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THE GLACIAL HISTORY OF CENTRAL CAÑON FIORD, WEST-CENTRAL ELLESMERE ISLAND, ARCTIC CANADA

BY

VALERIE F. SLOAN ©

A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA
FALL 1990

ISBN 0-315-65049-4

UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR: Valerie F. Sloan

TITLE OF THESIS: The Glacial History of Central Canon Fiord, West-central Ellesmere Island, Arctic Canada.

DEGREE: Master of Science

YEAR THIS DEGREE GRANTED: 1990

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled THE GLACIAL HISTORY OF CENTRAL CAÑON FIORD, WEST-CENTRAL ELLESMERE ISLAND, ARCTIC CANADA submitted by VALERIE F. SLOAN in partial fulfilment of the requirements for the degree of MASTER OF SCIENCE.

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Charles Schweger.

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Date: September 28, 1990.

ABSTRACT

The last glaciation in central Canon Fiord, west-central Ellesmere Island, was characterized by the inception and expansion of ice caps on the plateaus surrounding Foster Creek and South Bay. At Foster Creek, east-central Canon Fiord, the precursor to the Agassiz Ice Cap advanced 8 km to the coast. While advancing to the fiord coast, this valley glacier overrode preglacial deltaic sediments which represent a former relative sea level at least 20 m above present. The valley glacier retreated in contact with a 131 m sea level, depositing deep water rhythmites. These are interpreted as turbid plume deposits which may represent sedimentation by a warm-based, tidewater glacier. Marine shells in growth position within these rhythmites at 94 m asl provide a minimum date of deglaciation of 8020 ± 90 BP. At South Bay, along west-central Canon Fiord, Wolf Ridge was capped by coalescing cirque glaciers whose meltwater drained into a 119 m sea level. Marine shells in growth position at 90 m asl in sediments likely coeval with a 119 m sea level were deposited at least 7930 ± 70 BP.

Prior to the last glaciation, a trunk glacier occupied the valley east of Wolf Ridge and flowed into relative sea level around 150 - 160 m asl, forming an ice shelf. Shell fragments in fossiliferous diamicton likely deposited by the ice shelf provided three AMS dates which are considered to be non-finite in age (all > 29,000 BP). This event may have been a recessional stage of a more regional glaciation that deposited high elevation erratics.

ACKNOWLEDGEMENTS

Working in the Canadian high arctic has been a treasure of an experience, from the glistening bright, snow-covered Disraeli Fiord to the deafening wind storms in central Canon Fiord, and I am very grateful to John England for having given me that opportunity. My interest in the Earth sciences and love for the Arctic has grown as a result. John's enthusiasm for his work is infectious, and his imagination inspiring. I also appreciate John's prompt and thorough editing of my thesis, and his caring supervision.

Thanks to Jan Bednarski, for always being there to answer questions and discuss the data and concepts - his patience was golden to a beginner. Bruce Rains' continual moral support and encouragement throughout my degree have been invaluable - and are very much appreciated. Don Lemmen first introduced me to the wonders of the arctic and has since been both supportive and instructive. Don has generously discussed and read my work, providing thoughtful, astute feedback, as well as encouragement. Thanks also go to Doug Hodgson, who has kept me questioning, and who provided helpful comments on the thesis manuscript as well as hour-by-hour humour during its production.

A number of people, in addition to those mentioned above, have been generous with their time in looking at slides, discussing data and ideas, and providing help of other kinds. Thanks go to Fred Bachhuber, Art Dyke, D.J.A. (Evs) Evans, W. Fieguth, Janice Hanright, James Hooper, John Kulig, Tom Morris, Mark Seemann and Stephen Traynor. Valuable comments on the thesis were made by the supervisory committee, Drs. England, Rains, and Schweger.

Trixa Marshall's enthusiasm and diligence as a field assistant were precious and are appreciated. David Bechtal, an honours student, also provided able assistance and good company in camp.

Assistance with preparation of diagrams and text for the thesis and conference papers was given by Dave Epp, Geoff Lester, Diana Liu, Randy Pakan, and André Prégent and is appreciated.

Logistical support was provided by Polar Continental Shelf Project, Department of Energy, Mines, and Resources. Financial support for this project was provided by the Boreal Institute for Northern Studies, University of Alberta, and by the Natural Sciences and Engineering Research Council of Canada (NSERC grant A6880 to John England).

A special mention goes to Trevor Bell, who has been a good friend and colleague throughout. Trevor has shared fieldtrips, conferences, deadlines, ideas, papers, coffee, and Big Rock with me all through these years at the U. of A. I treasure that shared experience, and very much appreciate his encouragment, his friendship, and having him as a colleague.

Finally, I would like to thank my parents, a pair of real rocks, for their continual support and good humour.

Here's to roaming thermarests!

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Chapter 1

Previous Research and Project Rationale

1.1 Introduction

West-central Ellesmere Island has been the focus of only a few Quaternary studies, and yet it is a key area in the controversy regarding the extent of the last glaciation in the Queen Elizabeth Islands. The principal debate concerns whether the geological evidence and the pattern of postglacial emergence record an extensive regional ice sheet or a minor advance of present day ice caps accompanied by the inception of plateau glaciers. One problem in deciphering the glacial record lies in the difficulty of distinguishing between evidence from the last glaciation and previous glaciations, particularly given the age limitations of radiocarbon dating. The following section outlines the regional models of the last glaciation and then focuses on the interpretations of the glacial history of west-central Ellesmere Island.

1.2 Regional Studies

The first glacial map of Canada that portrayed the ice cover of the Queen Elizabeth Islands was presented by Prest (1957). Subsequently, Craig and Fyles (1960), in a review of arctic Quaternary literature, proposed that the last glaciation of the arctic islands was characterized by coalescing glaciers that formed the Ellesmere-Baffin Glacier Complex. This complex was portrayed as a regional ice sheet that covered most of the northern and eastern arctic islands, contacting the Laurentide Ice Sheet along its south border (Fig. 1.1). Craig and Fyles (1960) suggested that the western Queen Elizabeth Islands supported either local ice caps or were ice-free during the last glaciation.

In the early 1970's, Blake (1970, 1972, 1975) also proposed a regional ice cover for the last glaciation based on the pattern of postglacial emergence along the south coast of Ellesmere Island and a broad, NE - SW oriented ridge of maximum emergence in the Queen Elizabeth Islands (Fig. 1.2). Blake believed that the arrival of driftwood and shells in the high arctic islands coincided with the disappearance of a former regional ice sheet (Blake, 1970, 1972). This reconstruction was based on the well established glacioisostatic principal that crustal depression -- and subsequent emergence -- was greatest where the ice was formerly the thickest (cf. Andrews, 1970). Blake did not place boundaries on the proposed last glacial "Innuitian Ice Sheet"; however, Prest (1967) and Mayewski et al. (1981) produced maps in which the ice margins encompassed all of the Queen Elizabeth Islands. These have generally been used to depict the extent of the Innuitian Ice Sheet (Fig. 1.3).

The reconstruction of ice sheets based on emergence patterns alone has been questioned repeatedly (England, 1976; Boulton, 1979). For example, these authors argued that postglacial emergence does not categorically indicate a former ice cover at a specific site because crustal loading also occurs in the peripheral depression beyond the margin of an ice load. Boulton (1979) emphasized that the glacial geological evidence should take precedence over

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Fig. 1.1. The Ellesmere-Baffin Glacier Complex proposed by Craig and Fyles (from Craig and Fyles, 1960).

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Fig. 1.2. Corridor of maximum postglacial emergence in the eastern Queen Elizabeth Islands (from Blake, 1970).

Diagram omitted due to copyright restriction.

Fig. 1.3. Extent of the last glaciation in Canada and Greenland as portrayed by Mayewski et al. (1970).

emergence data in the reconstruction of former ice sheets.

An alternative model supporting a limited ice cover during the last glaciation of the eastern Queen Elizabeth Islands challenged the hypothesis of an Innuitian Ice Sheet (England, 1976). On northeastern Ellesmere Island, a prominent moraine system less than 60 km from the present ice margin in the Grant Land Mountains (Hazen Moraines, Fig. 1.4) was interpreted as the last glacial limit (England, 1978). Throughout Archer Fiord and Lady Franklin Bay, radiocarbon dates on in situ marine shells at or close to marine limit indicated that the sea was relatively stable between 11,000 and 8,000 BP (England, 1983). The stability of the sea suggested that postglacial emergence, and therefore deglaciation, had not yet begun, and that the sea therefore coincided with full glacial conditions. Synchronous emergence after 8,000 BP throughout Archer Fiord and Lady Franklin Bay, interpreted as glacioisostatic rebound of the peripheral depression, provided further evidence that glaciers did not occupy the fiord during the last glaciation. This pattern of postglacial emergence combined with the position of the last ice limit led England to propose that during the last glaciation the fiords were occupied by the "full glacial sea" (England, 1983). Rather than forming a contiguous ice sheet, icefields expanded into a non-contiguous "Franklin Ice Complex" (England, 1978).

Much research has followed the initial debate concerning the Innuitian Ice Sheet published during the 1970's. Three doctoral dissertations recently completed in Clements Markham Inlet, Marvin Peninsula, and Philips Inlet on

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Fig. 1.4. Northeastern Queen Elizabeth Islands showing placenames in text, ice cover (shaded), and ice margins attributed to the last glaciation in previous studies. Hodgson's (1985) drift belt is along west-central Ellesmere Island and England's Hazen moraines are on northeastern Ellesmere Island (from England, 1990).

northern Ellesmere Island all supported a limited ice cover during the last glaciation (Bednarski, 1984; Lemmen, 1988; and Evans, 1988 respectively). Of particular interest are radiocarbon dates of full glacial age (11 to $\geq$ 23 ka) on redeposited terrestrial organics in northernmost Ellesmere Island which demonstrate the former presence of biologically productive refugia (Lemmen, 1988, 1989). Lastly, evidence of a limited last glaciation accompanied by a full glacial sea was observed on Hall Land, NW Greenland (England, 1985, 1987b).

1.3 Local Studies: West-central Ellesmere Island

In one of the earliest studies on western Ellesmere Island, Troelsen (1952, p.209) stated that:

> it was evident that the whole region had been covered by glaciers during Pleistocene time, and evidence was found [on Stor Island] that at one time this ice cover was continuous and that it had its centre, or centres, in eastern Ellesmere Island.

However, no reference to the age of this event was made nor was the evidence for this statement developed.

In a report on the Quaternary geology of western Ellesmere Island, Fyles (in Jenness, 1962 - in Hodgson, 1985) noted differences in relative weathering of bedrock and tentatively suggested that a late Pleistocene glaciation may have been partial in extent and that a regional glaciation occurred earlier. Similarly, Hattersley-Smith's (1969) interpretation of glacial deposits of two distinct glacial ages at the head of Tanquary Fiord "suggested that the newer drift dates from a later 'Meighen type' of ice cap, in contrast to a much more severe and widespread earlier glaciation."

More recently, Hodgson (1985) identified a 500 km long depositional boundary on west-central Ellesmere Island termed the "drift belt." The drift belt occurs 10 - 60 km beyond the present-day ice caps and is regarded as marking a regional ice margin of paleoclimatic significance (Fig. 1.4). The glacial landforms comprising the drift belt were deposited by the coalesced central Ellesmere Island and Sydkap ice caps between 9000 and 7000 BP. Three different types of paleo ice margins were identified: i) confluent valley glaciers common along the drift belt which generally terminated less than 20 km from present ice margins; ii) trunk valley glaciers which flowed at approximately right angles to the modern ice margins; and iii) trunk valley ice which drained low-lying plateaus supporting small ice domes or an expanded central Ellesmere Island ice cap.

Hodgson (1985) presented three alternative models for the last glaciation. Model A suggested that central and western Ellesmere Island were completely covered by ice which formed part of the Innuitian Ice Sheet proposed by Blake (1970). Hodgson pointed out that this reconstruction demands the catastrophic breakup of a thick ice sheet in the early Holocene for which there is no unequivocal paleoclimatic or geologic evidence. Model B proposed that the inner fiords' drift belt represented the last ice limit along west-central Ellesmere Island, independent ice caps occupied Axel Heiberg and Ellesmere Islands, and the sea remained raised throughout the last glaciation. However, a lack of evidence for a high stable sea or a transgressive sea in the outer fiords and Eureka Sound prevented full support of this model. Model C suggested that the last ice limit lay some distance down-fiord from the drift belt, based on the presence of striated bedrock beneath glaciomarine deposits distal to the drift belt in Strathcona and Bay Fiords (Fig. 1.4). Evidence for an ice cover over Braskeruds Plain suggests that low-lying plateaus at similar elevations (ca. 350 - 500 m) were also ice covered during the last glaciation. Hodgson suggests that perhaps neither the extensive pan-archipelago ice sheet (the Innuitian Ice Sheet) nor the limited expansion of present day ice caps properly explain the glacial and marine observations, and that an expansion of ice caps accompanied by the inception of plateau glaciers better characterize the last glaciation. Nonetheless, this model of a limited advance, concomitant with the inception of plateau ice caps, is similar to the minimal glaciation proposed in model B and unlike that of the Innuitian Ice Sheet of model A, for it suggests a land-based rather than a marine-based pan-archipelago ice cover.

In a recent study of Greely Fiord and its tributaries, England (1990) identified a prominent former ice margin encircling the fiord marked by moraines, meltwater channels, and former grounding lines (Fig. 1.5). Based on the glacial geology and associated raised marine deposits, this former ice margin was interpreted as recording the limit of the last glaciation. This suggested that tidewaters glaciers, plateau ice caps, and outlet glaciers advanced only 5 - 35 km

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Fig. 1.5. The last ice limit in Greely Fiord and outer Canon Fiord as documented by England (1990) (from England, 1990).

beyond their present margins where they contacted and calved into the sea (England, 1990). Evidence for a high, stable sea between 8,400 BP and 7,400 BP in Greely Fiord indicated that the sea remained relatively stable prior to initial deglaciation.

1.4 Project Rationale

Central Canon Fiord was chosen for study because of its importance to the glacial history of west-central Ellesmere Island. The area lies just south of England's (1990) study area and beyond the drift belt identified by Hodgson (1985). Central Canon Fiord also lies along the axis of maximum postglacial emergence interpreted by Blake (1970) as postglacial rebound caused by a regional ice load. The glacial and marine record of central Canon Fiord therefore will test these various reconstructions of the last glaciation. For example, was this area occupied by regional ice during the last glaciation (cf. Blake, 1970, 1972) or by local, plateau glaciers and an expanded Agassiz Ice Cap (cf. Hodgson, 1985; England, 1990)? These alternative models have been carefully outlined by Hodgson (1985).

This study was conducted around two base camps, South Bay and Foster Creek, in central Canon Fiord visited between June and July, 1988 (Fig. 1.5). The principal objectives included: (1) mapping Quaternary landforms and deposits, (2) determining the limit of the last glaciation versus earlier glaciations, (3) surveying raised marine shorelines associated with these changing ice loads and collecting material suitable for ^{14}C dating, and (4) synthesizing the glacial and marine evidence for a local interpretation that could provide insights into the last glaciation and the nature of postglacial emergence in west-central Ellesmere Island.

Chapter 2

Study Area

2.1 Location and Physiography

Canon Fiord is a NW - SE trending body of water which divides the western side of central Ellesmere Island at ca. 80^{0}N (Fig. 2.1). The fiord is 118 km long, ca. 10 km wide, and its bathymetry is unknown. East of Canon Fiord, a highland plateau supports the Agassiz Ice Cap; an outlet glacier of the Agassiz Ice Cap calves at the head of Canon Fiord. This plateau is incised by several small, deep valleys which are occupied by outlet glaciers of the Agassiz Ice Cap. Two such glaciers flow into one valley located on the east coast of Canon Fiord, 10 km south of Caledonian Bay (Fig. 2.1). These glaciers terminate five to eight kilometres from the coast and the narrow, U-shaped valley, which lies 600 m below the plateau surface, is referred to here informally as Foster Creek. The abundance of glacial and raised marine sediments in Foster Creek led to its selection as one of two sites in this study.

Canon Fiord separates the ice-covered plateau from the lower relief of the Fosheim Peninsula to the west. To the south of Canon Fiord, strike ridges such as the locally ice-capped Sawtooth Mountains extend southward across the neck of Fosheim Peninsula (Fig. 2.1). Broad, low valleys between these glacierized ridges drain to the north and south, supporting lush arctic vegetation. The valley extending southward from South Bay, southwest Canon Fiord, is bordered by a glacierized ridge to the west rising to 1200 m - here informally named Wolf

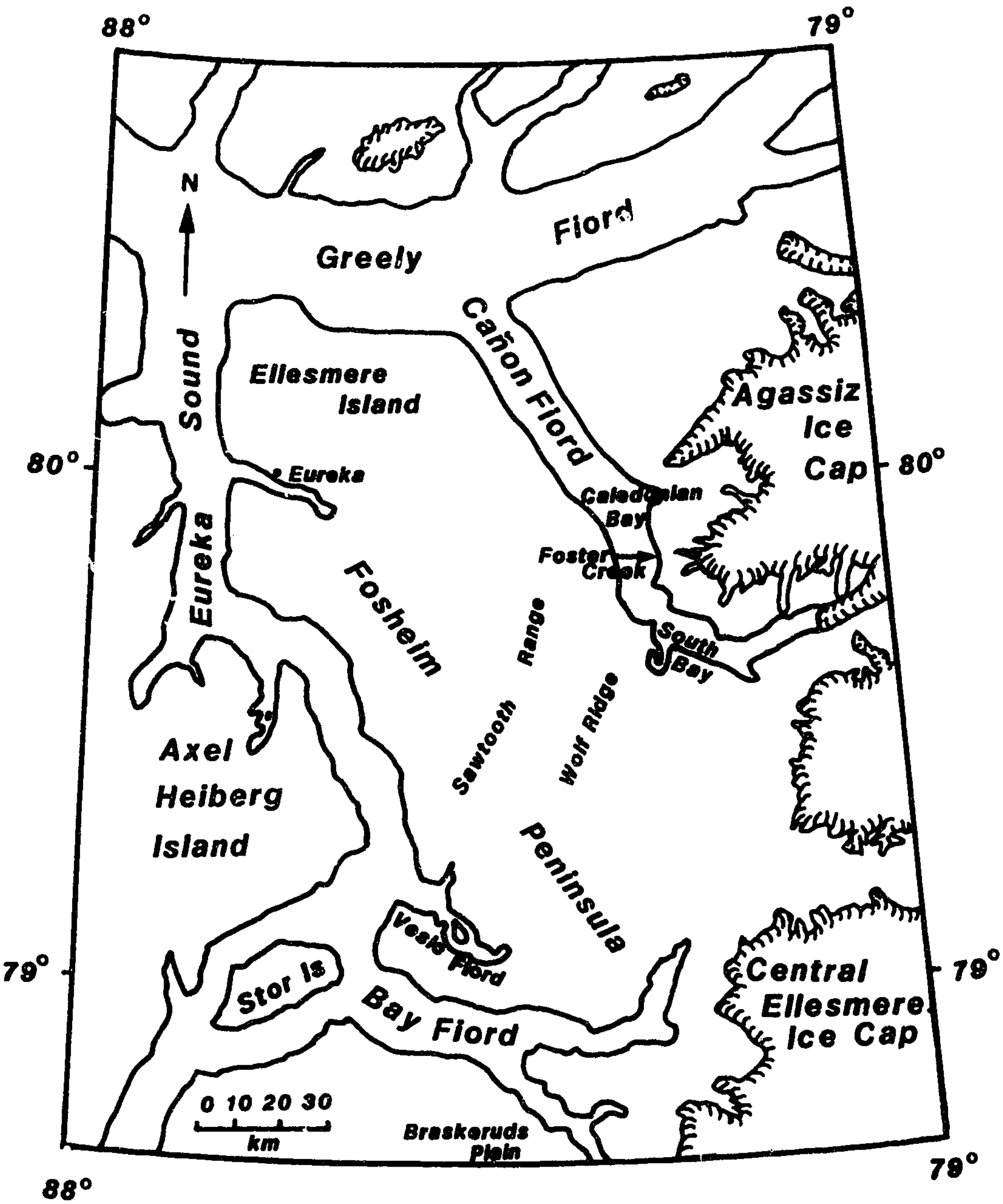

Fig. 2.1. Map of west-central Ellesmere Island showing study areas, South Bay and Foster Creek.

