial gravels in the foreground and hummocky terrain in background formed by the meltout of buried ice .

PLATE 3.6

Photo looking eastward showing the sinuous ridge of rounded pebbles and cobbles at Good Friday Bay that is interpreted as a kame. Note adjacent kettle lake.

PLATE 3.7



Foreset sands at Good Friday Bay that extend to 30m asl and location of mollusc collection (A) which dates 5800±80 BP (Beta 137921, Table 4.1). Foresets grade to 45m asl. Photo taken looking northwards.

PLATE 3.8

Photo showing thermo-erosional scarps on the terrace of marine fines bordering the west side of the Amorak River at Duck Bay. Note person for scale.

CHAPTER 4: LAST GLACIAL MAXIMUM ICE COVER AND DEGLACIATION OF WEST-CENTRAL AXEL HEIBERG ISLAND.

4.1 INTRODUCTION

Air photo interpretation and field investigations are used to reconstruct the configuration of ice during the last glacial maximum (LGM) on west-central Axel Heiberg Island. Emphasis is placed on determining the subsequent pattern of deglaciation and the nature of relative sea level change.

4.2 LAST GLACIAL MAXIMUM

4.2.1 Configuration

Lateral meltwater channels crosscut the highest terrain within the study area (~760m asl) and those at the mouths of the fiords record the extension of ice offshore onto the continental shelf (A,B,C; Figure 4.1). At several sites, the lowest channels can be traced to Holocene marine limit (Figure 4.1). Collectively this distribution of meltwater channels suggests that during the last glaciation the entire topography of west-central Axel Heiberg Island was inundated by ice.

The distribution of till also records widespread glaciation on west-central Axel Heiberg Island. Till blankets and veneers are common on lowlands and valley floors and often extend to sea level providing a minimum estimate of former ice extent (Figures 3.1, 3.2, 3.3). At higher elevations (up to ~550m asl) till veneers become more sporadic. Mountain summits were not investigated and therefore the distribution of erratics at the highest elevations within the study area remains unknown.

4.2.2 Chronology

Heavily mineralised valves of *Hiatella arctica* were collected at 131m asl from the surface of till overlying an outcrop of the Beaufort Fm (Site 1, Figure 4.1; Section 3.3). The presence of molluscs within till above marine limit (123m asl) requires that the shells were glacially entrained and redeposited. The configuration of meltwater channels adjacent to the shell site indicates that the last ice that occupied this site flowed westward from Sand Bay onto the peninsula separating it from Good Friday Bay (D, Figure 4.1). An AMS date of 46 800±1200 ka BP (Lawrence Livermore, 61420; Table 4.1) was obtained on a single *H. arctica* valve, indicating that ice advanced over west-central Axel Heiberg Island after this time. However, because finite radiocarbon dates (~30-35 ka BP) have been obtained on marine molluscs of presumed Pliocene age elsewhere in the High Arctic (Bednarski 1995), these shells could be much older than 46.8 ka BP. Indeed, it is possible that the till containing these shells was preserved beneath cold-based ice during the LGM and if so, it represents an earlier glaciation of unknown age (cf. England 1976). Consequently, the chronology of ice advance on west-central Axel Heiberg Island remains undated. However, lateral meltwater channels in the study area commonly descend to Holocene marine limit marked by ice-contact and ice proximal deltas, indicating that the last ice to retreat from the study area post-dates the LGM.

4.3 DEGLACIATION

4.3.1 Configuration

The configuration of ice margins on west-central Axel Heiberg Island during deglaciation is indicated principally by lateral meltwater channels (Figure 4.1). At Mosquito Creek, meltwater channels document the eastward retreat of ice into the main (unnamed) river valley (E, Figure 4.1) followed by southeastward retreat into the highland terrain (F, G, H; Figure 4.1). Meltwater channels (I, J) and a lateral moraine (II) record upvalley retreat within the highland terrain leading to the Steacie Ice Cap in the east (Figure 4.1).

Around the coast of Duck Bay, meltwater channels depict the retreat of small valley glaciers (K, L, M; Figure 4.1). Nested meltwater channels (N, O) and moraines (III) indicate that this ice subsequently thinned on the plateau to the south of Duck Bay (Figure 4.1). On the walls of the principal valleys, nested meltwater channels indicate that larger valley glaciers also thinned and retreated towards the high terrain to the south (P,

Figure 4.1) and southeast (Q, Figure 4.1). Meltwater channels within the valleys tributary to the Amorak River similarly record ice retreat to the southeast (R, S; Figure 4.1).

Meltwater channels on the north coast of Sand Bay and on the lowland separating Sand Bay from Good Friday Bay (B, D; Figure 4.1) record the retreat of trunk ice within the fiord and into the adjacent U-shaped valley (Figure 4.1). Meltwater channels (T,U) and lateral moraines (IV) record subsequent retreat of the tributary ice towards the highlands to the east (Figure 4.1). Nested meltwater channels in this highland also record the progressive thinning of ice to the east (V, Figure 4.1).

A till blanket at the mouth of the main valley entering the lowlands of Good Friday Bay exhibits northeast-southwest oriented flutings (0.5km^2; VI, Figure 4.1). These flutings indicate the westward deflection of ice exiting the valley, suggesting its coalescence with trunk ice in Good Friday Bay. Along the north coast of Good Friday Bay, fiord-parallel meltwater channels show that trunk ice occupied the outer fiord (C, W; Figure 4.1) and at one site, a channel terminates at a small fan that probably marks marine limit (VII; Figure 4.1). Inland, nested meltwater channels (X,Y), a lateral moraine (VIII), and a kame (IX) delineate margins of a tributary valley glacier after it became detached from trunk ice in the fiord (Figure 4.1). A morainal bank (XI, Figure 4.1) and an ice-contact delta (Site 8, Figure 4.1) flank the east side of the main valley at Good Friday Bay. On the west side, a moraine is superimposed on the surface of a till blanket (XII, Figure 4.1). The abundance of glacial landforms in this area likely represents a temporary stillstand of the ice front during retreat as it became stabilised at the narrow mouth of the valley. If the pattern of ice retreat was the reverse of the advance pattern (cf. Bednarski 1998) then west-central Axel Heiberg Island was probably glaciated by the northward and westward expansion of the Steacie Ice Cap which lies to the east of the study area (Figure 1.5).

4.3.2 Deglacial chronology and marine limit.

At several sites within the study area, ice marginal meltwater channels descend to marine limit which is marked by ice-contact or ice-fed deltas. This indicates that ice retreat and marine transgression were concurrent. Therefore, dating marine limit establishes the chronology of deglaciation. Unfortunately, molluscs were often absent in raised marine sediments, limiting the number of dateable samples. Lemmen *et al.* (1994) previously attributed the scarcity of shells in this area to low original numbers of colonising species or to dissolution or crushing. In most cases, molluscs were confined to surface collections that occurred below marine limit and hence have an uncertain association with it. Therefore, all dates are treated as minimum estimates on deglaciation and the establishment of marine limit. Five radiocarbon dates of Holocene age on marine molluscs have been determined from the field area and these provide a preliminary chronology of deglaciation for west-central Axel Heiberg Island.