Ridge - and an ice-free plateau at 500 m to the east. The Agassiz Ice Cap lies 35 km to the east. The South Bay area constitutes the second area of focus in this study. South Bay has a 1000 km^2 catchment area and drains a valley which extends 55 km southward across the Fosheim Peninsula. The valley is characterized by a broad, low-grading valley floor flanked by steep bedrock walls. The valley widens to 5 km at its mouth where it opens into a hook-shaped embayment called South Bay.

2.2 Geology

The geology of Canon Fiord consists entirely of sedimentary rocks deposited in two successive basins which were present for much of the Phanerozoic. These include the Shelf Province of the Franklinian Mobile Belt and the subsequent Sverdrup Basin (Fig. 2.2). Each basin filling was concluded by tectonism which imparted a strong NE / SW grain to the present landscape. The sediments range from fine- to coarse-grained, well- to poorly-consolidated clastic and carbonate rocks, with some evaporites (Trettin, 1989). Granitoid rocks of the northernmost Canadian Shield are exposed 50 km to the southeast.

Sediments deposited in the Shelf Province of the Franklinian Mobile Belt outcrop to the east of Canon Fiord, whereas the Sverdrup Basin sediments make up the whole of the Fosheim Peninsula to the west and southwest of the fiord; however, an outcrop of Sverdrup Basin sediments occurs on the east side of central Canon Fiord between Caledonian Bay and Foster Creek (Fig. 2.2). The

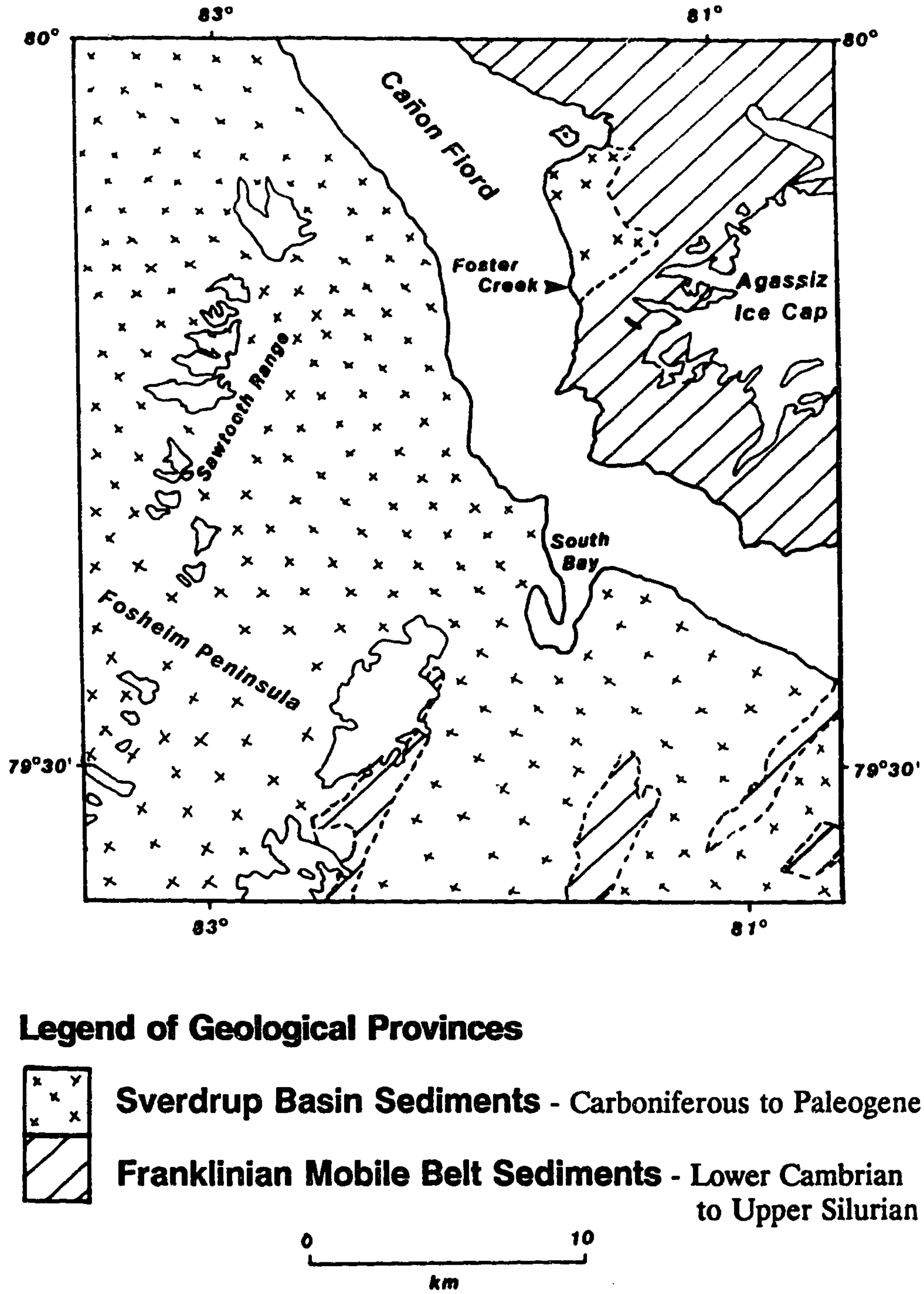

Fig. 2.2. Map of the bedrock geology in Canon Fiord (Source: Geological Survey of Canada, 1972, Geology map 1308).

contact between the two geologic provinces crosses outer Foster Creek along a SW - NE strike, providing a useful boundary across which erratics can be traced. Ordovician limestones and dolomites outcrop east of the contact in upper Foster Creek, and Upper Paleozoic redbeds of the Sverdrup Basin outcrop to the west.

2.3 Climate

Fosheim Peninsula and adjacent lowlands are classified as an "intermontane region" which is protected from the moderating effects of the central Arctic Ocean and southerly cyclonic activity by the surrounding mountains of Axel Heiberg and Ellesmere Islands (Edlund and Alt, 1989). This results in a climate which is characterized by extreme temperatures and aridity. Canon Fiord dissects both the mountainous and lowland regions; however, the study area is primarily within the lowlands of the intermontane region.

The intermontane region is characterized by regional minimums of cloud cover, high summer insolation, and maximums of temperature and melt season duration. The area has exceptionally warm July temperatures (ca. $>10^0$C) for such a high latitude and supports abundant high arctic vegetation (Edlund and Alt, 1989). During the winter, minimal cloud and a severe energy deficit produce strong temperature inversions and extreme surface temperatures (Maxwell, 1981). The mean annual temperature range at Eureka, Fosheim Peninsula, is 43^0C, with a mean monthly temperature of -35^0C in January (maxwell, 1981) and a mean July maximum temperature of 7.8^0C (1964 - 1972)

Peninsula, experience warmer temperatures in summer and have a larger annual temperature range due to increased continentality. For example, Hot Weather Creek, 25 km east of Eureka, had a mean July temperature of 12.7^{0}C in 1988, averaging 5.5^0 C warmer than Eureka (Edlund et al., 1989). The Canon Fiord area likely experiences temperatures similar to Hot Weather Creek, for it lies within the interior of Ellesmere Island.

The rainshadow effect caused by the mountains surrounding the intermontane region results in extreme aridity. The Fosheim Peninsula receives the lowest mean annual precipitation in Canada (Maxwell, 1981). Eureka's 30 year mean annual precipitation is 64 mm (Edlund et al., 1989), 35-40% of which falls as rain (Maxwell, 1981). Central Canon Fiord likely receives a similar amount of precipitation, whereas inland locations such as Hot Weather Creek are even more arid due to their increased summer temperatures, hence evapotranspiration. The adjacent highlands on Ellesmere and Axel Heiberg Island receive slightly more precipitation, approximately 200 mm per year (Maxwell, 1981). This regional aridity has put severe constraints on past glacial activity elsewhere on northern Ellesmere Island (England and Bradley, 1978).

One critical effect of the low temperatures on Ellesmere Island is the presence of permafrost. The active layer is within 50 - 100 cm of the surface while the permafrost base is > 400 m below the surface (Ritter, 1978). Permafrost an important factor influencing the geomorphological processes affecting the landscape.

2.4 Paleoglaciation Levels

The glaciation level is the elevation which separates higher glacierized mountain summits from lower non-glacierized summits (Miller et al., 1975). Above this level, most upland areas with an adequate surface area are ice covered. The level is calculated from the arithmetic mean of the highest ice-free summit and the lowest ice-covered summit in an area. Glaciation levels vary substantially across the Queen Elizabeth Islands and locally from one part of Ellesmere Island to another. For example, on the northwest coast of Ellesmere Island glaciers accumulate at sea level whereas the glaciation level rises rapidly to > 1300 m asl in the Grant Land Mountains less than 100 km to the south.

Glaciation levels reflect the mean elevation of the long term snowline and are indicative of regional climatic gradients. If considered in relation to the topography, glaciation levels can also reflect the "sensitivity of an area to changes in local mass balance (climate change)" (Miller et al., 1975, p.156). Factors controlling glaciation levels are poorly understood; however, climate and topography are evidently the primary controls (Miller et al., 1975).

Glaciation levels generally increase with relief; however, the low lying intermontane region of Ellesmere Island has relatively high glaciation levels relative to other areas of similar relief because of high summer net radiation and the rainshadow effect which increases ablation and reduces winter precipitation, respectively. The Sawtooth Mountains and the plateaus bordering the low relief Fosheim Peninsula have glaciation levels ranging from 900 to >1100 m.

In Quaternary studies, paleoglaciation levels are determined by the distribution of former glaciers. For example, numerous studies have established paleoglaciation levels for the Little Ice Age (Falconer, 1966; Locke and Locke, 1977; Dyke, 1983; Edlund, 1985). Extensive lichen-free surfaces indicate that plateau areas presently a few hundred metres below the present glaciation level were formerly ice-covered. On eastern Baffin Island, the glaciation level during the Late Wisconsin was ca. 350 to 450 m below the present glaciation level (Miller, 1973).

Based on the elevation of a topographically unrestricted cirque end moraine, as well as the distribution of former plateau ice caps, England (1986) proposed that the paleoglaciation level in inner Greely Fiord (presently 900 m asl) dropped to ca. 475 m asl during the last glaciation. In western Ellesmere Island, if the paleoglaciation level were at 475 m asl, then ice caps would develop on numerous plateaus, present day glaciers would expand, and only lowlands would be ice-free. This scenario corresponds with the model (C) proposed by Hodgson (1985) for western Ellesmere Island. This is supported by evidence that ice formerly covered Braskeruds Plain, western Ellesmere Island, which is 350 - 500 m asl (Fig. 1.4).

Chapter 3

Methodology

3. 1 Methodology

The reconstruction of late Quaternary paleoenvironments in this study involved the investigation of glacial and marine sediments and landforms. This included preliminary air photo interpretation followed by fieldwork in central Canon Fiord, Ellesmere Island, and the subsequent dating of organic samples. Initially, airphoto interpretation was carried out in order to gain familiarity with the field area, to identify and interpret glacial and marine landforms, and to plan which areas should be emphasized in the field. During the field work, surficial deposits and related landforms were described using standard mapping techniques of the Geological Survey of Canada. Former ice margins, denoted by moraines, kames, and meltwater channels, were mapped in relation to raised marine deposits in South Bay and Foster Creek. The marine deposits were investigated in terms of their stratigraphy, sedimentology, and organic content and were related to former sea levels which were surveyed using altimetry. Marine shells found within these marine sediments were extracted and subsequently ^{14}C dated, thereby providing a chronology for former ice margins and sea levels in central Canon Fiord. These techniques and their associated problems are outlined below.

Stratigraphic sections observed in the field area were logged, interpreted, and sampled using standard stratigraphic techniques. This involved recording the location, geomorphic context, and elevation of the section, as well as describing the units based upon their thickness, structure, texture, and lithologic composition.

The elevation of raised marine deposits was measured using a Wallace and Tiernan altimeter which has an accuracy of $\pm$ 2 m in 100 m. Since the instrument is temperature and pressure dependent, temperature readings and corrections were made for each measurement, and transects were closed as frequently as possible using high tide or camp as the datum in order to allow for pressure correction. Air pressure is relatively stable in the intermontane region of west-central Ellesmere Island and therefore presented little problem for most of the surveying; however, occasionally air pressure would change dramatically during a transect and adjustments were made to the measurements based on the rate of pressure change during the transect. To eliminate as much error as possible, key sites had at least two corresponding elevations from two different transects. The tidal range in this area is less than 50 cm, hence little error was introduced by this (Evans, 1988).

The age of marine sediments was established through ^{14}C dating of shells found in growth position. Throughout the study area marine shells were collected from beaches, deltas, and marine rhythmites exposed by rivers downcutting through raised marine deposits. These samples were excavated by knife and placed in sterile plastic sample bags for transportation. Shells were prepared for analysis by scraping foreign material off the shell surfaces and immersing them in an ultrasonic bath of distilled water. Two samples weighing > 30 g were dated using conventional radiocarbon analysis (beta counting) at the Geological Survey of Canada, Ottawa. Smaller samples (2 to 9 g) were sent to the ISOTRACE Laboratory, University of Toronto for accelerator mass spectrometry (AMS) dating in which the carbonate fraction was analyzed. One advantage of AMS dating is that individual shell fragments can be dated rather than larger bulk samples of shells which could contain a mixture of ages. Priority for dating was given to those shells most closely associated with a former sea level marking marine limit.

Errors in the interpretation of raised marine features may lead to inaccurate emergence data (cf. Andrews, 1970). For example, the uppermost observed shoreline may not necessarily represent the local marine limit if a higher shoreline was altered by erosion or if sea ice prevented its establishment. Problems also exist in attributing in situ shells within isolated, fine-grained sediments to a specific former sea level, because most arctic shells live tens of metres below sea level. In northeastern Ellesmere Island, shells are commonly found in such fine-grained sediments and often these do not underlie the foreset or topset beds of well defined deltas. Adjacent to Canon Fiord, most of the shells selected for dating were found in deltaic and littoral deposits which were considered coeval with marine limit.

In South Bay, fieldwork was restricted to the west side of the valley because it was not possible to cross the river. Airphoto interpretation was conducted and altitudinal measurements on delta apexes were taken with an inclinometer and altimeter from the west side of the valley; however, these observations are only supplemental and would require further verifications.

Chapter 4

Geomorphology and Raised Marine Deposits: South Bay

4.1 Introduction

In this chapter, field observations on the glacial geomorphology and raised marine deposits of South Bay will be presented and discussed, beginning with Wolf Ridge adjacent to South Bay, followed by the area surrounding South Bay. Glacial loading and the sea level history are closely interrelated, and so glacial and marine landforms will be considered together.

4.2 Wolf Ridge (informal name)

Glacial landforms and marine deposits occur on Wolf Ridge, west of South Bay (Figs. 2.1, 4.1). A plateau ice cap presently occupies the highest part of Wolf Ridge (Fig. 4.1). Four kilometres to the north of the glacier, a low saddle area extends across the ridge at ca. 400 m asl. The contrast in the vegetation adjacent to the ice cap and within the saddle is striking. The area within 4 km of the ice cap, above ca. 600 m asl, is lichen-free, relatively unweathered, and only slightly eroded, whereas the saddle is deeply gullied, lichen-covered, and exhibits periglacial stripes (Fig. 4.1). These differences are due to the saddle's longer exposure to subaerial erosion and revegetation. Elsewhere in the central arctic similar lichen-free zones are interpreted as representing the extent of ice during the Little Ice Age (Locke and Locke, 1977; Edlund, 1985). In the case of Wolf Ridge, the length of exposure of the lichen-free area is unknown.