4.3.2(a) Duck Bay-Mosquito Creek (Outer Strand Bay)

At Mosquito Creek, an ice-contact delta marks marine limit at 73m asl (Site 2, Figure 4.1). The delta is flanked by widespread hummocky terrain and large ice-wedge polygons suggesting the meltout of buried ice (Sections 3.4.1; 3.5.1). The delta is non-fossiliferous. However, molluscs were collected ~10km seaward of the delta at 40m asl from finely laminated, stony marine mud (Site 3, Figure 4.1). These sediments are interpreted to have been deposited by suspension settling and ice-rafting in an ice-proximal environment (Section 3.6.1). An AMS date of 8370±50 BP (Lawrence Livermore 61416; Table 4.1) on a single valve of *H. arctica* provides a minimum estimate for the deglaciation of outer Strand Bay. The ice-contact delta (73m asl) 10km upvalley is likely younger than 8370±50 BP as it is well below local marine limits (~105m asl) that date close to this age (Lemmen *et al.* 1994).

A non-fossiliferous delta composed of truncated foreset sand was located at 68m asl on the SE shore of Duck Bay (Site 4, Figure 4.1). As topsets are absent, the elevation of the foresets must be a minimum estimate for relative sea level at the time of delta deposition. Whole valves of *Astarte borealis* and *Mya truncata* were collected at 54m asl from the surface of marine mud ~3km to the east of the delta (Site 5, Figure 4.1). A sample from this collection dated 7780±70 BP (Beta 137919; Table 4.1) and provides a minimum estimate on the timing of deglaciation at Duck Bay but cannot confidently be associated with a former relative sea level.

4.3.2(b) Sand Bay

A delta at 123m asl records marine limit at Sand Bay (Site 6, Figure 4.1). Whole valves of *A. borealis* and *H. arctica* collected at 10m asl from silt at this site date 4700±70 BP (Beta 137923; Table 4.1). The delta surface at 123m asl is equal to the highest marine limit recorded in the field area. The date of 4.7 ka BP at Sand Bay is ~4 ka younger than estimated for a similar marine limit to the south (see below) and must therefore date a relative sea level well below marine limit.

4.3.2(c) Good Friday Bay

Marine limit at Good Friday Bay is marked by a delta at 123m asl (Site 7, Figure 4.1). The delta lacks a modern sediment source and its foreset beds contain dropstones suggesting deposition in an ice-proximal environment (Section 3.6.2). Approximately 12km to the east, a meltwater channel descends to a prominent ice-contact delta at 120m asl (Site 8, Figure 4.1). The delta is adjacent to a morainal bank (XI, Figure 4.1) and these landforms are regarded to be contemporaneous and may have been formed during a period of ice-marginal stability at the mouth of the main tributary valley entering Good Friday Bay (Section 4.3.1). A third delta located within this main valley (Site 9, Figure 4.1) is composed of truncated foreset sands at 119m asl. This delta provides a minimum estimate for marine limit which varies locally from ≥119-≤123m asl. The three marine limit sites at Good Friday Bay are closely accordant (± 4m) and are consequently considered to be of similar age.

Five kilometres to the southwest of the 123m asl delta whole valves of *H. arctica* were collected at 96m asl from the surface of silty mud (Site 10, Figure 4.1). The molluscs were dated at 8820±60 BP (Beta 137922; Table 4.1) and provide a minimum age estimate for marine limit along the outer north shore of Good Friday Bay. A second sample of molluscs (whole valves of *H. arctica* and *M. truncata*) was collected at 79m asl from the surface of clast-rich silty mud nearer the 123m asl delta (Site 11, Figure 4.1). This sample was slightly younger and dated 8690±70 BP (Beta 137920; Table 4.1) and its standard error overlaps with the previous date. Therefore, the date of 8820±60 BP is taken as the best estimate on marine limit. This is also the oldest postglacial radiocarbon date recovered from the study area and it provides a first estimate on the deglaciation of the outer west-central coast of Axel Heiberg Island.

Topset and foreset sand form a prominent delta at an intermediate relative sea level (45m asl) in the lowlands of Good Friday Bay (Site 12, Figure 4.1). Paired valves of *H. arctica*, *M. truncata* and *A. borealis* were collected in growth position from the foreset beds at an elevation of 30m asl. The shells dated 5800±80 BP (Beta 137921; Table 4.1) and are assumed to date the 45m asl shoreline.

4.4 DISCUSSION

4.4.1 Sediments, landforms and basal thermal regime

In the High Arctic, the distribution of sediments and landforms deposited by glaciers and ice caps has been used to draw inferences on former basal thermal regimes (cf. Dyke 1993; Ó Cofaigh *et al.* 1999; England *et al.* in press). The distribution of glacial sediments and landforms on west-central Axel Heiberg Island can be similarly utilised. Widespread Holocene glacimarine sediment shows that a large volume of debris was delivered to former ice margins in the study area. This suggests that subglacial meltwater passed through debris-rich basal ice during deglaciation and hence indicates a warm-based thermal regime (cf. Aitken & Bell 1998; Ó Cofaigh *et al.* 1999). At Good Friday Bay, a small field of flutings (0.5km^2; VI, Figure 4.1) record warm-based ice exiting the tributary valley to the east. Approximately 2.5km up-valley, a morainal bank (XI, Figure 4.1) and an ice-contact delta (Site 8, Figure 4.1) document the stabilisation of this ice as it retreated towards the interior highlands. A morainal bank is a subaqueous landform created by the deposition of sediment from subglacial meltwater discharge (cf. Ó Cofaigh *et al.* 1999). Consequently, its presence indicates that the ice occupying the valley was warm-based, at least during deglaciation.

Unaltered residuum (Figures 3.1, 3.2, 3.3) and preglacial drainage channels (Section 2.2.1) are conspicuous within the study area. They indicate that the majority of west-central Axel Heiberg Island was unmodified during the last glaciation and suggest that it was crossed by ice that was largely protective and hence cold-based (Dyke 1993). The most widespread and conspicuous ice-marginal landforms within these areas are lateral meltwater channels (Figure 4.1). Cold-based thermal regimes may have existed where the duration of ice cover was insufficient to dissipate pre-existing permafrost by the input of geothermal heat and strain heat produced by ice flow (Dyke 1993). Cold-based thermal regimes may also have been favoured in valleys due to the advection of cold ice from upstream (highland) locations (cf. Blatter 1987) or on plateaux where ice cover was too thin to insulate the bed from cold atmospheric temperatures.

Till containing striated and faceted clasts provides evidence for warm-based ice throughout the study area (Figures 3.1, 3.2, 3.3). Thick till blankets are particularly common on the floors of larger valleys where ice thickness would have been greatest. In these areas, geothermal heat and strain heating would have been used to melt basal ice following the removal of any underlying permafrost (Dyke 1993). The abundance of till blankets in the main valleys of the study area may reflect the glacial redistribution of pre-existing sediment (cf. Whillans 1978) that would be expected to be abundant in pre-glacial valleys. The chronology of till deposition on west-central Axel Heiberg Island is unknown. Till in the study area could have been deposited during the LGM or may have been protected beneath cold-based ice and therefore pertain to previous glaciations. Indeed, the presence of late Tertiary Beaufort Formation sediments (Section 3.3) indicates that surfaces of considerable antiquity have been preserved within the study area.

The distribution of glacial landforms and sediments suggests the presence of polythermal ice on west-central Axel Heiberg Island, at least during deglaciation. The present landscape is a palimpsest of glacially modified and unmodified terrain with an unknown temporal resolution. Similar landscape configurations have been reported from the Canadian High Arctic (cf. Dyke 1983,1993; England 1986) and from Scandinavia (cf. Kleman 1994) where they have been attributed to variable basal thermal regimes during former glaciations.

4.4.2 Marine limit

Marine limit on the outer coast of west-central Axel Heiberg Island at Good Friday Bay and Sand Bay is ~123m asl whereas the ice-contact delta at Mosquito Creek (73m asl) is the lowest marine limit reported from Axel Heiberg Island. Because these discordant marine limits (123 vs. 73m asl) occur in the same locality, they cannot be explained by different ice load histories. Hence, they must reflect differing ages. The age of marine limit at Good Friday Bay is ~8.8 ka BP whereas at Mosquito Creek the age is unknown, but younger than 8.4 ka BP. Consequently, the earlier deglaciation of Good Friday Bay would have resulted in a longer period of postglacial rebound thus explaining its higher marine limit. The low marine limit (73m asl) is best explained by late deglaciation at this site as ice retreated into the local highlands, indicated by the pattern of meltwater channels (Figure 4.1).

4.5 REGIONAL SYNTHESIS

4.5.1 Regional Isobases

Research on west-central Axel Heiberg Island augments the regional pattern of postglacial sea level change by providing data from the little studied western margin of the Queen Elizabeth Islands. The pattern of differential postglacial emergence over the study area can be investigated by comparing relative sea level at 8.5 ka BP from two sites on western Axel Heiberg Island: Expedition Fiord and Good Friday Bay (Figure 4.2). The 8.5 ka BP shoreline is of interest to this study as it can be used to build upon a well-established database of postglacial shorelines at 8.5 ka BP that are mapped by Ó Cofaigh (1999). At the head of Expedition Fiord, to the northeast of the study area, marine limit is 102m asl and dates 8.4 ka BP (Lemmen *et. al* 1994). Using the emergence curve published by Lemmen *et al.* (1994), the initial rate of emergence at the head of the fiord is estimated at ~3m/century. Therefore, relative sea level at 8.5 ka BP at the head of Expedition Fiord is calculated to be 105m asl (marine limit at 8.4 ka BP) plus 3m (3m of emergence in the previous 100 years) or 105m asl (A, Figure 4.2). This assumes that the final interval of restrained rebound (Andrews 1970) occurring beneath the retreating ice was similar to the initial rate of postglacial emergence (~3m/100yrs) following the entry of the sea. Relative sea level at 8.5 ka BP at the head of Strand Fiord, adjacent to Expedition Fiord (B, Figure 4.2), can be similarly calculated at 127m asl (marine limit 124m asl at 8.4 ka BP plus 3m of emergence in the previous 100 years). Marine limit at outer Good Friday Bay is 123m at 8.8 ka BP. An estimate of the 8.5 ka BP shoreline at Good Friday Bay can be made by applying the initial rate of emergence determined for Expedition Fiord to Good Friday Bay. The 8.5 ka BP shoreline is subsequently calculated at 123m (marine limit at 8.8 ka BP) minus 9m ((3x3m of emergence per 100 years) or 114m asl (C, Figure 4.2). The calculated values of the 8.5 ka BP shoreline can also be determined in calendar years which indicates that the initial rate of emergence is ~3.2m/century. This calculation does not significantly alter the estimates on the 8.5 ka BP shoreline presented above.

Isobases drawn on the 8.5 ka BP shoreline by Ó Cofaigh (1999; Figure 4.3) show a closed cell of maximum emergence aligned north-south along Eureka Sound, corresponding to the central loading axis of the IIS (Blake 1970, 1975; Walcott 1972). The 8.5 ka BP isobase that encloses Eureka Sound reaches 130m asl whereas along the outer west-central coast of Axel Heiberg Island the isobases are speculated to decline to ~110m asl (Ó Cofaigh 1999; Figure 4.3). The 8.5 ka BP sea level over western Axel Heiberg Island calculated in this study suggests that the isobases drawn by Ó Cofaigh should be repositioned farther west and reoriented northeast to southwest (Figure 4.2). The coastward deflection of the 8.5 ka BP isobases is consistent with recent isobases presented by Dyke (1998) that indicate a northeast-southwest ridge of postglacial emergence centred to the northwest of Devon Island. This ridge corresponds to a former loading centre of the IIS situated within the central Arctic (Dyke 1998, Figure 4.3). Isobases to the north of the loading centre are oriented parallel to it and hence northeast-southwest across west-central Axel Heiberg Island (Figure 4.3). The pattern of postglacial emergence proposed by this study appears to extend the Axel Heiberg Isobases towards Amund Ringnes and Cornwall islands. On the north coast of Cornwall Island, marine limit is recorded at 115m asl and dates 8.5 ka BP (Lamoureux & England in press). Hence, the 8.5 ka BP isobase over outer Good Friday Bay (115m asl; Figure

4.2) extends to Cornwall Island. It is unknown whether the 120m isobase at 8.5 ka BP forms a closed cell in the vicinity of Norwegian Bay or whether it extends farther into the western Arctic islands.

4.5.2 Last Glacial Maximum

In their reconstruction of the ice configuration in Strand Fiord during the LGM, Lemmen *et al.* (1994) suggested that valley glaciers advanced northwards from the Steacie Ice Cap and terminated at tidewater along the south shore of the fiord. In contrast, fieldwork carried out in this study suggests that during the LGM, the Steacie Ice Cap expanded farther to the north and west, inundated the topography of west-central Axel Heiberg Island and extended an unknown distance offshore. The absolute chronology of this ice advance remains undetermined. In this reconstruction, the glacial history of Strand Fiord is more in line with that of Expedition Fiord where lateral moraines demarcate the margin of a trunk glacier that extended at least as far as Middle Islands, ~20km from the head of the fiord (Lemmen *et al.* 1991, 1994). Furthermore, Lemmen *et al.* (1991, 1994) stated that there was no unequivocal evidence to suggest that Middle Islands marked the *limit* of the last glaciation. Indeed, they report that the outer fiord contains glacial landforms and deposits recording a more extensive ice cover of unknown age. Lemmen *et al.* (1994) also proposed that if Middle Islands do mark the limit of the last glaciation then a greater thickness of sediment should have accumulated on the fiord floor distal to the islands because it would have been deglaciated for a longer interval, presumably throughout the last glaciation. However, Lemmen *et al.* (1994) recorded similar sediment thickness in the inner and outer fiord distal to Middle Islands. This can be reconciled by a more advanced ice cover during the last glaciation that would have resulted in a similar interval of sedimentation throughout the fiord following deglaciation.