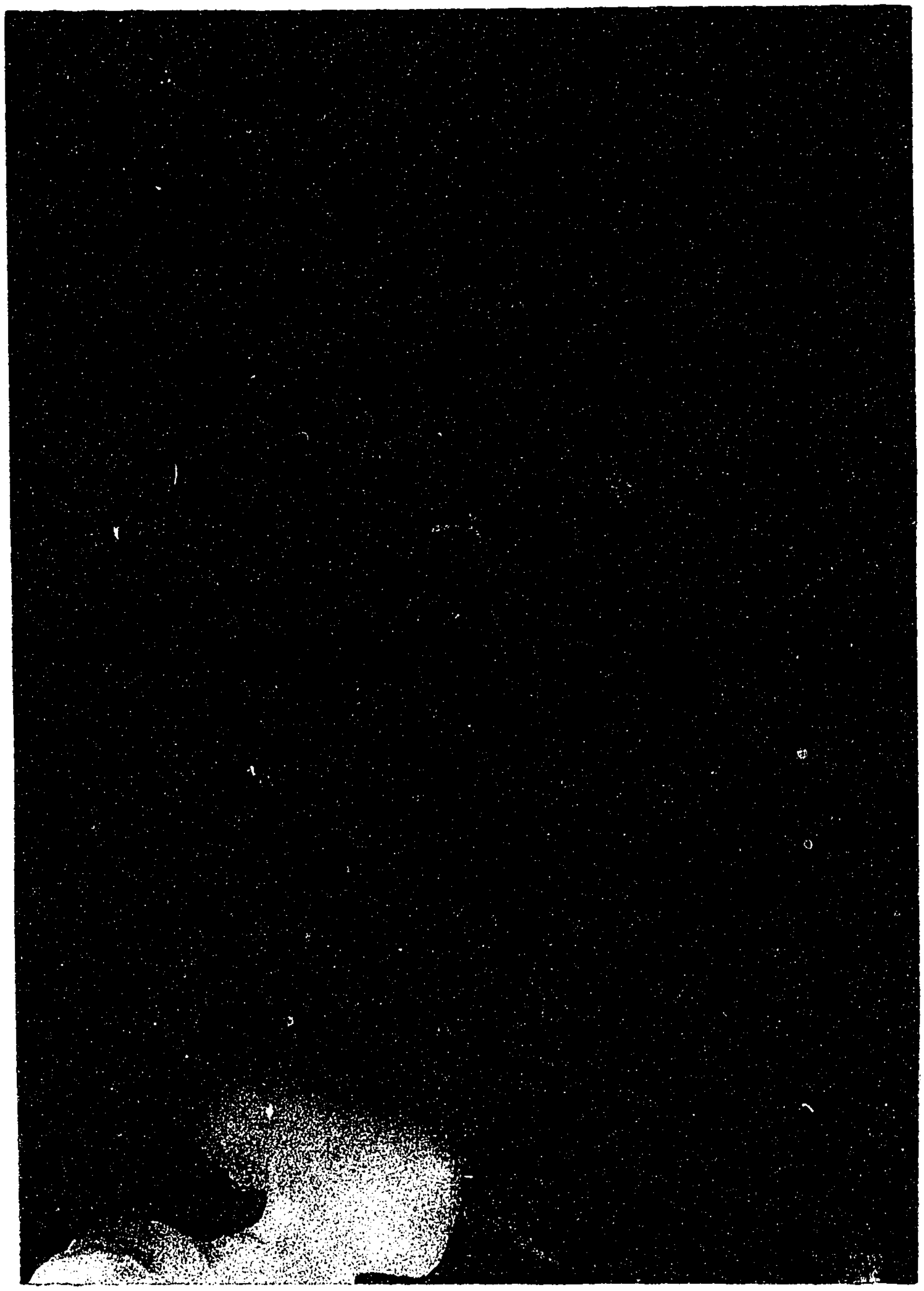

Fig. 4.1. Vertical airphoto of Wolf Ridge, west of South Bay, SW Canon Fiord, showing former glacial meltwater channels (barbed arrows, crossed lines), cirque basins (hatched arcuate lines), fossiliferous marine silts (dotted area) and their upper limit (dashed line), moraines (lines with dots), till blanket (enclosed areas with triangles) and raised marine terrace (arrow showing location of shell sample GSC-4967) discussed in text. The meltwater channels breaching the ridge crest are located at the head of the northernmost cirque. Width of photo ca. 12 km.

Prominent gullies cut in bedrock adjacent to the glacier and in the saddle are interpreted as former meltwater channels (Fig. 4.1). The saddle's eastern basin is cirque-like and contains a veneer of diamicton (polygonized in places) which may represent till. A moraine one kilometre long, 50 m wide and ca. 8 m high containing local lithologies borders the northern edge of the saddle (Fig. 4.1). These glacial features are interpreted as evidence of a cirque in the saddle linked to an expanded ice cap, which deposited till, a moraine, and cut meltwater channels upon its retreat.

At the northernmost end of Wolf Ridge, two small cirques are cut into the centre of the ridge (Fig. 4.1). One of the cirques contains a thick deposit of diamicton interpreted as a till blanket. A 2 km long moraine occurs downslope of the two cirques. The position of this moraine indicates that the former cirque glaciers were unconstrained by topography. The paleo-equilibrium line altitude (ELA) of former cirque glaciers can be estimated using the elevation of the ridge crest at ca. 500 m and the moraine at ca. 380 m. If the ratio of the accumulation area to total glacier area is 0.65 $\pm$ 0.05 (Porter, 1970, in England, 1986), then the accumulation area would have extended about two thirds of the distance from the ridge crest to the moraine, i.e. to ca. 420 m asl. This paleo-ELA is comparable to that of $\leq$ 400 m asl in inner Greely Fiord calculated by England (1986). Given that relative sea level was at least 100 m above present during the last glaciation, the paleo-ELA was in fact within 300 m of sea level.

Two gullies cutting diagonally across the crest of one cirque's headwall are interpreted as meltwater channels formed along the margin of a former glacier (Fig. 4.1). These channels lead to a gully that runs down the eastern side of Wolf Ridge. It is likely that the two former cirque glaciers coalesced to overtop the ridge, perhaps as part of a former plateau ice cap, and sent meltwater down the eastern flank of Wolf Ridge to Canon Fiord.

Several closely-spaced, parallel gullies lead down the eastern flank of Wolf Ridge (Fig. 4.1). The heads of these gullies are located near the ridge crest and have almost no watershed; therefore, it is proposed that these are meltwater channels which drained a former plateau ice cap on northern Wolf Ridge. These gullies commonly terminate at ca. 120 m asl (Fig. 4.1) which must represent the former base level for these channels. Below these channels, the slopes are blanketed with fossiliferous marine silts up to 119 m asl, the local marine limit. The termination of the gullies at the point of the uppermost marine deposits indicates that the local plateau ice cap was contemporaneous with a 119 m sea level.

A raised marine delta at 53 m asl occurs at the junction of the small tributary valley and South Bay. Several raised marine deltas also occur in the main valley where gullies descend from the adjacent plateau. The apexes of these deltas are 47 - 86 m asl, lower than the local marine limit. The deltas likely represent glaciomarine sedimentation at later stages of deglaciation as relative sea level dropped or, alternatively, they are remnants of earlier sediments deposited during higher sea levels.

4.3 Raised Terrace Below North Cliff

The northern end of Wolf Ridge terminates in a prominent cliff bordering Canon Fiord. A low, structural ridge extends from the cliff base towards the fiord, forming a projection in the coastline; a large, raised, depositional terrace is inset on the northwestern flank of this ridge (Fig. 4.1). This terrace is 119 m asl, ca. 100 m wide, and extends 400 m along the cliff base (Figs. 4.1, 4.2, 4.3). The bedrock at the back of the terrace has been notched across strike (Fig. 4.3). The outer part of the terrace is extensively gullied, revealing fossiliferous, fine-grained rhythmites, whereas the surface of the terrace consists of somewhat coarser sediments originating in local redbeds of the Canyon Fiord Formation (Fig. 4.2).

The presence of shells in growth position within the sediments demonstrates that the terrace is of marine origin. The terrace surface marks a former sea level at least 119 m asl. In situ bivalves of Mya truncata at 90 m asl within the sediments were dated 7930 ± 70 BP (GSC-4697) and provide a minimum age on the 119 m marine limit.

This raised marine terrace, which has a distinctive red hue, is inter[illegible]d as the product of longshore drift along the Canon Fiord coastline. Red sa[illegible] and other lithologies capping Wolf Ridge have been eroded by former [illegible] glaciers that incised the redbeds. The resulting sediments were carried by meltwater to the fiord. These sediments were then transported by longshore drift along the coast where they were trapped by the protruding bedrock ridge

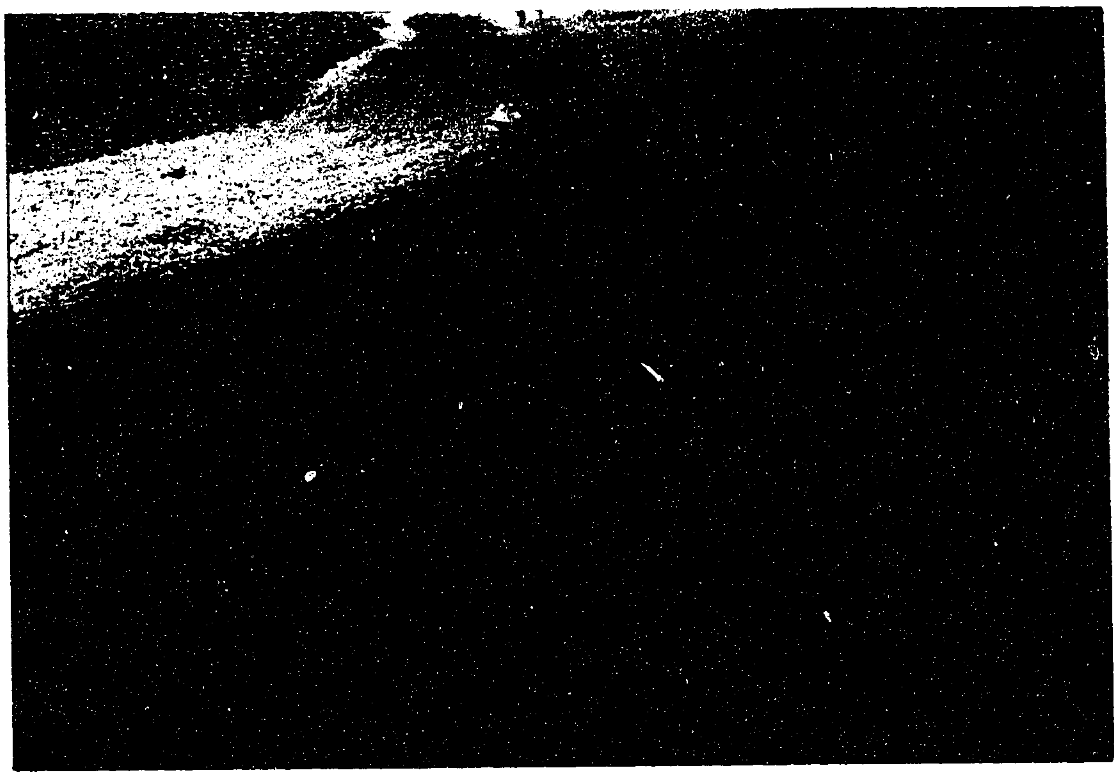

Fig. 4.2. An aerial view of the terrace at the northern end of Wolf Ridge at 119 m asl. Site of shell sample GSC-4697 is in the gully indicated with the arrow.

Fig. 4.3. The terrace surface at 119 m asl. Note the incision into bedrock with vertical dip at right background. Scale is provided by the individual on outer terrace, at left.

which formed a natural groyne projecting into the 119 m sea. Today, strong, northern winds blow up-fiord forming active sediment plumes; similar winds likely caused littoral drift in the past.

An alternative interpretation is that the terrace represents a kame or delta deposited by meltwater from a valley glacier in South Bay. However, if this were the case, one would expect lateral meltwater channels to have formed along the eastern flank of Wolf Ridge and coarse glaciofluvial materials to be associated with the terrace deposit. No such evidence exists. To the contrary, meltwater channels on Wolf Ridge are distributed radially and terminate at the upper level of the fossiliferous marine silts. Furthermore, the fossiliferous silts on the eastern flank of Wolf Ridge and the raised marine terrace occur at the same elevation indicating that they were deposited contemporaneously in a 119 m sea. An ice-contact origin for the terrace therefore is rejected in favour of its origin by littoral drift distal to a local upland ice limit.

In summary, the geomorphology and raised marine deposits on the eastern plateau and Wolf Ridge do not indicate the presence of a former trunk glacier in the main valley during the last glaciation but rather local plateau ice which was coeval with a 119 m sea dated at least 7930 $\pm$ 70 BP. Figure 4.4 is a schematic representing the extent of ice cover in South Bay during the last glaciation.

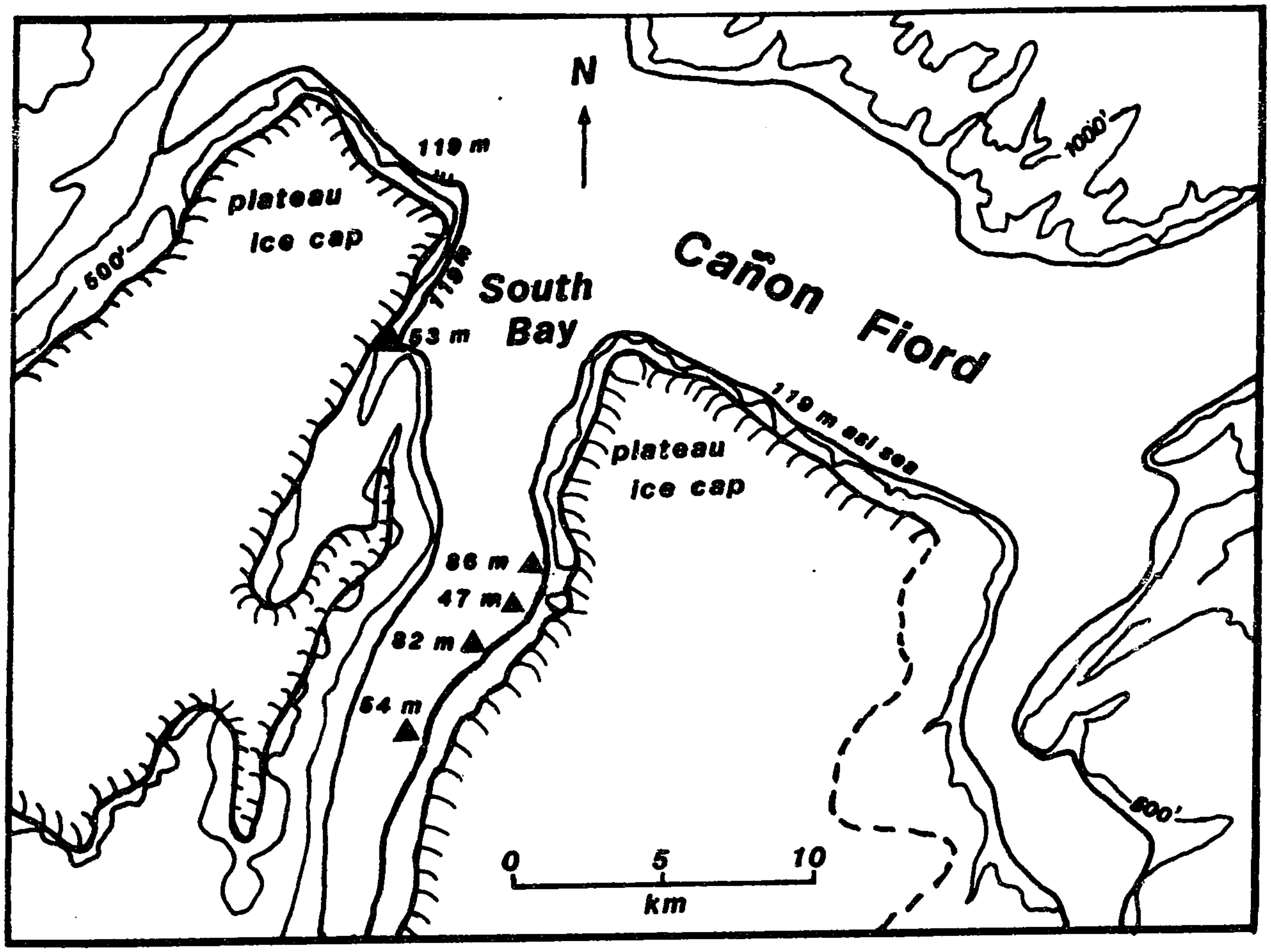

Fig. 4.4. Schematic diagram showing the extent of ice cover during the last glaciation at South Bay, central Cañon Fiord. Small, black triangles represent raised marine deltas. Contour interval 500 feet or 152 m.

4.4 South Bay

On the ridge which separates South Bay from its small tributary valley, a prominent bench crosses the strike of the steep northeastern slope (Figs. 4.5, 4.6). The bench is depositional and slopes downvalley from 177 to 167 m asl where it forms a broad terrace of sub-angular to sub-rounded cobbles and gravel (Figs. 4.5, 4.7). The bench and adjoining terrace are 1 km long with a north-inclined gradient of 1:100. Below the terrace, the slopes are draped with polygonized, poorly sorted, matrix-supported cobbles.

Twenty and forty metres upslope, similar but shorter (ca. 200 m long) landforms parallel the prominent bench (Figs. 4.5, 4.6). Both of these landforms have low gradients similar to the bench downslope. The uppermost landform is incised by a gully which leads down the western side of the bedrock ridge. The intervening slopes are draped by material similar to that which veneers the lower slopes. At elevations above these depositional landforms, quartzite erratics are found on the ridge and adjacent highlands.

The depositional bench at 177 - 167 m asl is interpreted as a lateral moraine deposited by ice which formerly occupied South Bay (Fig. 4.6). The moraine grades into a kame or delta at the northern end of the ridge. The shorter landforms upslope are believed to be moraines deposited just prior to the lower and more prominent moraine, while the gully incising the uppermost landform is interpreted as a meltwater channel. The poorly sorted, cobble-laden material draping the slopes is interpreted as ice-contact sediment.

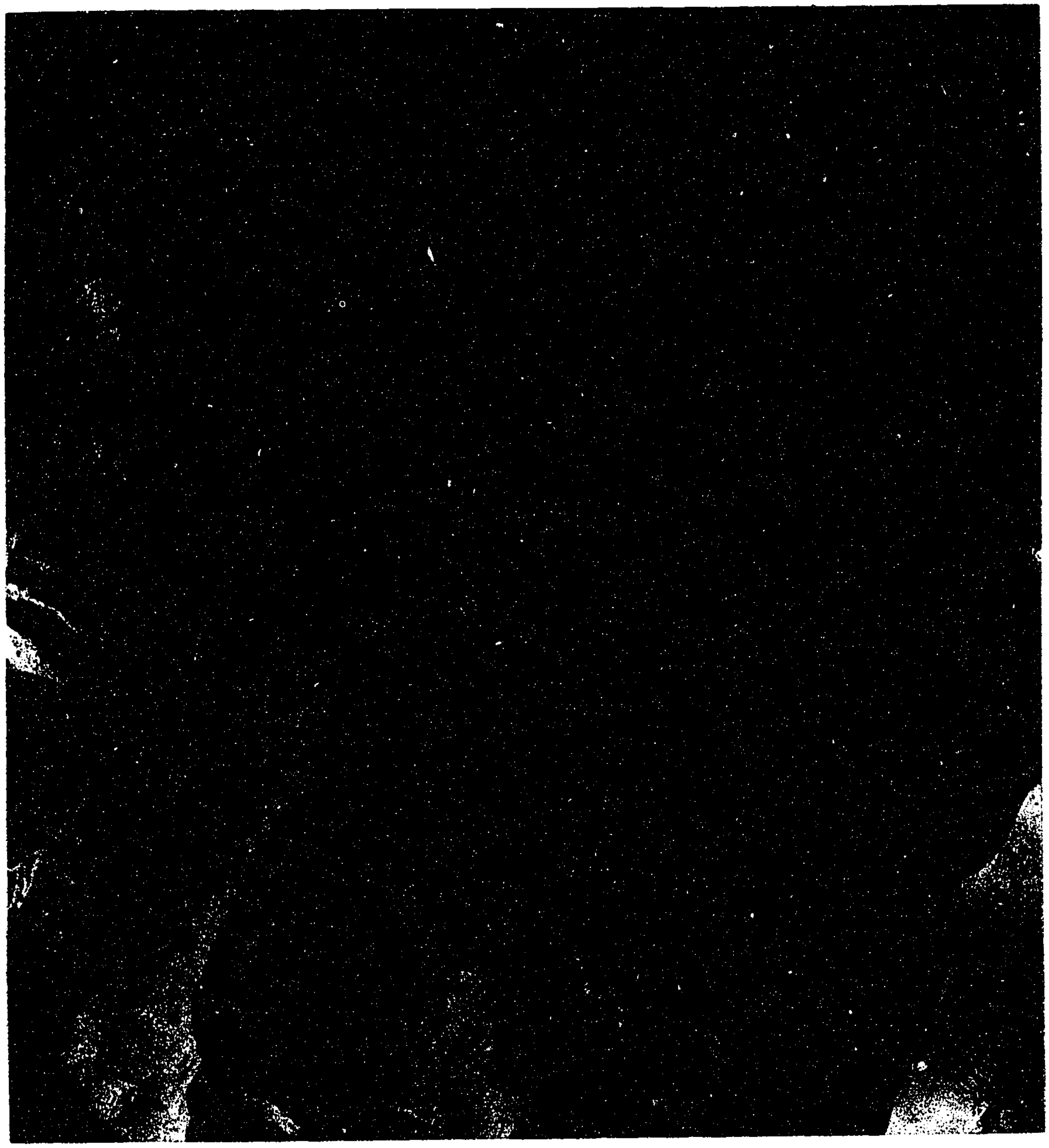

Fig. 4.5. Vertical airphoto of South Bay, SW Canon Fiord, showing subhorizontal ridges (between small black triangles), kame/delta (below arrow marked 167 m asl), fossiliferous diamicton (within enclosed, dashed lines), the former meltwater channel (barbed arrow, bottom right corner of photo), discussed in text. The locations and elevations of shell samples (TO-1198, 1200, 1201) on the peninsula and at the apex of the lobes are shown with long, narrow arrows. The width of the photo is ca. 7 km.

Fig. 4.6. View of subhorizontal, depositional bench which crosses the bedrock strike of a mountain in South Bay. The bench slopes from 177 m asl at the left to 167 m asl at the right. Two similar, shorter features can be seen upslope at the right end of the bench.

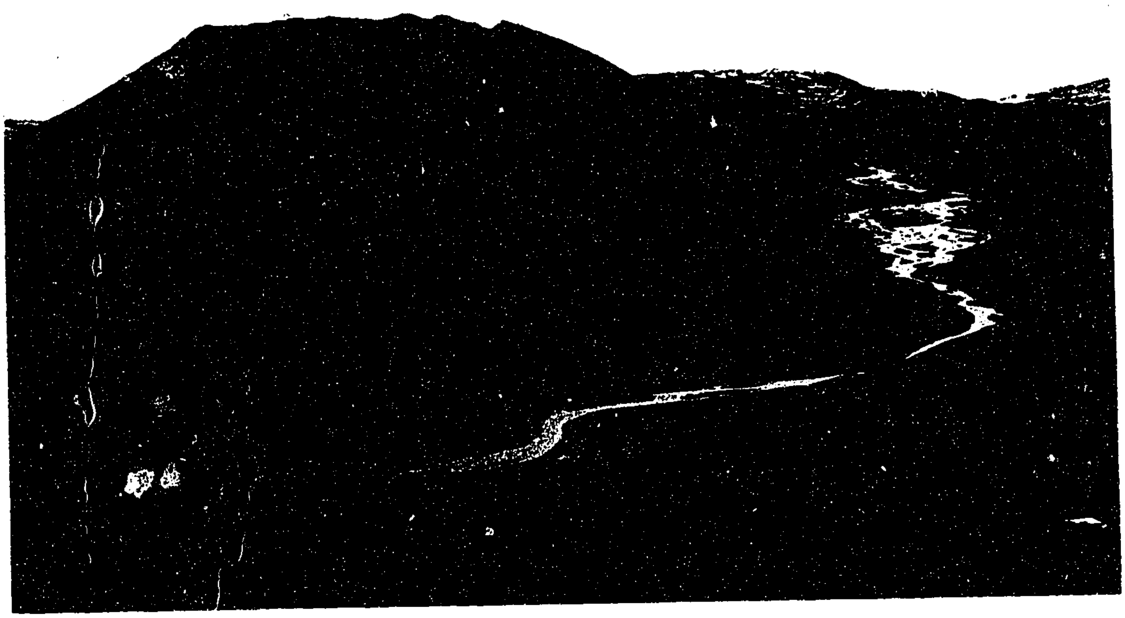

Fig. 4.7. The subhorizontal bench at 177 - 167 m asl in South Bay from the north.

Two kilometres downvalley from the kame/delta on the eastern flank of Wolf Ridge, fossiliferous diamicton mantles the slope (Fig. 4.5). This diamicton drapes two lobe-shaped slopes, the apexes of which are 167 m asl (Fig. 4.8). It consists of yellowish brown, massive silt and clay which contains clasts and shells. Shell fragments were found up to 167 m asl (Fig. 4.9). Two shell fragments from 167 m asl were submitted for AMS analysis and provided dates of 38,100 $\pm$ 380 BP (TO-1200) and 34,950 $\pm$ 340 BP (TO-1201). Radiometric dates > 25,000 BP are generally interpreted as minimum ages because of potential contamination by younger carbon (Bradley, 1985). This profile of the lateral moraine and adjacent kame/delta at 177 - 166 m asl corresponding to the upper limit of the fossiliferous diamicton downvalley is discussed below.

East of South Bay, a linear notch at ca. 150 m asl is cut into the side of the steep bedrock slope, forming a channel along the valley wall (Fig. 4.5). This notch cuts diagonally across the bedrock strike, indicating that it is not structurally controlled, and it slopes downvalley. Four kilometres upvalley, two gullies which parallel the valley are cut into the adjacent plateau surface at ca. 200 m asl. These are all interpreted as lateral meltwater channels formed along the margin of the ice which occupied the valley and South Bay.

The peninsula in South Bay is controlled by the curvilinear bedrock strike which forms a smooth arcuate ridge (Figs. 4.5, 4.8). A fossiliferous diamicton drapes the upper surface of the peninsula to its crest at 134 m asl. Angular quartzite erratics as large as 1 m^2 occurred at ca. 50 m asl. The diamicton has

Fig. 4.8. The two lobes (foreground) and the peninsula (background) which are draped with fossiliferous diamicton. Shell fragments collected from both locations provided AMS dates between 29 and 38 ka which are interpreted as infinite.

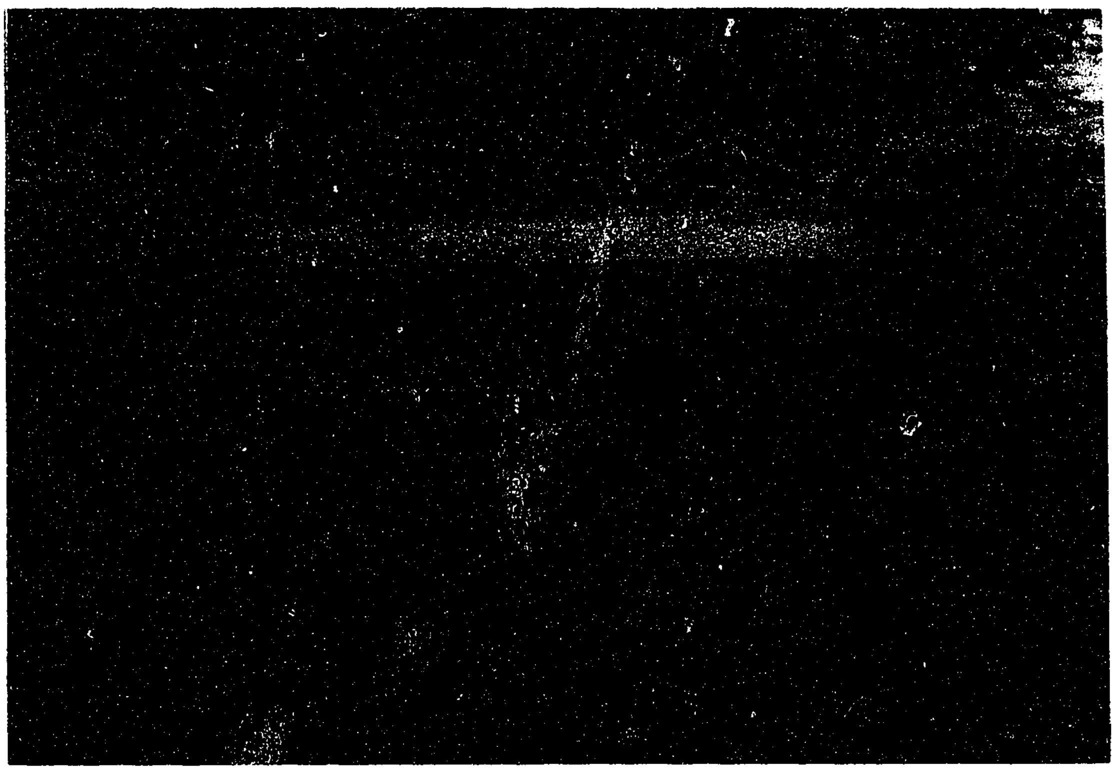

Fig. 4.9 The apex of the lobes draped with fossiliferous diamicton. Two shell fragments collected at 167 m asl (marked by the figure at the left) provided dates greater than 34,000 BP (TO-1200, TO-1201) which are considered infinite.

been reworked by marine processes which have formed small beach berms and washed away some of the fines, leaving a gravel lag on the surface. One shell fragment from 134 m asl provided an AMS date of 29,380 $\pm$ 230 BP (TO-1198) which is interpreted as a minimum age of the sample.

The prominent lateral moraine along the west shore of South Bay has a conspicuously gentle gradient (1:100) (Fig. 4.6). This moraine extends to a kame/delta at 167 m asl which corresponds to the upper limit of fossiliferous diamicton 2 km further downvalley. Two alternative interpretations are provided for this moraine profile and related marine sediments.

The first interpretation proposes that a former valley glacier occupied South Bay depositing the gently sloping moraines and forming lateral meltwater channels within the valley and inner bay. This glacier was grounded and entrained "old" marine shells from the floor of South Bay and redeposited them in till along the slopes of Wolf Ridge and downvalley on the peninsula. This advance could have occurred as late as the last glaciation.

This interpretation explains the occurrence of diamicton on the lower slopes, the fragmentation of the shells through glacial transportation, and the lateral meltwater channels within South Bay. On the negative side, it does not explain the gradient of the moraine recording a former glacier surface profile which was subhorizontal (1:100). This profile becomes horizontal 2 km further downvalley between the kame/delta and the apex of the diamicton lobes flanking Wolf Ridge (Figs. 4.5, 4.7). Valley glaciers have a much steeper gradient than this;

for example, outlet glaciers in Arctic Canada have a minimum surface gradient of 1:22 within 20 km of their termini, which then steepens dramatically (Buckley, 1969). If the former ice in South Bay was near its terminus - as indicated by the elevation of the moraine above the adjacent lowlands (ca. 167 m) - the surface gradient of the ice (1:100) was far less than that expected for a grounded glacier. Finally, if this ice gradient was too low for ice flow to occur, it certainly would not provide the compressive flow required to redeposit shells from South Bay at its upper surface.

The second interpretation proposes that a former glacier occupying the valley contacted the sea at South Bay and was forced to float, forming a small ice shelf. Ice would have been grounded on the peninsula, draping the bedrock with fossiliferous diamicton and possibly depositing the quartzite erratics. Downvalley from the peninsula, the glacier would have encountered greater depths of water forming an ice shelf. This is suggested by the gentle gradient of the non-fossiliferous, lateral moraines which extend into the horizontal surface of the kame/delta and fossiliferous diamicton 2 km downvalley. Ice shelves are characterized by their low gradient, in contrast to grounded glaciers which have steep surface gradients near the terminus, as noted previously (Buckley, 1969; England et al., 1978; England, 1985).

Similar evidence for former ice shelves was recorded on northeastern Ellesmere Island, where steeply descending moraines become abruptly horizontal for 2 km and are fossiliferous downvalley of their apparent grounding line

(England et al., 1978). The process by which these shells were deposited is uncertain; however, they may have been transferred to the surface by the freezing-on of sea water at the base of the ice shelf accompanied by ablation at the surface. This has been well documented at the Ward-Hunt Ice Shelf where siliceous sponges, pelecypods, and other organics have been transported from the fiord to the debris ridge of the 40 m thick ice shelf (Lyons and Mielke, 1973; England et al., 1978).

The process by which the shells were deposited is equivocal, whether by grounded glacier ice or by an ice shelf in South Bay. However, the gradient of the former ice surface was clearly too gentle for a grounded glacier and supports the interpretation of a former ice shelf within South Bay. A schematic diagram showing the plan view of such an ice shelf in South Bay is shown in Figure 4.10.

The radiocarbon dates on the shells associated with the ice shelf are all > 29,000 BP and are interpreted as minimum ages. If the shells are contemporaneous with the ice shelf, then the radiocarbon dates provide a minimum age of ca. 29,000 BP for this glaciation. However, the shells may predate the ice shelf, further diminishing the usefulness of the dates. The lack of an absolute age for the shells and the uncertainty of the depositional process prevents an unequivocal interpretation of these deposits.

The depth of water required to float an ice shelf can be determined from the thickness of the ice and the ratio of ice : water densities (0.88 : 1). The thinner the ice becomes, the less freeboard there is because the relative proportion of

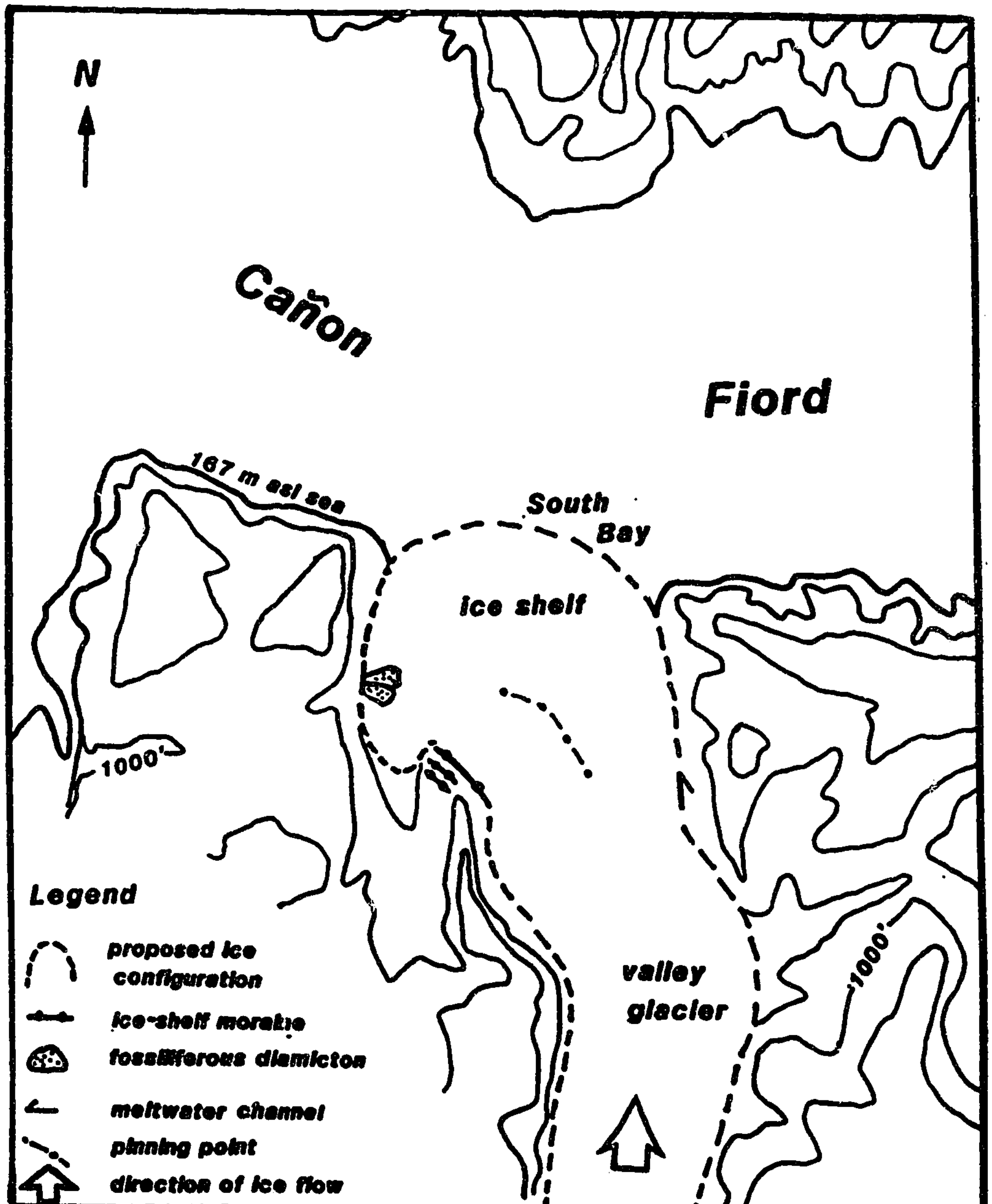

Fig. 4.10. Schematic map illustrating an ice shelf in South Bay which is fed by a trunk glacier in the valley to the south of South Bay. Contour interval 500 feet or 152 m.

ice above sea level decreases based on the ratio of the densities of ice to water. In South Bay the maximum ice thickness was ca. 167 m based on the elevation of the subhorizontal moraine. The relative sea level required to float a 167 m thick ice shelf would be 147 m asl based on the surface elevation of the ice and the amount of freeboard (20 m). If the ice shelf was only 25 m thick, it would have only 3 m of freeboard, requiring a relative sea level 164 m asl. Therefore, the sea level associated with the former ice shelf in South Bay must have been at least 147 m asl, with a possible maximum of ca. 164 m asl.