The extent of ice advance onto the continental shelf of west-central Axel Heiberg Island during the LGM remains unknown. In the interisland channels of the southwestern QEI that are known to have been glaciated during the LGM (cf. Dyke 1998), MacLean *et al.* (1989) consistently record the presence of bedrock overlain by till which is, in turn, overlain by Holocene glacimarine and postglacial marine sediments. Expansion of ice onto the continental shelf of west-central Axel Heiberg Island should have produced a similar stratigraphy. To date, only sedimentary cores from ~40km off northwestern Axel Heiberg Island (~200km from the study area) have been investigated (Hein *et al.* 1990; Hein and Mudie 1991). These authors report the presence of fine-grained Holocene marine deposits overlying bedrock and consequently propose that grounded ice did not occupy this part of the continental shelf during the last glaciation. However, an absence of ice cover off the northwestern tip of Axel Heiberg Island does not preclude the possibility of ice advance onto the shelf farther to the south. Furthermore, on the basis of similar sediments recovered from the Barents Sea, L. Polyack (pers. comm. to J. England 1999) suggests that some of the facies described by Hein *et al.* (1990) and Hein and Mudie (1991) may be reinterpreted as till. Such a conclusion would invoke grounded ice on the continental shelf. Due to the differing interpretations regarding former ice extent that may be inferred from the existing marine sedimentary record, further investigation of these sediments and their genesis is warranted.

4.5.3 Deglaciation

Regional deglaciation of the IIS was underway in the westernmost margin of the QEI at Prince Patrick Island by ~11 ka BP (Blake 1972) and reached the central axis of the ice sheet within Eureka Sound by ~9 ka BP (Bell 1996; Hodgson 1985; Ó Cofaigh 1999). This study suggests that west-central Axel Heiberg Island was deglaciated by ~8.8 ka BP which accords with the regional pattern of ice retreat.

A comparison of calibrated radiocarbon dates on deglaciation at the head of Strand Fiord (8430±80 ka BP (GSC-5411), 8924 cal BP; Lemmen *et al.* 1994) and at the outer west coast of Axel Heiberg Island (8820±60 ka BP (Beta 137922), 9197 cal BP; this study) indicates that ice retreated from the coast to the fiord head, a distance of ~52 km in 273 years (190m/yr). The retreat of fiord-based glaciers in the QEI has been widely attributed to calving at tidewater margins facilitated by increasing summer temperatures (particularly between ~9.5-8.5 cal BP; Koerner & Fisher 1990) in conjunction with an ongoing rise in eustatic sea level in the early Holocene (Fairbanks 1989). These factors combined to promote ice retreat by favouring the calving of marine-based ice fronts (cf. Ó Cofaigh 1999).

TABLE 4.1 RADIOCARBON DATES, WEST-CENTRAL AXEL HEIBERG ISLAND

Site	Location	Laboratory Dating No.	Material	Age (years BP) Conventional	AMS	Age *** (calibrated years BP)	Enclosing material	Sample Elevation (m asl)	Related RSL (m asl)
1	78° 43'N 93° 07'W	Lawrence Livermore* 61420	*H. arctica*		46800±1200	NA	Surface (till)	131	NA
3	79° 01'N 93 °52'W	Lawrence Livermore 61416	*H. arctica* *A. borealis*		8370±50	8890	Surface (stony marine mud)	40	≥40-≤73
5	79° 08'N 92° 36'W	Beta** 137919	*A. borealis* *M. truncata*	7780±70		8217	Surface (laminated marine mud)	54	>54
6	78° 48'N 93 °17'W	Beta 137923	*A. borealis*	4700±70		4889	Silt	10	≥10 <123
10	78° 38'N 93° 10W	Beta 137922	*H. arctica*	8820±60		9197	Surface (stony marine mud)	96	≥96-≤123
11	78° 39'N 93° 04W	Beta 137920	*A. borealis* *M. truncata*	8690±70		9061	Surface (stony marine mud)	79	≥79-≤123
12	78° 38'N 92° 56'W	Beta 137921	*H. arctica* *A. borealis* *M. truncata*	5800±80		6208	Foreset sand	30	45

*Radiocarbon dating laboratory: Lawrence Livermore, California; ** Radiocarbon dating laboratory: Beta Analytic Inc., Miami; ***All dates were calibrated using CALIB 4.2 (Stuiver & Reimer 2000); All radiocarbon dates were calculated using ^{13}C= -25 ‰ and subsequently corrected for a marine reservoir age of 410 years.

FIGURE 4.1: RADIOCARBON DATES, MARINE LIMIT ELEVATIONS AND GLACIAL GEOMORPHOLOGY, WEST-CENTRAL AXEL HEIBERG ISLAND.

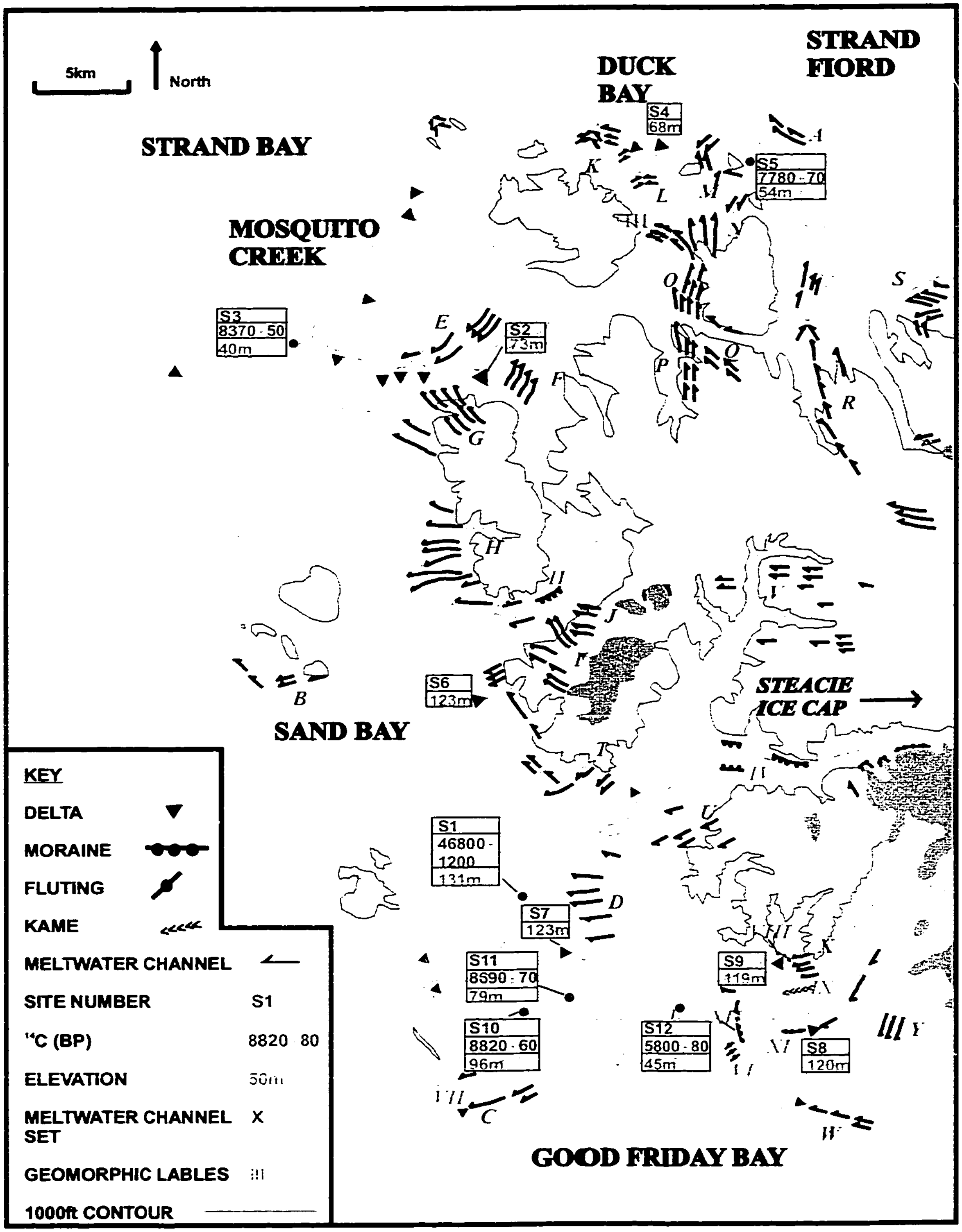


FIGURE 4.2: RELATIVE SEA LEVEL AND ISOBASES AT 8.5 ka BP, WEST-CENTRAL AXEL HEIBERG ISLAND.