The relative sea level associated with the former ice shelf in South Bay at 147 to 164 m asl is 28 to 45 m higher than the local Holocene marine limit surveyed in outer South Bay (119 m asl). Therefore, the amount of glacioisostatic loading associated with this trunk glacier and its adjoining ice shelf distinguish it from the Holocene marine limit (119 m asl) and therefore the last glaciation.

The juxtaposition of the 119 m sea level dated 7930 ± 70 BP and the evidence for a higher sea level (147 - 164 m asl) within South Bay suggests that two discrete intervals of glacioisostatic loading were responsible for their formation. For example, the former ice shelf could not have been formed in contact with the 119 m sea because this water depth would have been inadequate to float the observed ice thickness. Furthermore, it is not likely that the profile of the Holocene marine limit would slope upvalley from 119 m to ≥ 147 m asl over a distance of 5 km. This would require a slope of

$\geq$ 5.6 m/km which is inconsistent with the gradient observed throughout the area (less than 1 m/km in Canon Fiord; 0.73-0.85 m/km in northwest Ellesmere Island (Evans, 1990)). Also, the higher sea level (147 to 164 m asl) within South Bay does not correspond to the 148 m marine limit at the mouth of Canon Fiord because the latter is dated 8290 $\pm$ 100 BP (GSC-4626) to 8850 $\pm$ 60 BP (TO-1285) (England, 1990) and is in fact similar in age to the 119 m sea level in South Bay (7930 $\pm$ 70 BP).

4.5 Summary

The geomorphology and raised marine deposits on Wolf Ridge indicate that during the last glaciation, the ridge was covered by an ice cap which was coeval with a 119 m sea at ca. 7930 $\pm$ 70 BP. During an older, more extensive glaciation, a valley glacier drained into South Bay where it formed an ice shelf in a former relative sea level 147 - 164 m above present. Evidence also shows that a trunk glacier once occupied at least the inner half of Canon Fiord. The amount of glacioisostatic unloading associated with the older glaciation (147 - 164 m in South Bay) exceeds that recorded by the Holocene marine limit (119 m asl); however, the age of this older glaciation is unknown.

Chapter 5

Geomorphology and Raised Marine Deposits: Foster Creek

5.1 Introduction

This chapter presents the glacial geomorphology and raised marine deposits of Foster Creek and its adjacent plateaus. The observations include two segments of a section exposed along Foster Creek and two radiocarbon dates on related marine sediments. Collectively, the evidence is used to reconstruct the glacial and sea level history of Foster Creek.

5.2 Foster Creek and Adjacent Plateaus

Foster Creek drains two outlet glaciers at the western extremity of the Agassiz Ice Cap, central Ellesmere Island (Figs. 2.1, 5.1, 5.2). The plateaus adjacent to Foster Creek are ice-free apart from a small cirque glacier to the southeast (Fig. 5.1). Erratics including quartzite, gneiss, and granite were observed on the plateaus surrounding Foster Creek. The provenance of these erratics is likely eastern and central Ellesmere Island, recording regional dispersal by late Tertiary rivers and/or subsequent glaciation(s). Five kilometres to the south of Foster Creek, a deep, V-shaped valley dissects the plateau along strike; this valley and its tributaries are interpreted as meltwater channels of a former plateau ice cap (Figs. 5.1, 5.3).

On the plateau immediately north of Foster Creek, a suite of well-developed, parallel gullies cross the bedrock strike (Figs. 5.3, 5.4). These are interpreted as

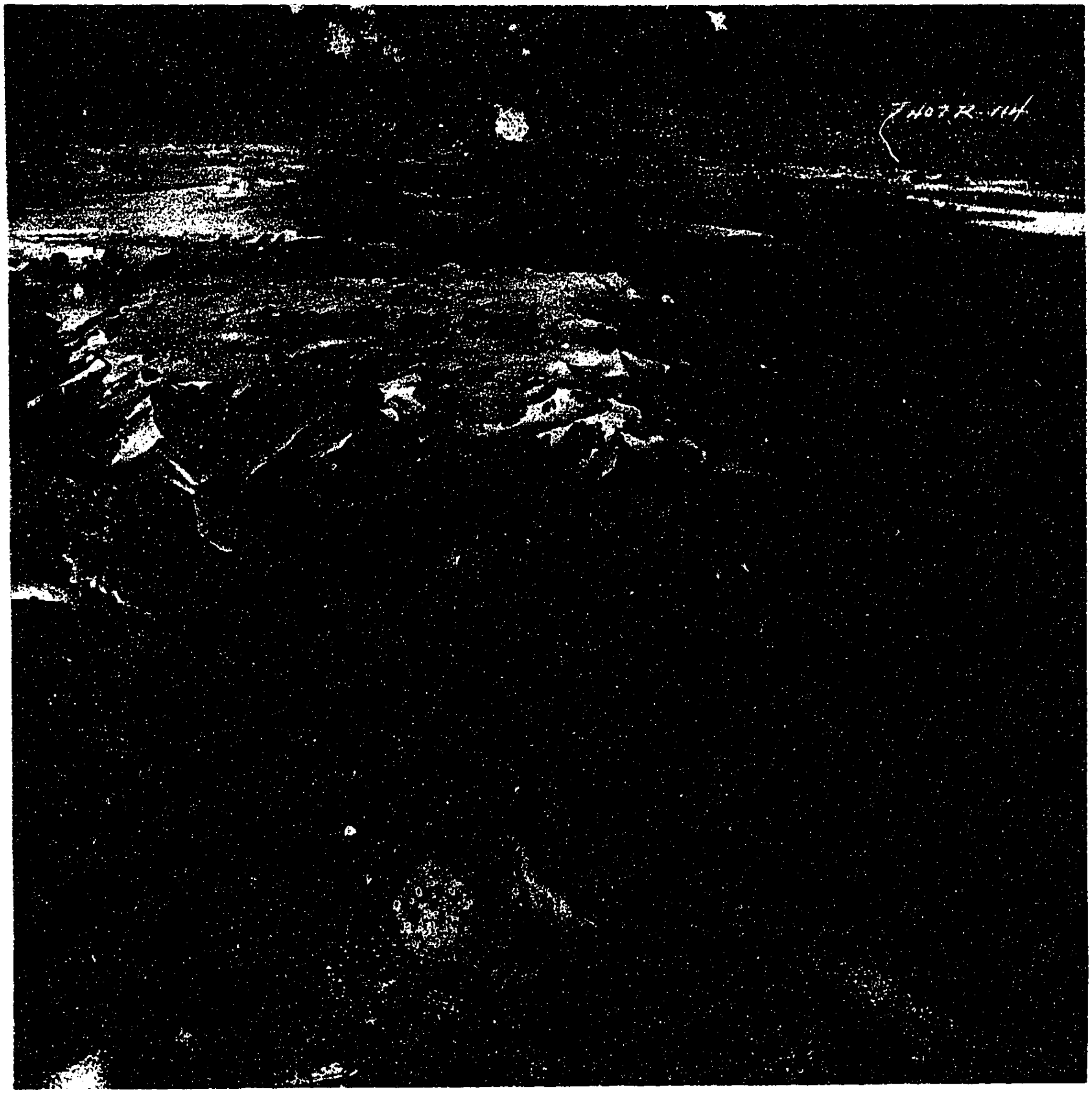

Fig. 5.1. Oblique airphoto of Foster Creek, east-central Canon Fiord.

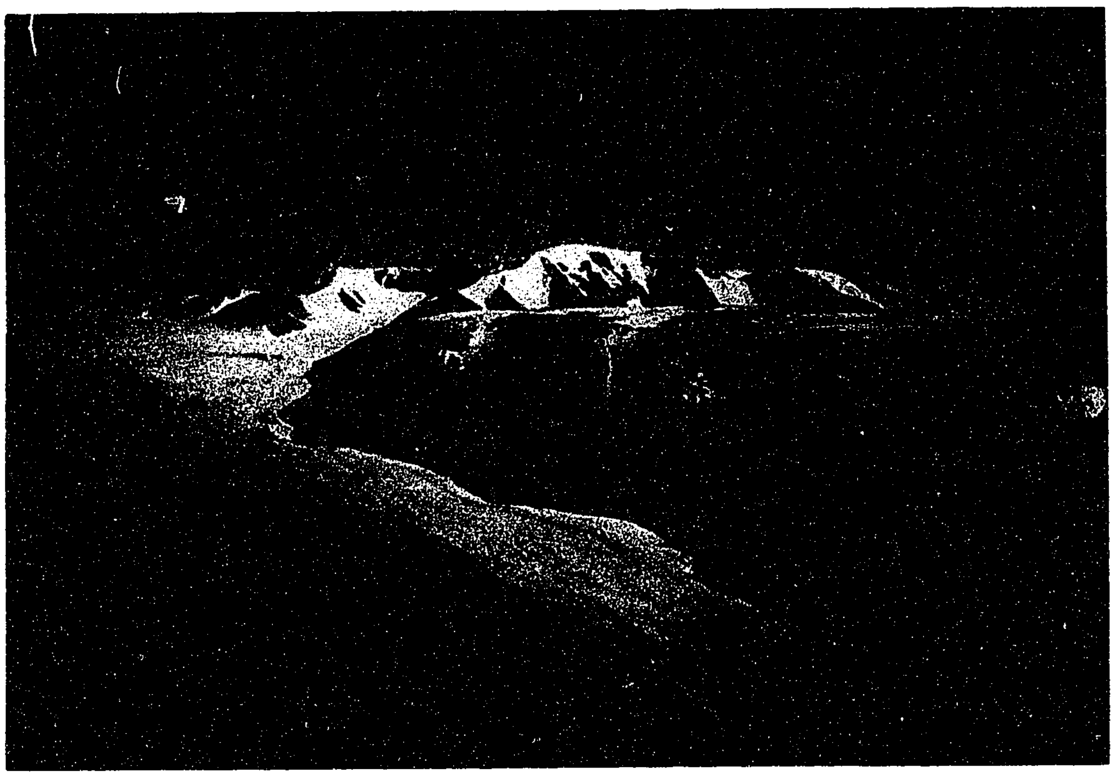

Fig. 5.2. One of two outlet glaciers of the Agassiz Ice Cap which drains into Foster Creek.

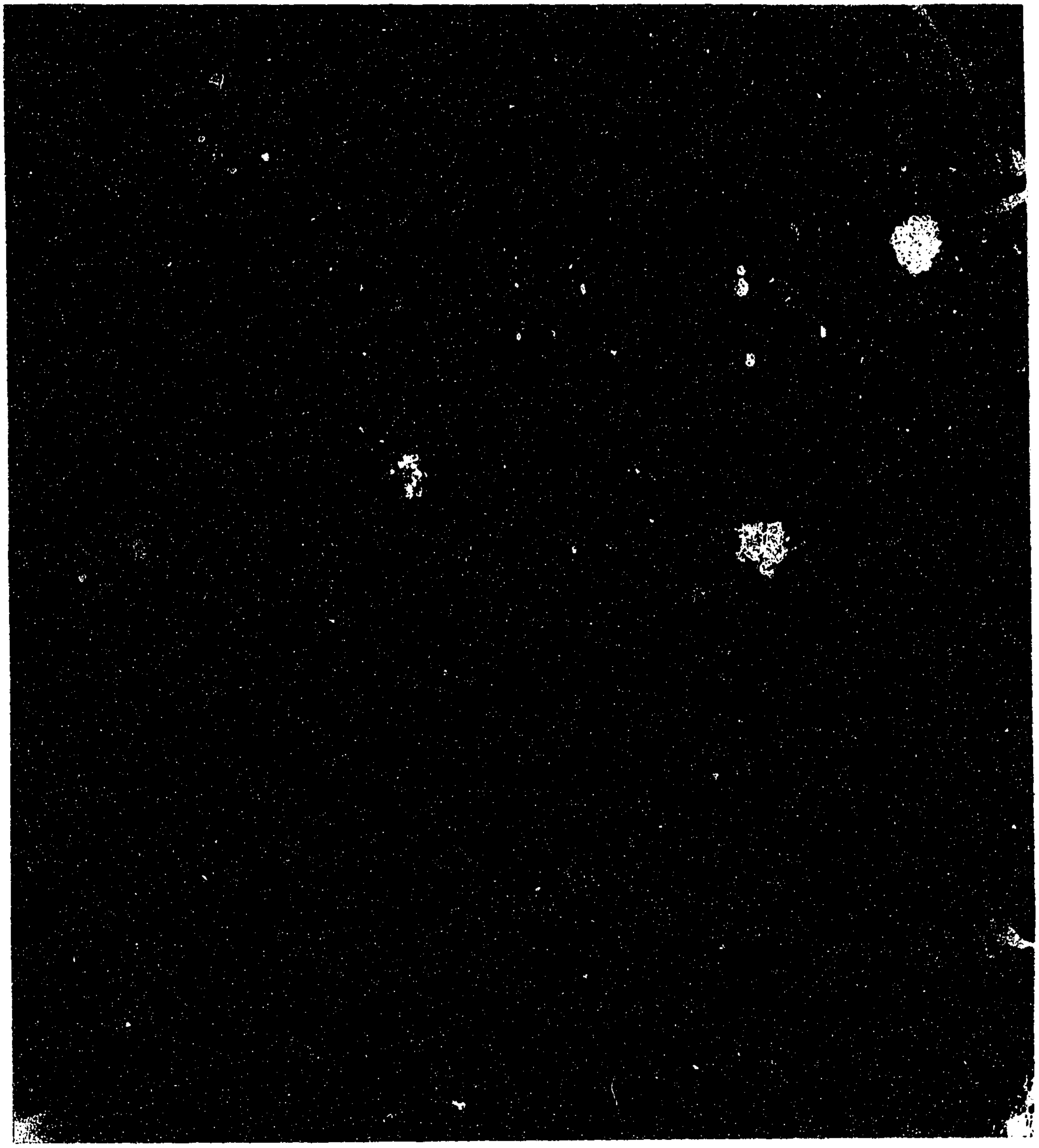

Fig. 5.3. Vertical airphoto of Foster Creek, central Canon Fiord, showing former meltwater channels (barbed arrows), moulded bedrock ( ), raised marine deltas (triangles), silts (dotted areas), terrace ( ), locations of shell samples, and sites 1 and 2, discussed in text.

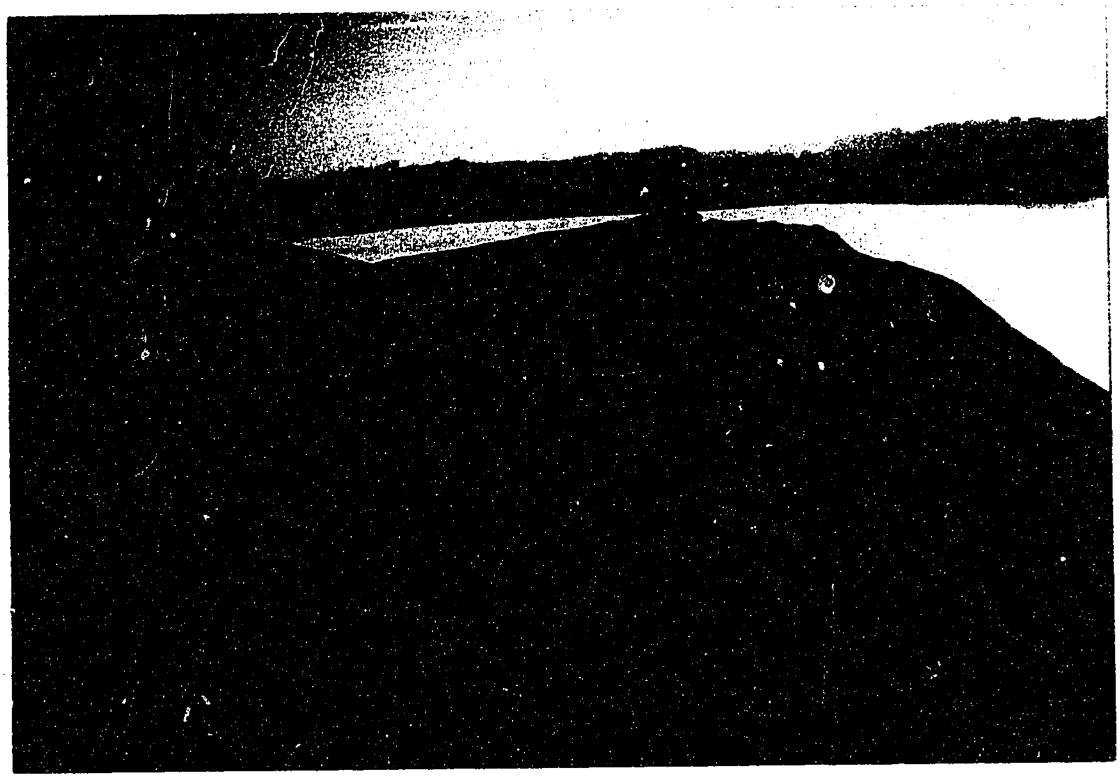

Fig. 5.4. View southwards across the plateau surrounding Foster Creek (obscured in valley crossing lower photo) with Canon Fiord and Fosheim Peninsula in background. Note the prominent meltwater channels in foreground (600 m asl) marking the retreat of a former valley glacier in Foster Creek. The arrow marks the location of streamlined features on the north-facing slope (see Figs. 5.5, 5.6).

meltwater channels, recording the former margin of a retreating valley glacier in Foster Creek. The elevation of these channels (600 m asl) indicates a former ice thickness of 470 m in the upper valley.