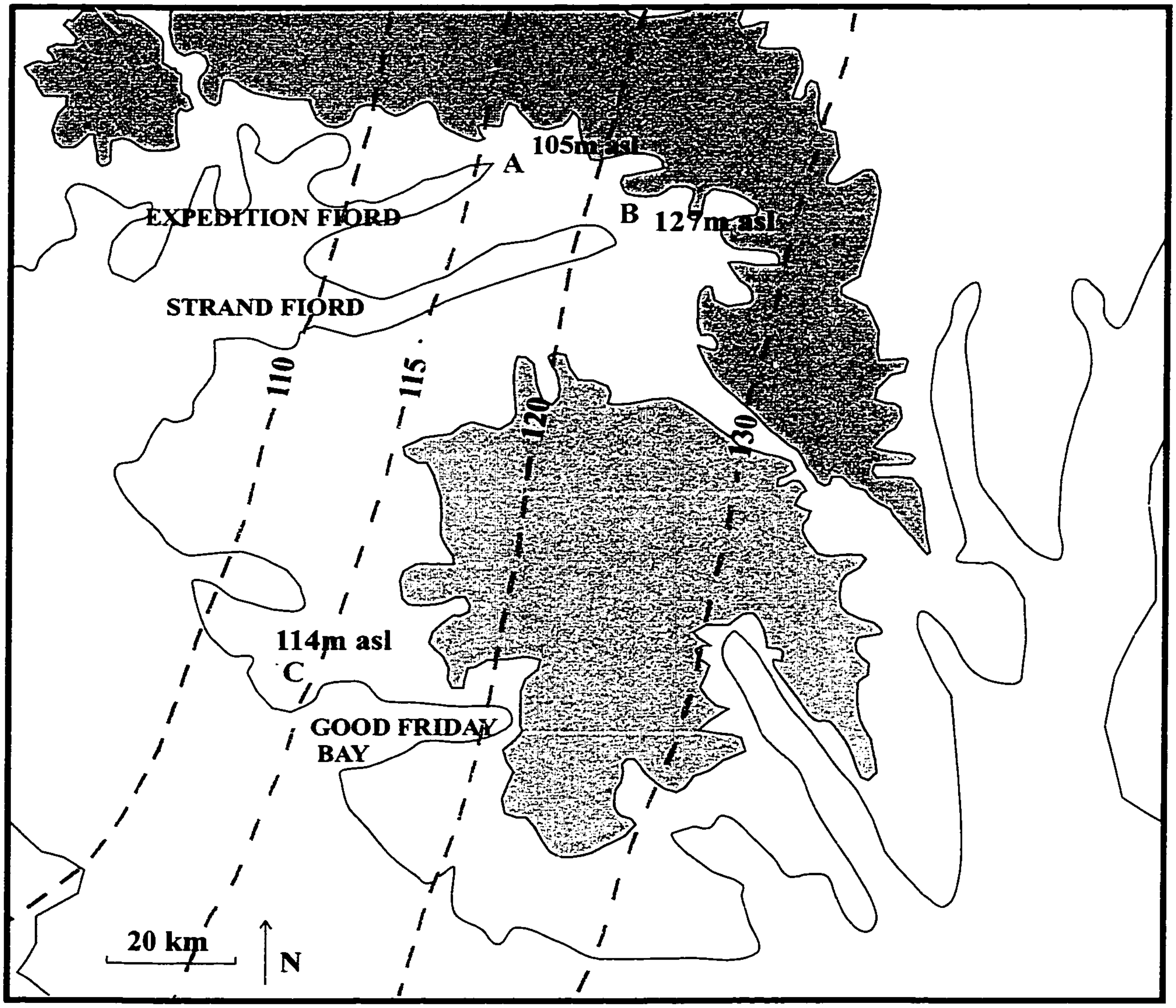
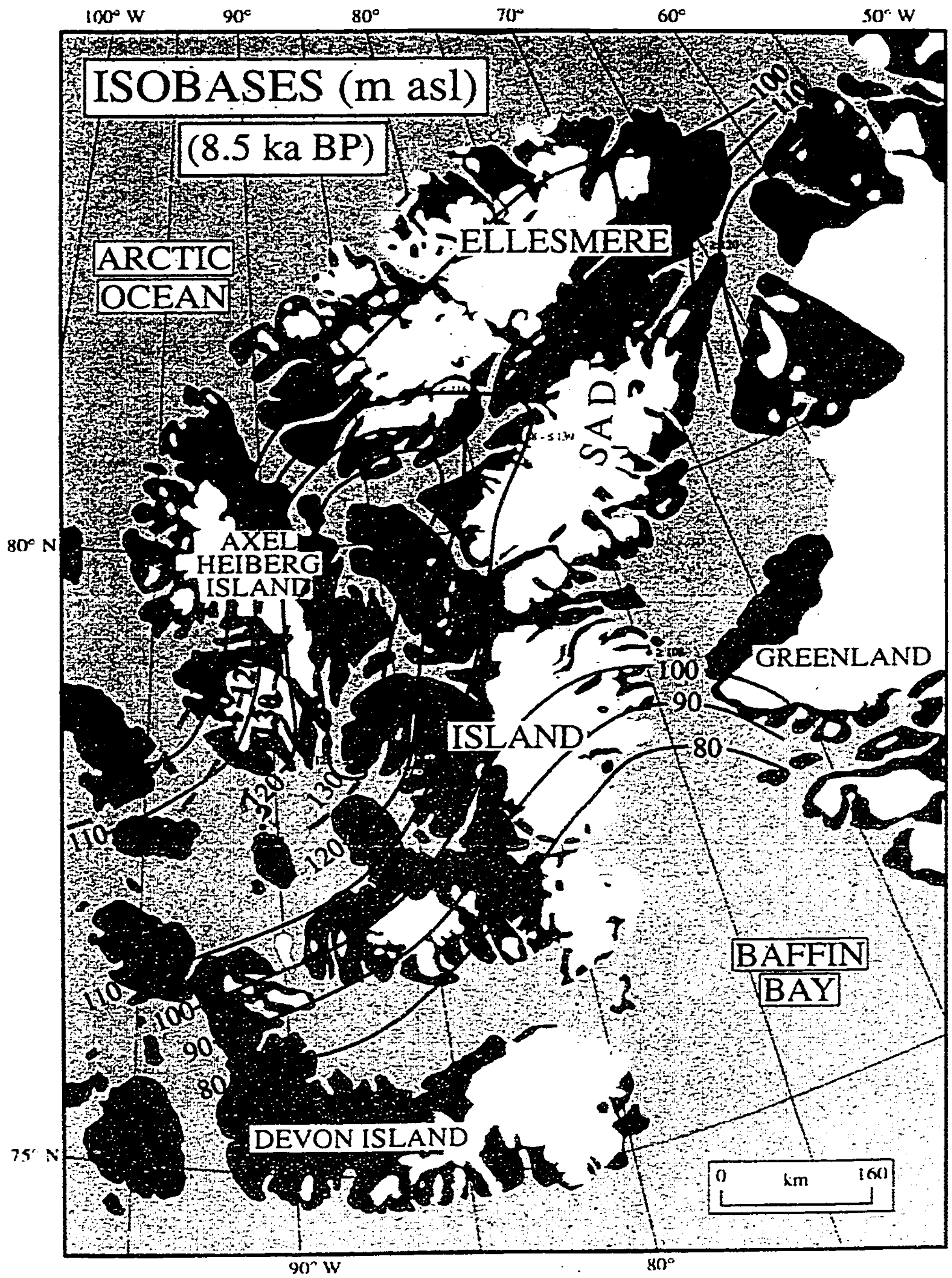


FIGURE 4.3: POSTGLACIAL ISOBASES AT 8.5 ka BP (Adapted from England 1999)



KEY

——— 8.5 ka BP ISOBASES (O'COFAIGH 1999) 1 Amund Ringnes Island
——— 8.5 ka BP ISOBASES (THIS STUDY) 2 Cornwall Island

CHAPTER 5: CONCLUSIONS

5.1 INTRODUCTION

Prior to this study, west-central Axel Heiberg Island constituted one of the uninvestigated sectors of the QEI and hence warranted investigation in order to augment the regional examination of late Quaternary glacial and sea level history. The objectives of the research were:

1) To determine the configuration, origin and chronology of former glaciation on west-central Axel Heiberg Island.
2) To determine the pattern and chronology of deglaciation and the nature of relative sea level adjustments on west-central Axel Heiberg Island.

5.2 CONFIGURATION, ORIGIN AND CHRONOLOGY OF GLACIATION

During the last glaciation, west-central Axel Heiberg Island was inundated by the expansion of local alpine ice masses. The distribution of glacial landforms and sediments within the study area indicates that the Steacie Ice Cap advanced to the north and west, engulfing the topography of west-central Axel Heiberg Island and extending an unknown distance offshore. Ice-transported shells from a till veneer at Good Friday Bay were AMS dated at 46 800±1200 ka BP (Lawrence Livermore 61420; Table 4.1). However, the molluscs may be older than this date. Therefore, the absolute chronology of ice advance over west-central Axel Heiberg Island remains unknown. Lateral meltwater channels in the study area commonly descend to Holocene marine limit indicating that the last ice to retreat from the study area post-dates the LGM.

5.3 PATTERN AND CHRONOLOGY OF DEGLACIATION AND RELATIVE SEA LEVEL ADJUSTMENTS

5.3.1 Pattern and chronology of deglaciation

Meltwater channels provide the principal evidence for the configuration of ice as it retreated from west-central Axel Heiberg Island. Nested meltwater channels record the thinning of fiord-based trunk ice and the retreat of terrestrially based ice from the main valleys of the study area towards the interior highlands. Whole valves of *H. arctica* collected from silty mud at Good Friday Bay dated 8820±60 BP (Beta 137922, Table 4.1) and provide a minimum age estimate for the onset of deglaciation. The deglacial chronology proposed for west-central Axel Heiberg Island is in accord with the regional retreat of the IIS from its westernmost margin (~ 11 ka BP; Blake 1972) to its former centre in Eureka Sound (~ 9 ka BP; Bell 1996; Hodgson 1985; Ó Cofaigh 1999).

5.3.2 Relative sea level adjustments

The nature of differential postglacial emergence over west-central Axel Heiberg Island is indicated by the distribution of the 8.5 ka BP isobases. The isobases trend northeast to southwest over the study area and suggest that the north-south trajectory of the 8.5 ka BP isobases for western Axel Heiberg Island presented by Ó Cofaigh (1999) should be revised. The repositioning and reorientation proposed by this study deflects the western Axel Heiberg Island 8.5 ka BP isobases farther west towards Amund Ringnes and Cornwall islands. This accords with a recently documented uplift centre of the IIS which displays the same orientation (Dyke 1998).

5.4 RECOMMENDATIONS FOR FUTURE RESEARCH

1) An extensive ice cover over west-central Axel Heiberg Island during the LGM, as proposed by this study, accords with widespread evidence for the IIS elsewhere in the QEI (cf. Blake 1970, 1975; Bednarski 1998; Dyke 1998, 1999; England 1998a, 1999; Ó Cofaigh 1999). During the LGM, Innuitian ice converged into Eureka Sound and Norwegian Bay (England 1998b; Dyke 1998; Ó Cofaigh 1999) where an ice stream may have extended northwestward through Massey Channel (J. England pers. comm. 1999). The expansion of local alpine ice masses on west-central Axel Heiberg Island could have prevented the onshore encroachment of this ice. Further studies should be directed towards clarifying whether or not an ice stream occupied Massey Sound and, if so, determining the relationship between this ice and local ice exiting the west coast of Axel Heiberg Island. To this end, future fieldwork could address the presence/absence of far-travelled erratics above marine limit. Small granite erratics have been observed along the west coast of Axel Heiberg Island (J. England & A. Dyke, pers. comm. 1999) and a more thorough investigation is warranted.

2) This study has presented a preliminary chronology for the glaciation and deglaciation of west-central Axel Heiberg Island. Further sampling and dating of fossiliferous till and raised marine sediments would provide greater detail on the timing of ice advance, retreat and postglacial emergence. Additional surveys of raised marine sediments would enable the determination of a postglacial emergence curve for west-central Axel Heiberg Island. Specifically, dated sea levels below 45m asl are unavailable. Hence, emphasis should be placed on the study of lower elevation marine landforms that can be radiocarbon dated. This would allow the form of the postglacial emergence curve to be established and compared to the regional pattern of response times presented by Dyke and Peltier (2000).

3) At present, the maximum extent of ice on the continental shelf of western Axel Heiberg Island is unknown. Further investigation of the marine sedimentary record may clarify the extent, origin and chronology of former glaciations from western Axel Heiberg Island that are not apparent within the terrestrial sedimentary record.

4) Regionally, the extent, configuration and dynamics of the IIS over northwestern Axel Heiberg Island and over the westernmost Arctic islands remains unknown and future Quaternary research should be directed towards these uninvestigated areas.

5) The IIS had a large marine-based component and was located adjacent to the Arctic Ocean. Therefore, glacial advance and retreat should be recorded within the sedimentary record of the Arctic Ocean Basin. A detailed comparison of these sediments and those on nearby terrestrial settings may allow a more thorough documentation of ice advance and retreat. In addition, it may enable the sources and transport pathways of sediment entering the Arctic Ocean to be determined (cf. Bischof & Darby 1999).

BIBLIOGRAPHY

Aitken, A.E., Bell, T. 1998. Holocene glacimarine sedimentation and macrofossil paleoecology in the Canadian High Arctic: environmental controls. *Marine Geology*, **145**, 151-171.