On the plateau to the south of Foster Creek, streamlined landforms oriented downvalley occur in exposed limestone at ca. 350 m asl (Figs. 5.4, 5.5, 5.6). The smoothly moulded bedrock is truncated sharply on the downvalley (west) side, exposing ragged, broken rock. These are interpreted as roche moutonées and/or subglacial meltwater features formed by or beneath the valley glacier occupying Foster Creek (Sugden and John, 1976; Shaw, 1987; Shaw et al., 1989).

Within the valley (Fig. 5.7), fluvial incision has isolated a bedrock knoll which bears further evidence of glacial flow (Figs. 5.3, 5.8). Exposed bedrock on the knoll is streamlined, striated, and grooved indicating former ice flow downvalley (Figs. 5.9, 5.10). The surface of the knoll is marked by a prominent gravel terrace which is interpreted as a wave-washed surface recording a former relative sea level of 130 m asl (Fig. 5.11).

5.3 Lower Valley of Foster Creek

One km to the west of the bedrock knoll, Foster Creek opens from a narrow, steep-walled valley to a broad, open foreland. Here a small, raised marine delta flanks the valley's south-facing slope marking the local marine limit at 131 m asl (Figs. 5.3, 5.8, 5.12). The elevation of 131 m is consistent with the gradient connecting the marine limit observed by Hodgson (1985) to the southeast and by

Fig. 5.5. Smoothly streamlined limestone on the plateau adjacent to Foster Creek moulded by subglacial meltwater or glacier ice indicates former flow in a downvalley direction.

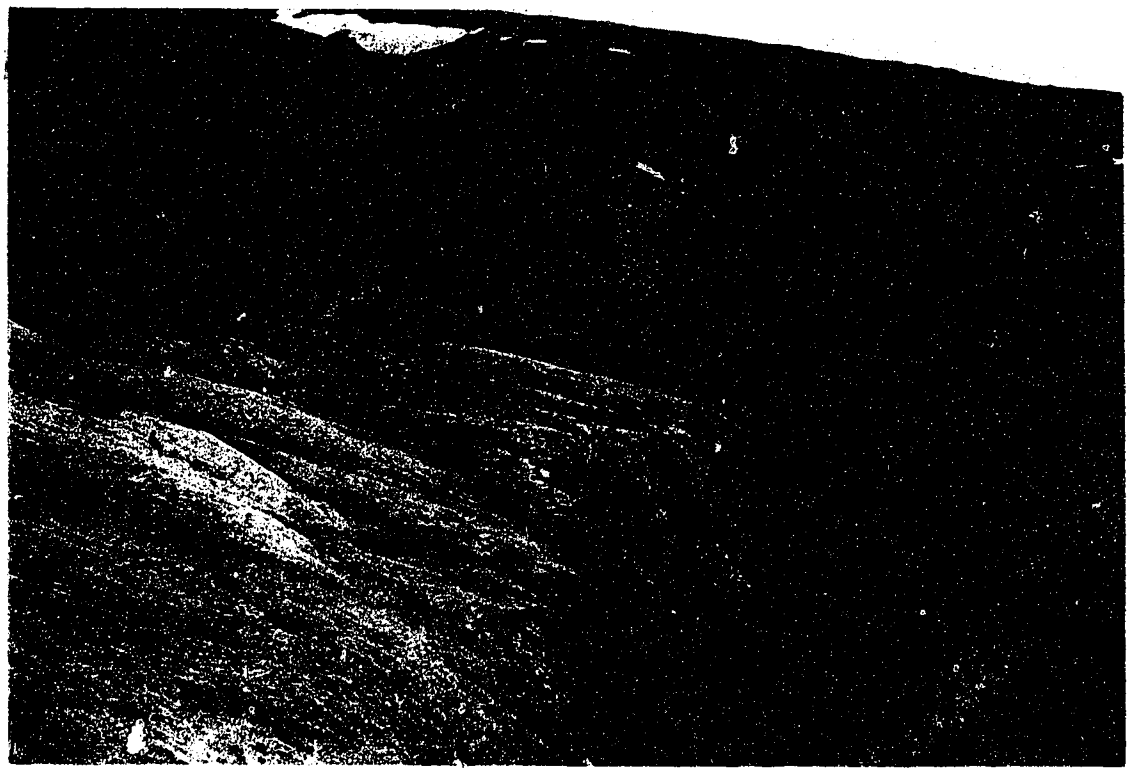

Fig. 5.6 Streamlined limestone forms on plateau are sharply truncated on their downvalley side, suggesting that they are roche moutonées. Notice boundary between limestone and dolomite bedrock in the background.

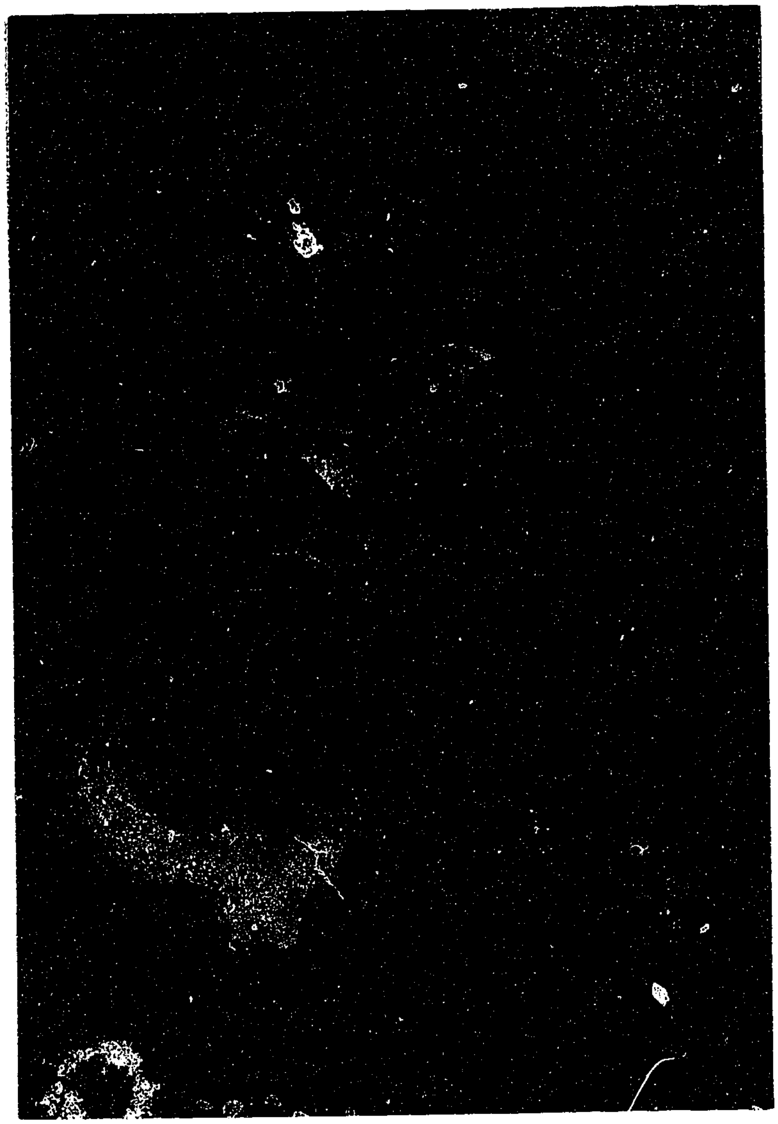

Fig. 5.7. The upper valley of Foster Creek which sits 500 m below the plateau surface. An unnamed outlet glacier of the Agassiz Ice Cap is seen at the head of the valley in the distance.

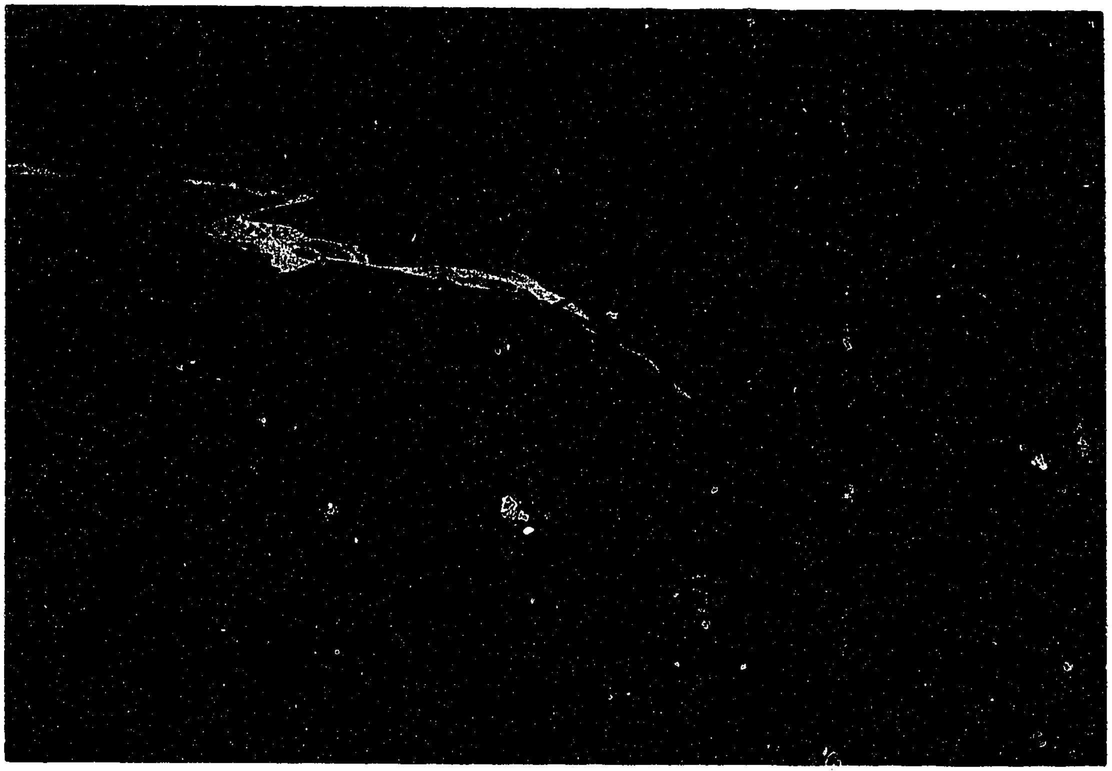

Fig. 5.8. The bedrock knoll created by river incision at mid-valley, Foster Creek (hollow arrow). Figures 5.9, 5.10 are photos from the surface of the knoll. A small raised marine delta marking marine limit (131 m asl)is seen in the distance (black arrows).

Fig. 5.9. Moulded limestone bedrock on knoll indicating former ice flow downvalley at Foster Creek.

Fig. 5.10. Glacial striae on the moulded limestone shown in Fig. 5.9.

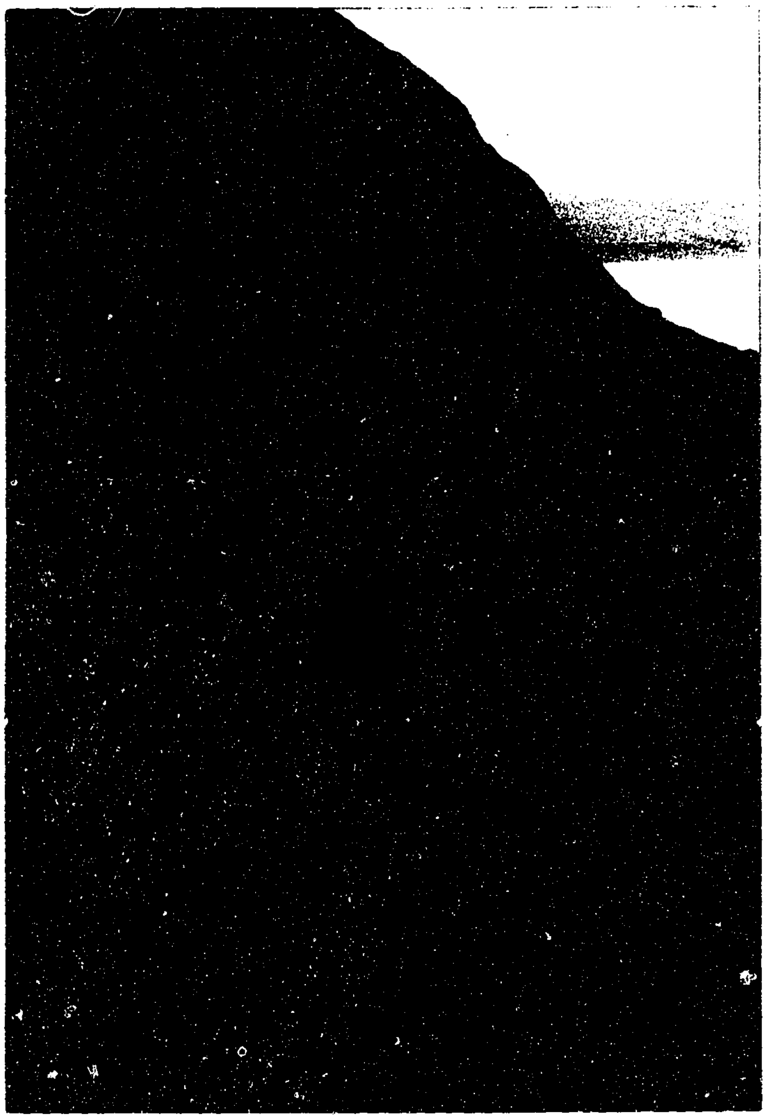

Fig. 5.11. The prominent gravel-covered terrace, on the bedrock knoll, interpreted as a wave-washed surface marking a former sea level of 130 m asl. Note the exposed bedrock in the left foreground of the terrace where Figures 5.9 and 5.10 were taken.

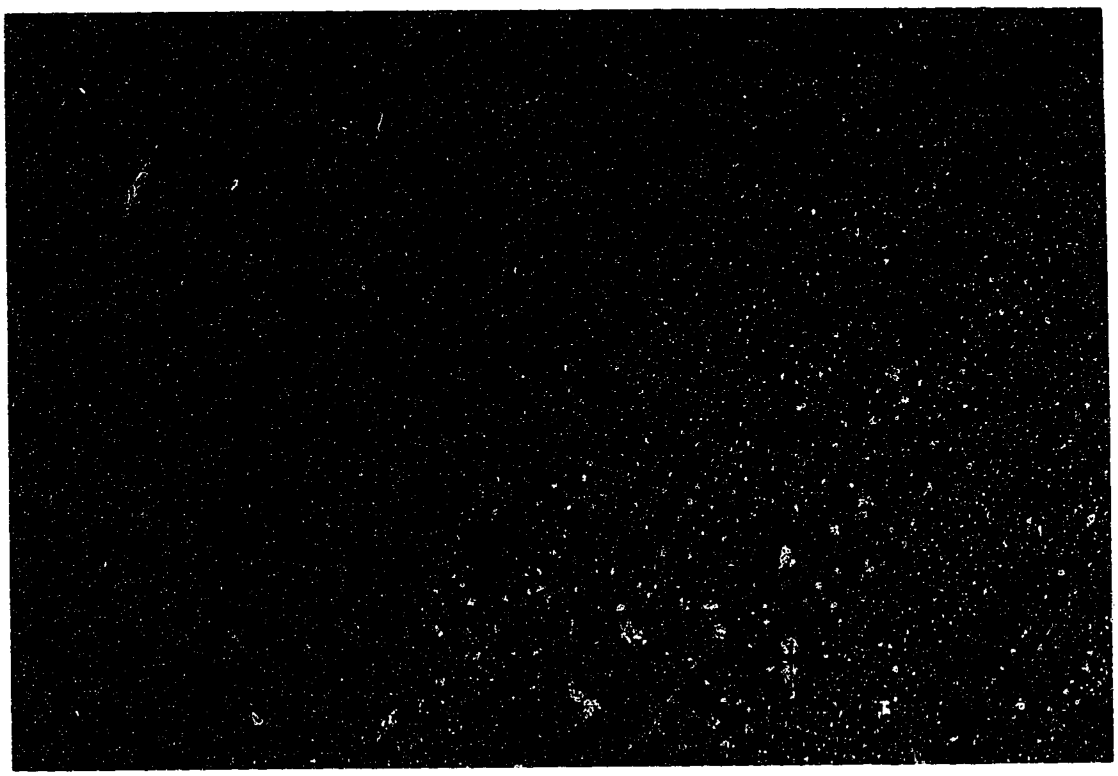

Fig. 5.12. The raised marine delta on the south-facing slope of Foster Creek valley which represents local marine limit at 131 m asl (arrow). Accordant (paired) surfaces occur on either side of the central ravine. The stream valley that fed the delta lies on the boundary of Sverdrup Basin (left) and Franklinian Mobile Belt (light) sediments.

England (1990) to the north.

In the lower valley, postglacial emergence has exposed a large marine delta to subaerial erosion, revealing sediments along the outermost 2 km of the valley (Figs. 5.13, 5.14). Two sections along the contemporary floodplain are presented below, beginning with the more coastal site (Figs. 5.1, 5.15). The stratigraphic units are described for each site and an interpretation of the associated depositional environments is presented.

5.4 Site 1: Description

Site 1 is located ca. 1 km from the present-day fiord coast (Figs. 5.1, 5.3). The section is ca. 60 m high and has been subdivided into five units (Figs. 5.15, 5.16). These units are exposed intermittently along 400 m of the section, but they are best exposed at site 1. Two stratigraphic columns are presented, for the upstream and downstream ends of the exposure at site 1 (Fig. 5.15).