Andrews, J.T. 1970. A geomorphological study of post-glacial uplift with particular reference to Arctic Canada, *Institute of British Geographers*, Special Publication Number 2.

Atkinson, N. 1999. The last glaciation and relative sea level history of central Baumann Fiord, southwest Ellesmere Island, Canadian High Arctic, *Unpublished MSc Thesis*, University of Alberta.

Bednarski, J. 1995. Glacial advances and stratigraphy in Otto Fiord and adjacent areas, Ellesmere Island, Northwest Territories. *Canadian Journal of Earth Sciences*, **32**, 52-64.

Bednarski, J. 1998. Quaternary history of northeastern Axel Heiberg Island, bordering Nansen Sound, *Canadian Journal of Earth Sciences*, **35**, 520-533.

Bell, T. 1992. Glacial and sea level history of western Fosheim Peninsula, Ellesmere Island, Canadian High Arctic, *Unpublished Ph.D. Thesis*, University of Alberta, Edmonton.

Bell, T. 1996. Late Quaternary glacial and sea level history of Fosheim Peninsula, Ellesmere Island, Canadian High Arctic, *Canadian Journal of Earth Sciences*, **33**, 1075-1086.

Benn, D.I., Evans, J.A., 1998. Glaciers and Glaciation. *Arnold*, London, pp734.

Beschel, R.E. 1961. Dating rock surfaces by lichen growth and its application to glaciology and physiography (lichenometry) in Raasch, G.O., ed. *Geology of the arctic.* Toronto, University of Toronto Press, 1961, Vol. 2, 1044-62.

Blake, W.R.Jr. 1970. Studies of glacial history in Arctic Canada. 1. Pumice, radiocarbon dates, and differential postglacial uplift in the eastern Queen Elizabeth Islands. *Canadian Journal of Earth Sciences*, **7**, 634-664.

Blake, W.R.Jr. 1972. Climatic implications for radiocarbon-dated driftwood in the Queen Elizabeth Islands, arctic Canada in Hyvarinen, Y, and Hicks S. (eds.), *Climatic Changes in Arctic areas during the Past Ten – Thousand Years*, 77-104. Acta Universitatis Ouluensis, Ser. A. Scientifae Rerum Naturalum, No 3, Geologica No. 1

Blake, W.R.Jr. 1975. Radiocarbon age determination and postglacial emergence at Cape Hershel, east-central Ellesmere Island, *Geofriska Annaler*, **57A**, 1-71.

Blake, W.R.Jr.1977. Glacial sculpture along the east central coast of Ellesmere Island, Arctic Archipelago. *Geological Survey of Canada Paper*, **77-A**, 107-115.

Blake, W.R.Jr. 1992a. Holocene emergence at Cape Herschel, east-central Ellesmere Island, Arctic Canada: implications for ice sheet configuration, *Canadian Journal of Earth Sciences*, **29**, 1958-1980.

Blake, W.R.Jr.1992b. Shell-bearing till along Smith Sound, Ellesmere Island - Greenland: Age and significance. *Sveriges Geologiska Undersokning*, **81**, 51-58.

Blatter, H.B., 1987. On the thermal regime of am Arctic valley glacier: a study of White Glacier, Axel Heiberg Island, N.W.T., Canada. *Journal of Glaciology*, **33**, 200-209.

Boesch, H. 1963. Notes on the geomorphological history in: Preliminary report 1961-1962, McGill University, Montreal, *Axel Heiberg Island Research Reports*, 163-168.

Boulton, G.S. 1979. A model of Weichselian glacier variation in the north Atlantic regions, *Boreas*, **8**, 373-395.

Bradley, R.S. 1990. Holocene paleoclimatology of the Queen Elizabeth Islands, Canadian High Arctic. *Quaternary Science Reviews*, **9**, 365-384.

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

Bustin, R. M. 1982. Beaufort Formation, Eastern Axel Heiberg Island, Canadian Arctic Archipelago. *Bulletin Canadian Petroleum Geology*, **30**, 140-149.

Cogley, G.J., Adams, W.P., Ecclestone, M.A., Jung-Rothenhauser, F., Ommanney C.S.L. 1996. Mass balance of White Glacier, Axel Heiberg Island, N.W.T., Canada, 1960-91. *Journal of Glaciology*, **42**, 548-563.

Dawes, P.R. and Christie, R. L. 1991. Geomorphic Regions; Chapter 3 in H.P. Trettin (ed.) *Geology of the Innuitian Orogen and Arctic Platform of Canada and Greenland*, Geological Survey of Canada, Geology of Canada, no. 3.

Dyke, A.S., 1983. Quaternary Geology of Somerset Island, District of Franklin. *Geological Survey of Canada*, Memoir 404.

Dyke, A.S., 1993. Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada, *Progress in Physical Geography*, **17**, 223-247.

Dyke, A.S. 1998. Holocene delevelling of Devon Island, Arctic Canada: Implications for ice sheet geometry and crustal response, *Canadian Journal of Earth Sciences*, **35**, 885-904.

Dyke, A.S. 1999. Last Glacial Maximum and deglaciation of Devon Island, Arctic Canada: support for an Innuitian Ice Sheet, *Quaternary Science Reviews*, **18**, 393-420.

Dyke, A.S., Morris, T.F., Green, D.E.C. 1991. Postglacial Tectonic and Sea Level History of the Central Canadian Arctic, *Geological Survey of Canada*, Bulletin 397.

Dyke, A.S., Morris, T.F., Green, D.E.C., England J. 1992. Quaternary geology of Prince of Wales Island, Arctic Canada, *Geological Survey of Canada*, Memoir 433.

Dyke, A.S., W.R. Peltier. 2000. Forms, response times and variability of relative sea level curves, glaciated North America, Geomorphology, **32**, 315-333.

Edlund, S.A., Alt, B.T. 1988. Regional Congruence of Vegetation and Summer Climate Patterns in the Queen Elizabeth Islands, Northwest Territories, Canada. *Arctic*, **42**, 3-23.

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

England, J.1987. Glaciation and the evolution of the Canadian High Arctic landscape, *Geology*, **15**, pp. 419-424.

England, J. 1992 Postglacial emergence in the Canadian High Arctic: integrating glasioisostacy, eustacy and late deglaciation, *Canadian Journal of Earth Sciences*, **29**, 984-999.

England, J. 1997. Unusual rates and patterns of Holocene emergence, Ellesmere Island, Arctic Canada. *Journal of the Geologic Society, London*, **154**, 781-792.

England, J. 1998a. Support for the Innuitian Ice Sheet in the Canadian High Arctic during the Last Glacial Maximum. *Journal of Quaternary Sciences*, **13**, 275 - 280.

England, J. 1998b. Consensus on the Innuitian Ice Sheet during the last glacial maximum; new evidence and opportunity. 1998 Annual meeting, *Geological Society of America*, Toronto, October 26-29. Abstracts with Program.

England, J. 1999. Coalescent Greenland and Innuitian Ice during the Last Glacial Maximum: revising the Quaternary of the Canadian High Arctic, *Quaternary Science Reviews*, **18**, 421-456.

England, J. and Ó Cofaigh, C. 1998. Deglacial sea level history along Eureka Sound: the effects of ice retreat from a central basin to alpine margins. *Abstracts Geological Association of Canada*, Quebec City, **A-52**.

England, J., Smith, R., Evans, D.J.A. In press. The last glaciation of east-central Ellesmere Island, Nunavut: Ice dynamics, deglacial chronology and sea level change. *Canadian Journal of Earth Sciences.*

Fairbanks, R.G. 1989 A 17, 000 year glacio-eustatic sea level record: influence of glacial melting rate on the Younger Dryas event and deep ocean circulation, *Nature*, **342**, 637-642.

Falconer, G. 1966 Preservation of vegetation and patterned ground under a thin ice body in northern Baffin Island, N.W.T. *Geographical Bulletin*, **8**, 194-200.

Fyles, J.G. 1990. Beaufort Formation (Late Tertiary) as seen from Prince Patrick Island, Arctic Canada. *Arctic*, **43**, 393-403.

Hattersley-Smith, G.1974 Present arctic ice cover, Chapter 4 in Ives J.D. and Barry R.G. (eds.) *Arctic and Alpine Environments.*

Hättestrand, C., Stroeven, A.P. 1996. Field evidence for wet-based ice sheet erosion from the south-central Queen Elizabeth Islands, Northwest Territories, Canada. *Arctic and Alpine Research*, **28**, 466-474.

Hein, F.J., van Wagonger, N.A., Mudie, P.J. 1990. Sedimentary Facies and Processes of Deposition: Ice Island Cores, Axel Heiberg Shelf, Canadian Polar Continental Margin. *Marine Geology* **93**, 243-265.

Hein, F.J., Mudie, P.J. 1991. Glacial-marine sedimentation, Canadian Polar margin, north of Axel Heiberg Island. *Geographie Physique et Quaternaire*, **45**, 213-227.

Hodgson, D. A. 1982. Surficial materials and geomorphological processes, western Sverdrup and adjacent islands, District of Franklin. *Geological Survey of Canada*, Paper **81-9.**

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

Kleman, J., 1994. Preservation of landforms under ice sheets and ice caps. *Geomorphology*, **9,** 19-32.

Koerner, R.M. and Fisher, D.A. 1990. A record of Holocene summer climate from a Canadian high-Arctic ice core, *Nature*, **342**, 630-631.

Lamoureux, S. 2000. Five centuries of interannual sediment yield and rainfall induced erosion in the Canadian High Arctic recorded in lacustrine varves. *Water Resources Research*, **36**, 309-318.

Lamoureux, S., England J. In Press. Late Wisconsinan glaciation of the northwest sector of the Canadian Arctic Archipelago, *Quaternary Research*.

Lemmen, D.S., Gilbert, R., Aitken, A.E. 1991. Quaternary investigations in the Expedition Fiord area, west-central Axel Heiberg Island, Northwest Territories. *Geological Survey of Canada*, Paper **91-1B.**

Lemmen, D.S., Aitken, A.E., Gilbert, R.1994. Early Holocene deglaciation of Expedition and Strand Fiords, Canadian High Arctic. *Canadian Journal of Earth Sciences*, **31**, 943-958.

Locke, C.W., Locke, W.W. 1977. Little Ice Age snow-cover extent and paleoglaciation thresholds: North-central Baffin Island, N.W.T., Canada. *Arctic and Alpine Research*, **9**, 291-300.

MacLean, B., Sonnichsen G., Vilks, G., Powell, C., Moran, K., Jennings, A., Hodgson, D., Deonarine, B., 1989. Marine geological and geotechnical investigations in Wellington, Byam Martin, Austin and adjacent channels, Canadian Arctic Archipelago. *Geological Survey of Canada*, Paper **89-11**.

Maxwell, J.B. 1981. Climatic Regions of the Canadian Arctic Islands. *Arctic*, **34**, 225-240.

Miall, A.D. 1991. Late Cretaceous and Tertiary Basin Development and Sedimentation, Arctic Islands: Chapter 15 in *Geology of the Innuitian Orogen and Arctic Platform of Canada and Greenland*, H.P., Trettin (ed.); Geological Survey of Canada Geology of Canada, no. 3,

Miller, G.H., Bradley, R.S., 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.

Müller, F. 1963. Radiocarbon dates and notes on the climate and morphological history, in: Preliminary report 1961-1962, McGill University, Montreal, *Axel Heiberg Island Research Reports*, 169-172.

Ó Cofaigh, C. 1999. Late Quaternary glaciation and postglacial emergence, southern Eureka Sound, High Arctic Canada. *Unpublished Ph.D. thesis*, University of Alberta.

Ó Cofaigh, C., England, J., 1998. Configuration and dynamics of Late Wisconsinan glaciation in southern Eureka Sound, High Arctic Canada, *Geological Society of America,* Toronto, October 26-29.

Ó Cofaigh, C., England, J., Zreda, M. 1998. Configuration and dynamics late Wisconsinan glaciation in southern Eureka Sound, High Arctic Canada, Annual meeting, *Geological Society of America,* Toronto, October 26-29.

Ó Cofaigh, C., Lemmen, D.S., Evans, D.J.A., Bednarski, J. 1999. Glacial sediment-landform assemblages in the Canadian High Arctic and their implications for late Quaternary glaciation. *Annals of Glaciology* **28**, 195-201.

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

Stewart, T.G.,1991. Glacial marine sedimentation from tidewater glaciers in the High Arctic. *Geological Society of America*, **Special Paper 261.**

Stuiver,M.,Reimer J. 2000. Calib 4.0. http://radiocarbon.pa.qub.ac.uk/calib/calib.html

Sugden D.E. 1974. Landscapes of glacial erosion in Greenland and their relationship to ice, topographic and bedrock conditions in: Brown E.M., Waters R.S. (eds.) *Progress in Geomorphology*. Institute of British Geographers, Special Publication No. 7, 177-195.

Thorsteinsson, R.T. 1970. Geology Map 1299A *Geological Survey of Canada.*

Thorsteinsson, R.T. 1970 Geology Map 1303A *Geological Survey of Canada.*

Thorsteinsson, R.T., and Tozer, E.T. 1976. Geology of the Arctic Archipelago, Chapter 10 in: Geology and Economic Minerals of Canada, part B, *Geological Survey of Canada,* Economic Geology Report No.1.

Trettin, H.P. 1991. Tectonic Framework; Chapter 4 in Geology of the Innuitian Orogen and Arctic Platform of Canada and Greenland, H.P. Trettin (ed.); *Geological Survey of Canada,* Geology of Canada, no. 3.

Tushingham, A.M.1991. On the extent and thickness of the Innuitian Ice Sheet: A postglacial-adjustment approach. *Canadian Journal of Earth Sciences*, **28**, 231-238.

Walcott, R.I. 1972. Late Quaternary movements in North America; Quantitative evidence of glacio-isostatic rebound. *Reviews of Geophysics and Space Science*, **10**, 849 – 884.

Whillans, I.M., 1978. Erosion by continental ice sheets. *Journal of Geology*, **86**, 516-524.

Zreda, M., England, J., Phillips, F., Elmore, D., Sharma, P. 1999. Unblocking of the Nares Strait by Greenland and Ellesmere ice-sheet retreat 10,000 years ago. *Nature*, **398**, 139-142.