Unit A - This unit consists of steeply dipping beds of rounded to sub-angular cobbles, gravel, and sand (Figs. 5.15, 5.16, 5.17). Unit A is ca. 10 m thick. The beds dip 23^0 downvalley and are differentiated primarily by colour, which ranges from yellow-beige to dark grey. The lithologies include locally derived red sandstone, and limestone and dolomite which originate in upper Foster Creek. The dipping beds of unit A are truncated by unit B forming an erosional contact at 26.5 m asl (Fig. 5.18). Unit A is not present at the upstream section of site 1;

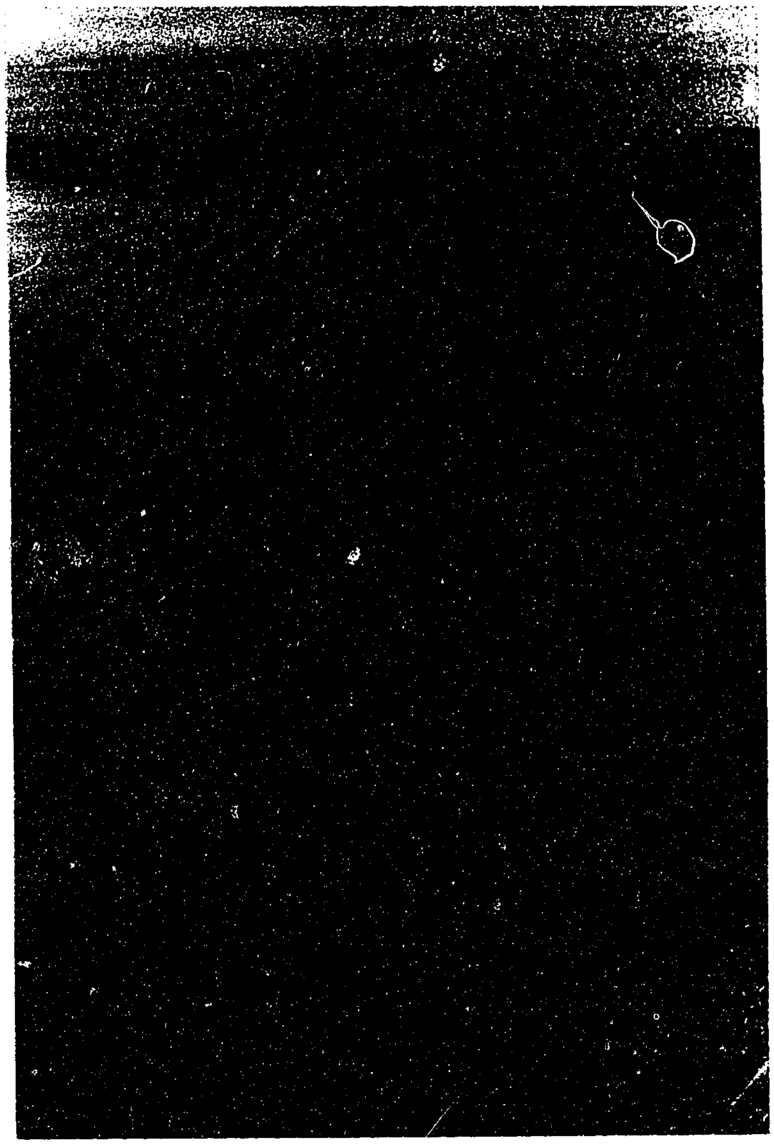

Fig. 5.13. An oblique view looking over the raised delta remnants in outer Foster Creek and the delta which is presently being deposited into Canon Fiord. Notice the brownish grey colour of the delta deposits in contrast to the local red sandstone bedrock. Caledonian Bay is seen in the distance. Fig. 5.14 is a close-up the delta on the south side of Foster Creek.

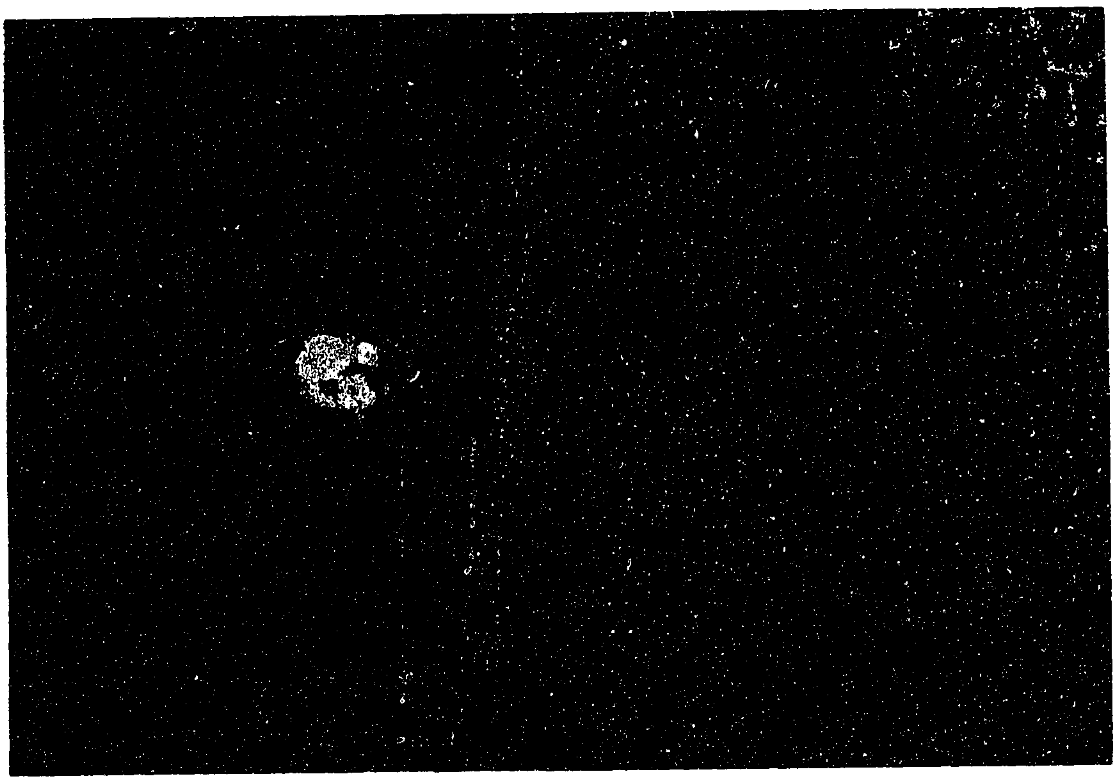

Fig. 5.14. Raised marine delta surfaces in outer Foster Creek which have been subjected to subaerial erosion, exposing a section along the river. The prominent delta surface in centre of photo is 49 m asl.

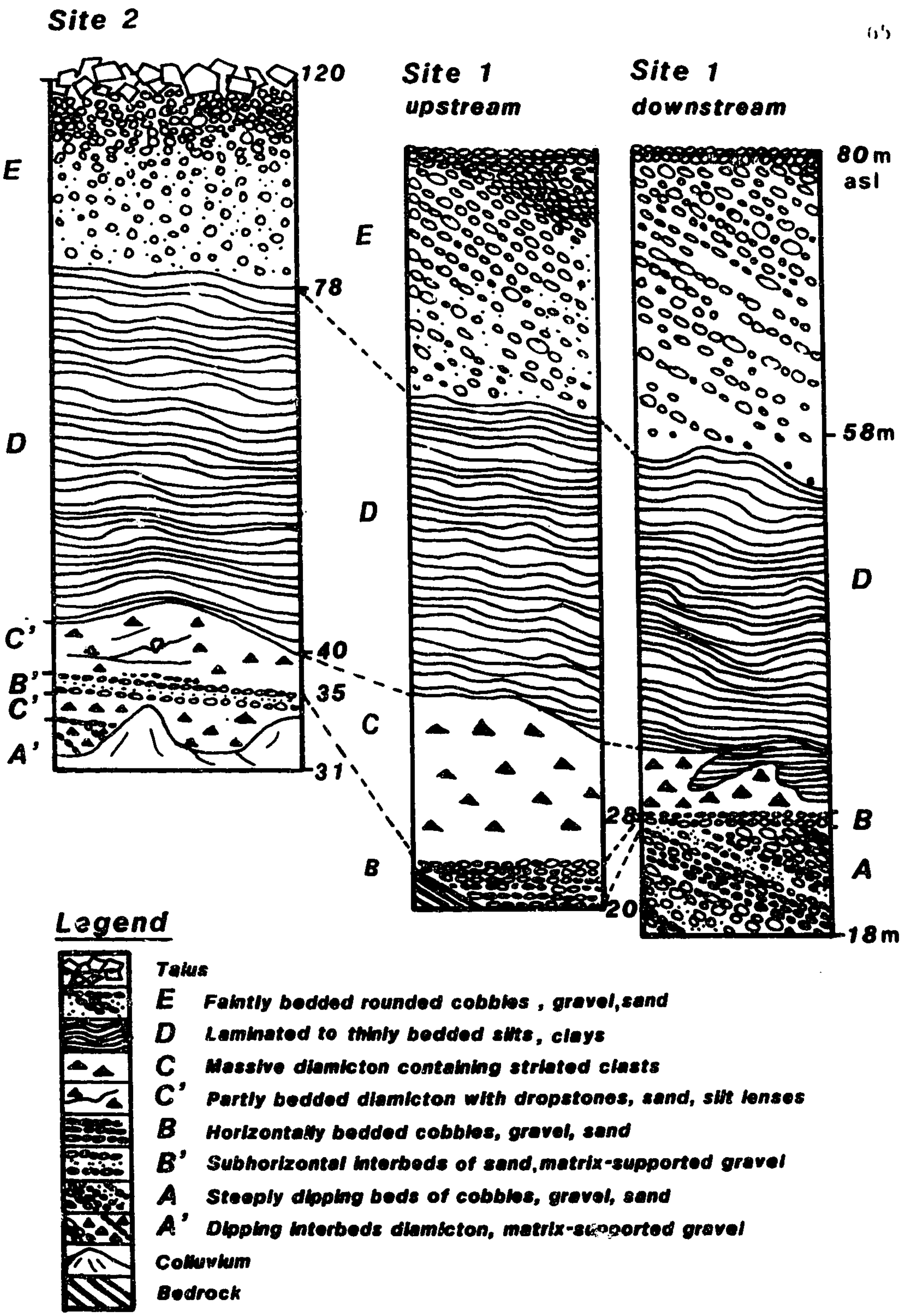

Fig. 5.15. Stratigraphic columns of site 1 (upstream and downstream) and site 2. The locations of the sites are shown in Figs. 5.1, 5.3.

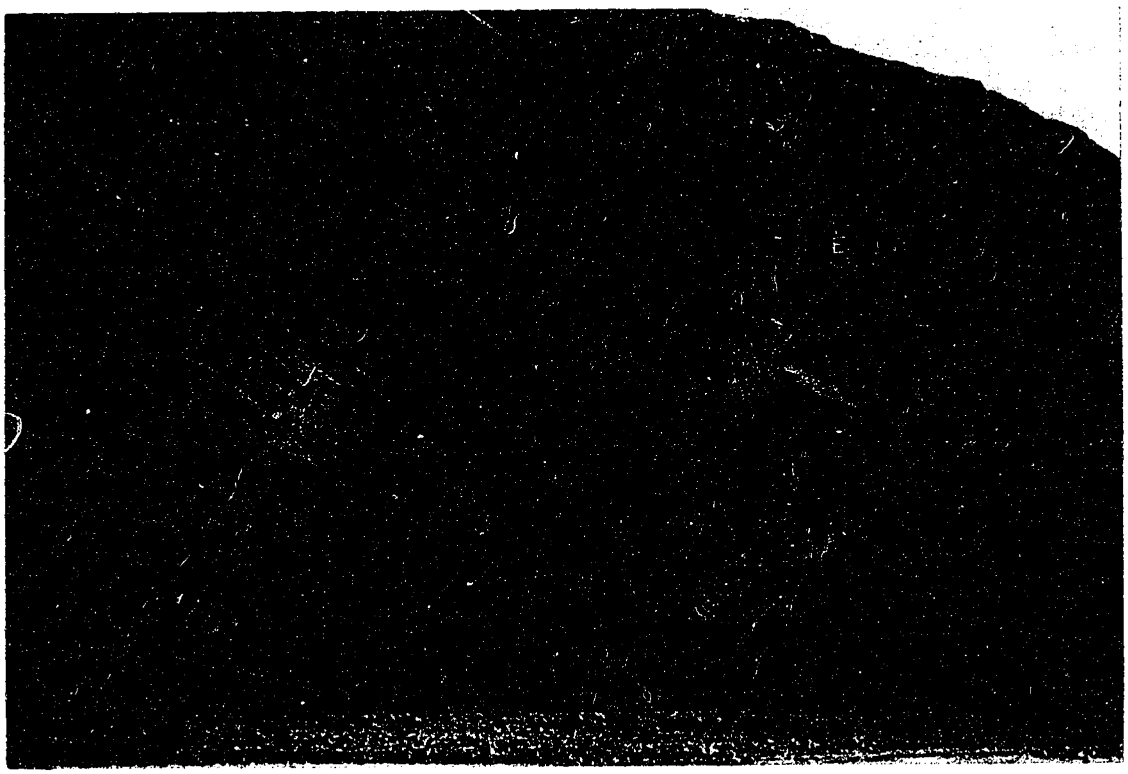

Fig. 5.16. Site 1, located 1 km from the coast, reveals a good exposure of units A and B which are interpreted as deltaic foreset beds overlain by topset beds. Unit B at 28 m asl represents a preglacial sea level. Units A through E are shown with arrows indicating contacts. Note the dipping beds in unit E which are interpreted as deglacial deltaic foresets.

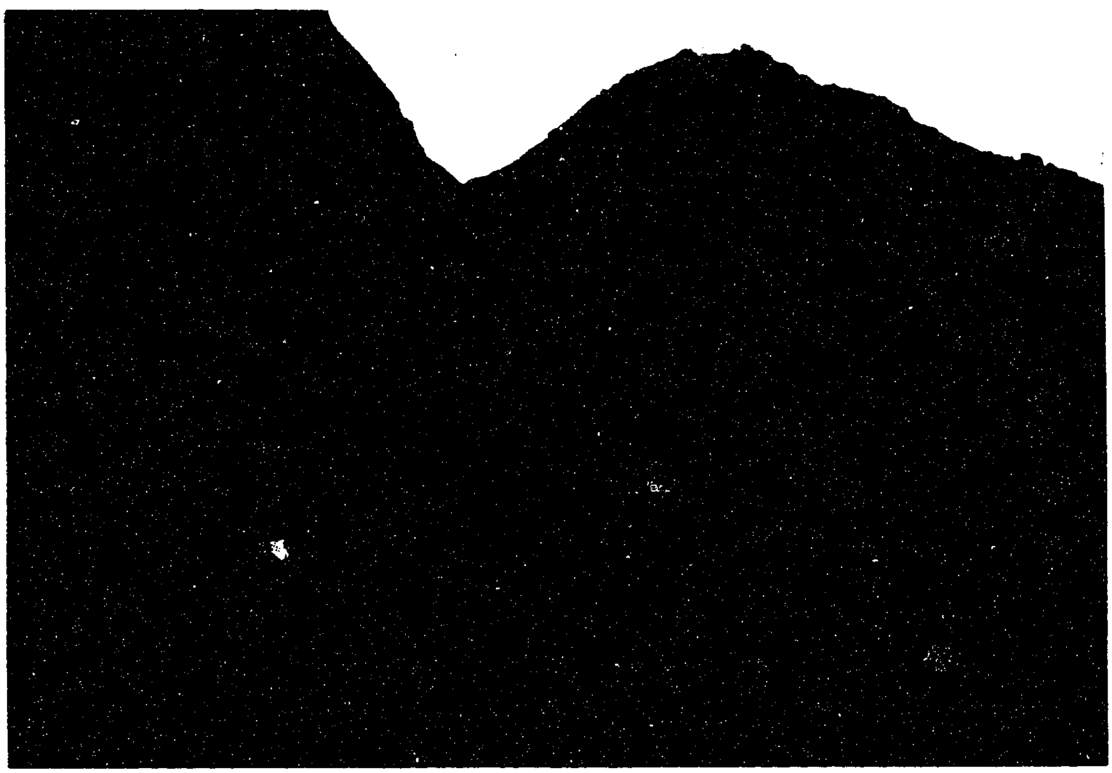

Fig. 5.17. A good exposure of unit A interpreted as deltaic foreset beds. Although unit B cannot be seen here, the horizontal, rust coloured bed can be seen in Fig. 5.16 which is taken to the left of this.

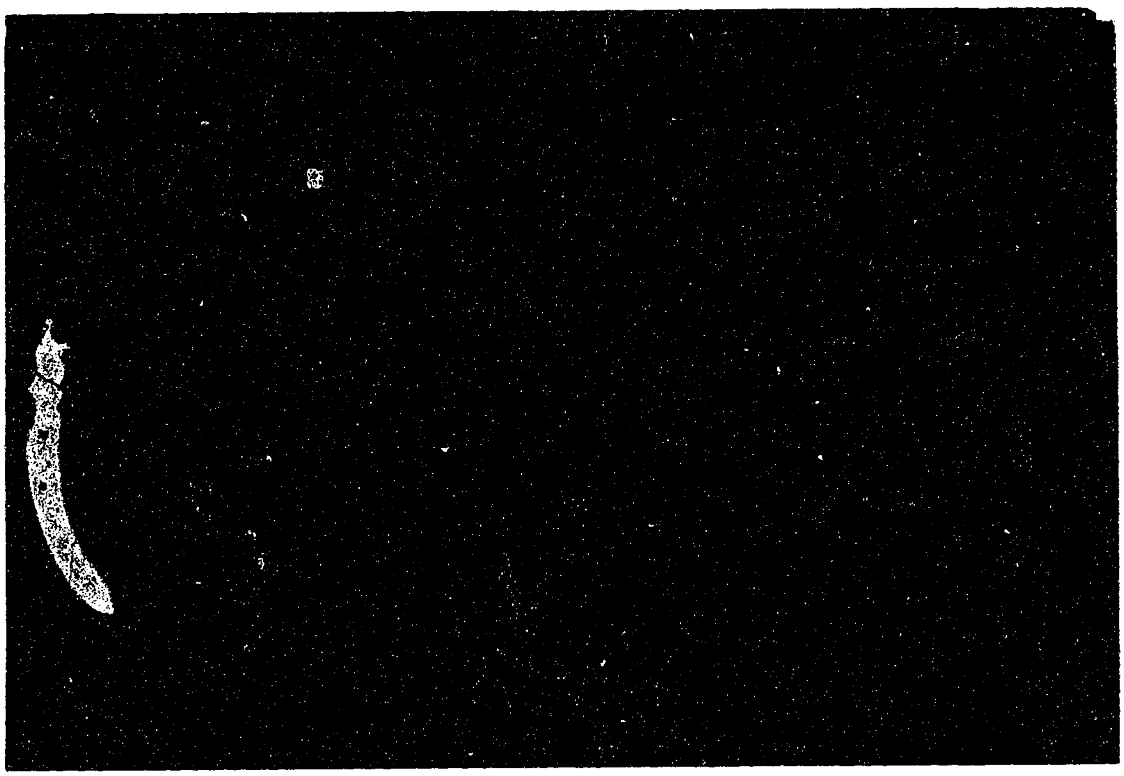

Fig. 5.18. Photograph of site 1 illustrates the contact between the dipping beds of unit A and the horizontal beds of unit B. Note the truncation of dipping beds at left of photograph.Units C and D are also visible. A rock hammer (circled) provides scale.

instead unit B outcrops at the base of the section (Fig. 5.15).

Unit B - This unit consists of horizontally bedded, rounded to sub-angular cobbles, gravel, and sand, and is lithologically similar to unit A which it overlies unconformably (Figs. 5.15, 5.18). Unit B is 1.5 m thick at the downstream exposure and is overlain by unit C at 28 m asl. Unit A and B outcrop together for ca. 300 m along the section, but further downstream they thin until they no longer outcrop where unit C extends to the base of the section.

At the upstream exposure, unit B outcrops at the base of the section between 20 and 24 m asl where it occupies a small bedrock syncline (Figs. 5.15, 5.19). Subhorizontal beds within unit B are truncated by the overlying unit, C, at 24 m asl (Fig. 5.20).

Unit C - This unit consists of grey to brownish grey (10 YR 6/1 d) massive, calcareous, silty matrix diamicton which contains striated, angular to sub-rounded, limestone and dolomite clasts. These lithologies outcrop upvalley in the Mid Paleozoic Allen Bay Formation and Cornwallis Group. Unit C varies considerably in thickness (0.1 to 13 m) (Figs. 5.15, 5.18, 5.20). This diamicton outcrops to within 500 m of the present coast, where it occurs at the base of the section. A fabric recorded by clasts within this diamicton has a preferred orientation parallel with the valley (east-west) and an upvalley dip (Figs. 5.21a). The diamicton is conformably overlain by, and in places grades up into, silts and

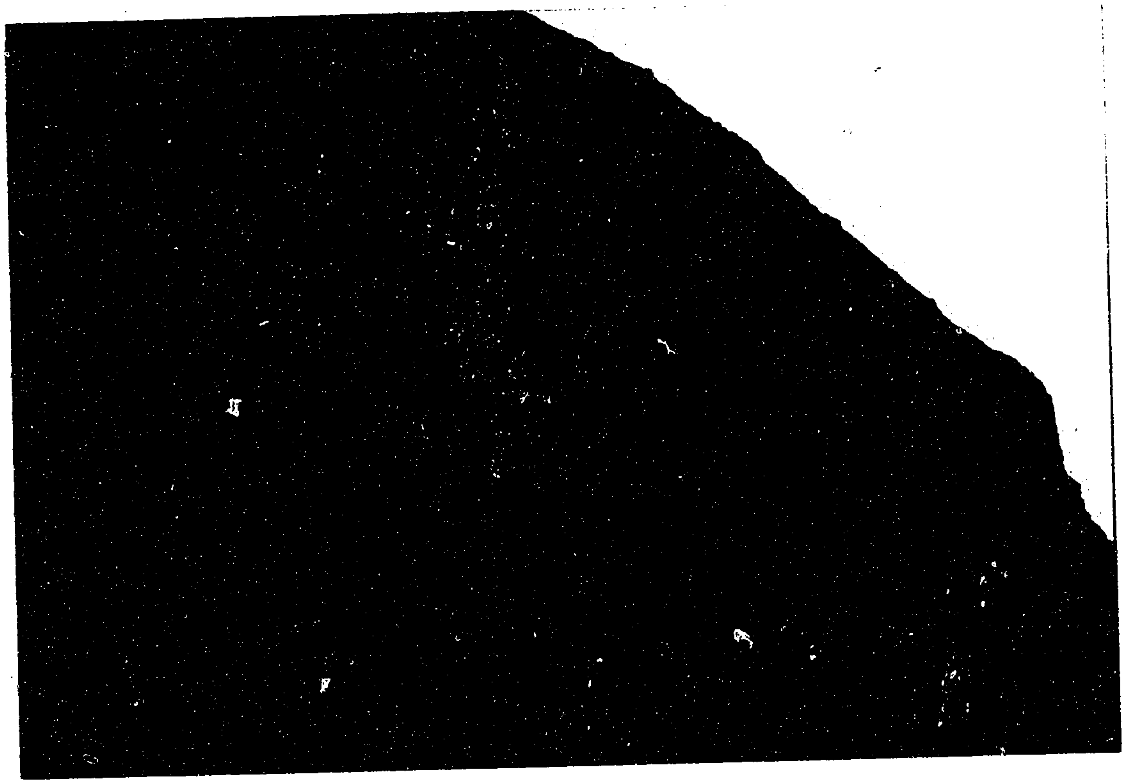

Fig. 5.19. Units B and C (upstream, site 1). Note the subhorizontal, discontinuous beds in unit B which are truncated at their upper boundary by unit C. The contact between the two units is 24 m asl and may represent a former sea level.

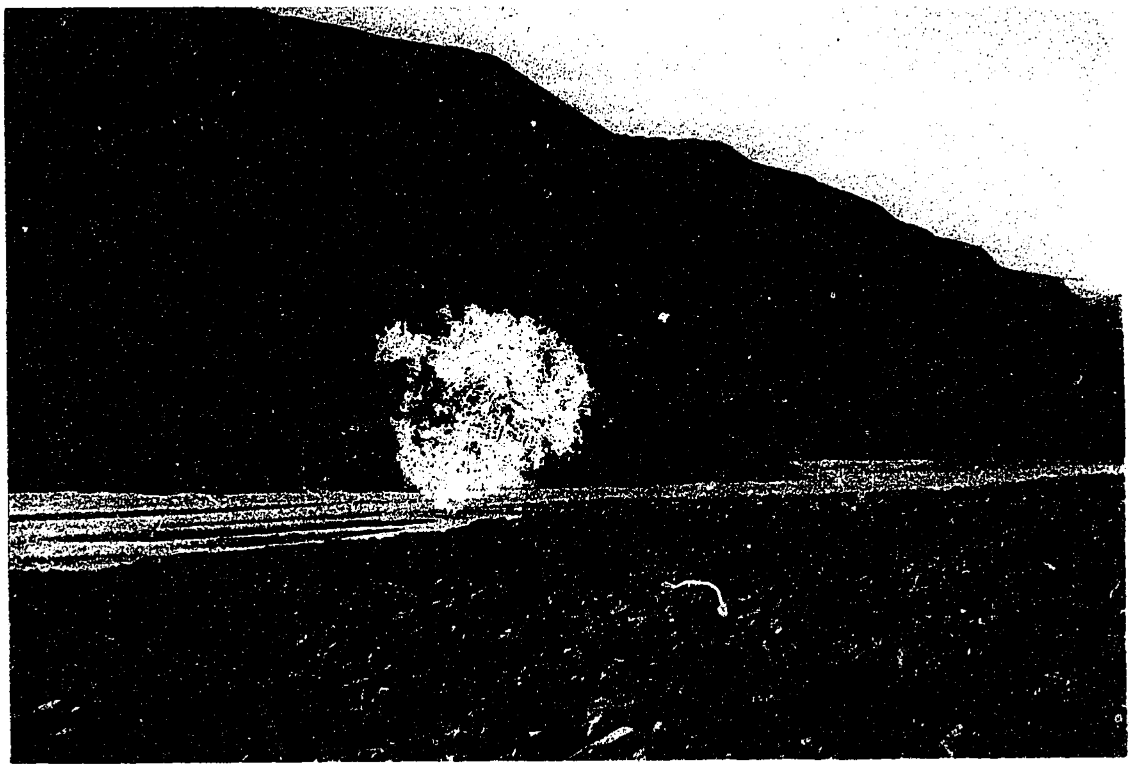

Fig. 5.20. Unit B occupies a small syncline at the upstream site and is overlain by 13 m of diamicton, unit C.

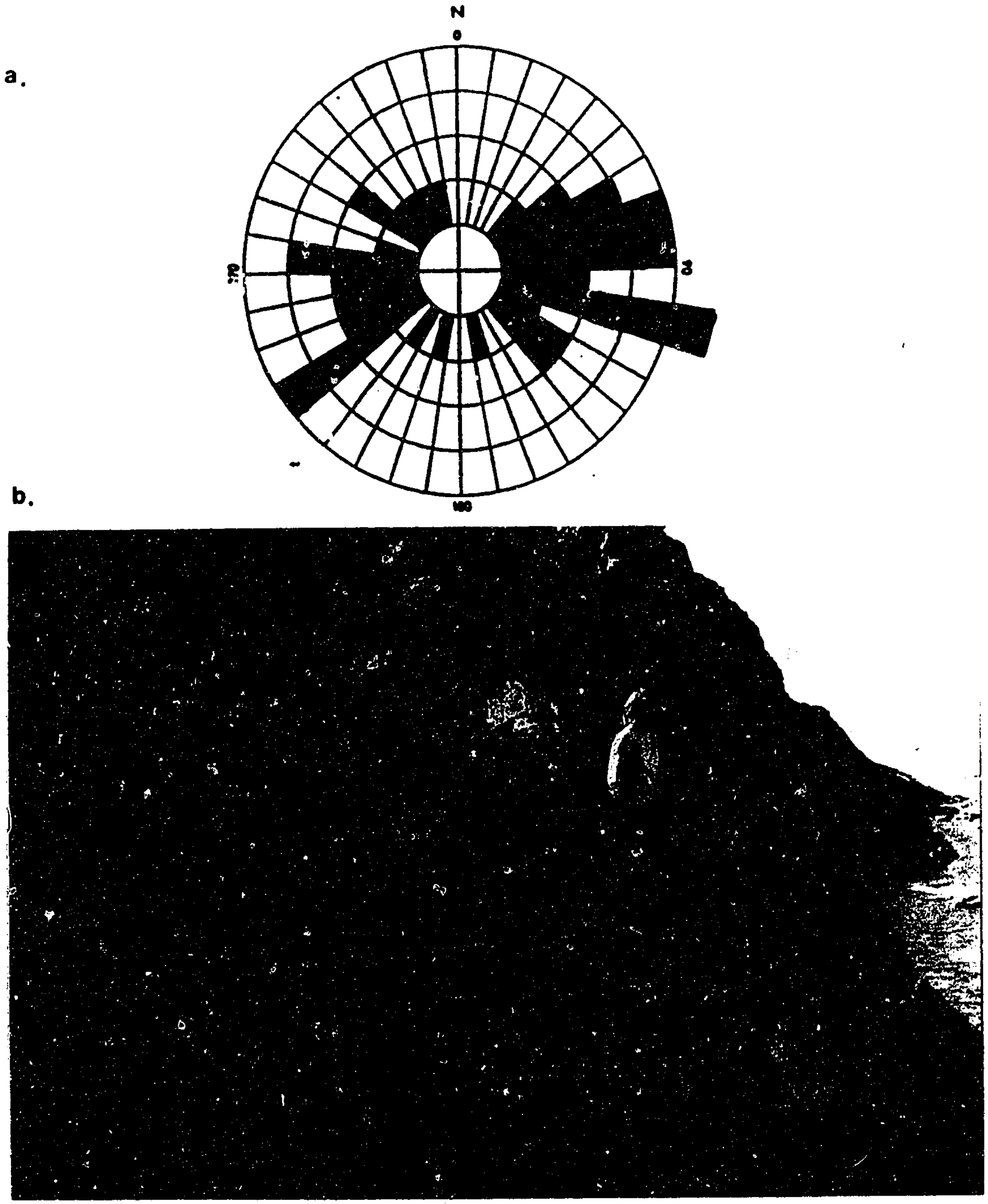

Fig. 5.21 (a) Orientation of clasts from fabric measured in unit C. (b) Unit C, which originates upvalley is conformably overlain by unit D. Results of a fabric measured in unit C 100 m downvalley indicate a preferred east-west orientation parallel to the valley with an upvalley dip. The diamicton is interpreted as a basal till. Trixa Marshall for scale.

clays of unit D (Fig. 5.21b). At the downstream end of the exposure, 4 m of diamicton is interfingered with a silt-clay interbed of unit D (1 m thick and 4 m wide) (Fig. 5.15). Upstream, unit C is 13 m thick (Figs. 5.15, 5.20).

Unit D - Unit D consists of dark brown to brownish grey (10 YR 5/1 m), horizontally- and thickly-laminated to thinly-bedded silts and clays (Figs. 5.21b, 5.22, Appendix: Table 1). The individual beds are characterized by massive clay and silt (> 25% clay) and are barren of organic deposits. This unit conformably drapes the underlying diamicton (Fig. 5.21b). Unit D is ca. 30 m thick, outcropping between 30 and 60 m asl, although its upper contact is concealed by colluvium (Fig. 5.16). Unit D extends 400 m downvalley from site 1 to the western end of the section. Here the barren, dark silts and clays grade upward into light grey (10 YR 7/1 d), fossiliferous silts, which are also laminated to thinly-bedded (Fig. 5.23). Shell fragments of Mya truncata collected at this boundary occur at 23 m asl. One fragment submitted to ISOTRACE provided an AMS date of 7570 ± 60 BP (TO-1199).

Unit E - Unit E consists of 25 m of non-cohesive, rounded, coarse sediments which coarsen upwards from sand and fine gravel to gravel and cobbles. The lithologies include limestone and dolomite. Although this unit is partially obscured by colluvium, faint cross-bedding is observable (Fig. 5.16). The steepest dip (observed in an exposure oriented downvalley) is ca. 20^0.

Fig. 5.22. Barren, laminated to thinly bedded silts and clays which make up much of unit D. Tape measure is 50 cm long.

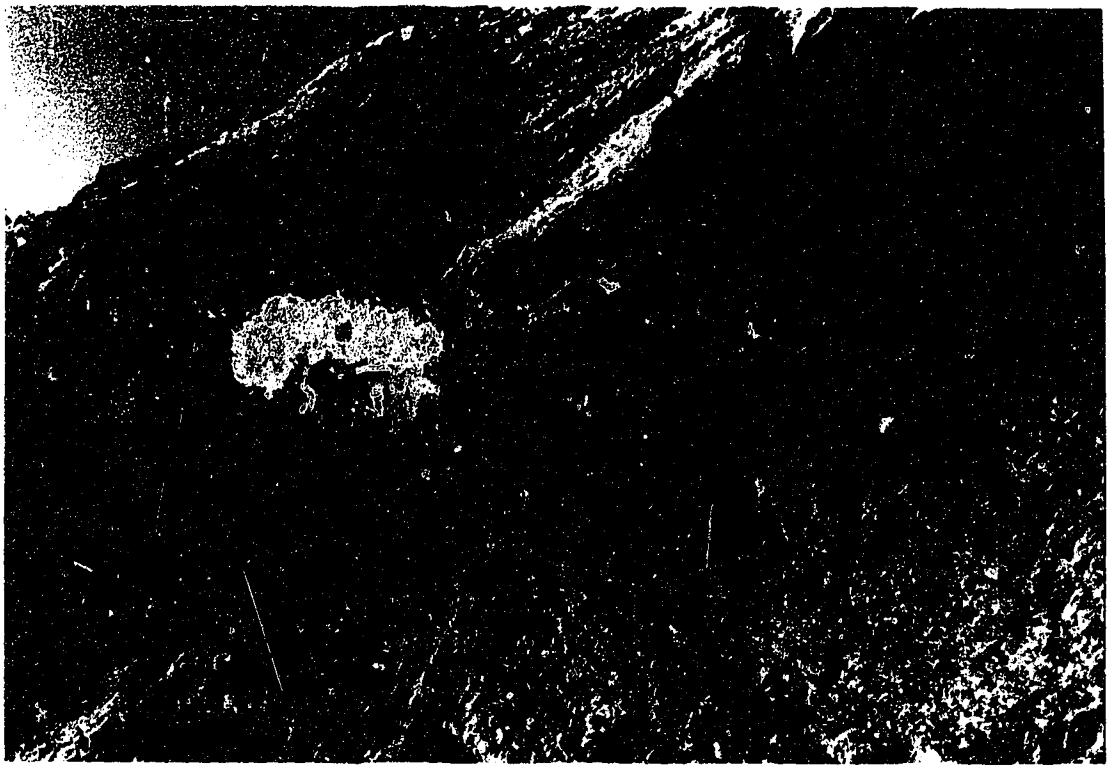

Fig. 5.23. Light grey, laminated to thinly bedded silts containing shells overlie darker, barren silts and clays (all of which comprise unit D). Note the prominent fold above the hammer (circled) which is overlain by horizontal and undisturbed beds. Deformation is likely due to slumping in the marine environment.

The upper limit of this unit forms a lower, prominent delta surface at 80 m asl which extends upvalley to 90 m asl and is composed of rounded gravels and cobbles, likely concentrated by the deflation of sand. Downvalley, the surface of unit E drops to 49 m asl where it forms an outer, prominent delta surface.

5.5 Site 1: Interpretation

The conspicuously dipping beds of sand, gravel, and cobbles at the base of the section (unit A) are interpreted as foreset beds formed in a prograding Gilbert-type delta (Matthews, 1984; Miall, 1984). The regular bedding exhibited in this unit dips 23^0 downvalley which is characteristic of delta foresets (Miall, 1984). If the unit were a debris flow, the beds would be irregular and would dip towards the centre of the valley, parallel with the valley walls.

The overlying horizontal beds of sand, gravel, and cobbles (unit B) which truncate the dipping beds of unit A are interpreted as possible topset beds of the same delta, deposited in a relative sea level 26.5 - 28 m asl. The outcrop of unit B upstream of site 1 may be either a bedrock-controlled meander of the same topsets or outwash which graded to an earlier, lower relative sea level at 20 - 24 m asl. However, this interpretation is not unequivocal, for horizontally bedded sands, gravels and cobbles may be interpreted as subglacial outwash or ice-proximal subaquatic outwash. It can only be said that the relative sea level indicated by units A and B was *at least* 20 m asl prior to the onset of the last glaciation.

The presence of limestone and dolomite erratics and carbonate content in unit C indicate that the diamicton originates upvalley. Combined with the marked striation of the clasts and the strong downvalley fabric, this demonstrates deposition by a glacier rather than by local debris flows; therefore, the diamicton is interpreted to be a basal till. The occurrence of this till 500 m from the present-day coast suggests that the former valley glacier advanced and was grounded at least to this point. The truncation of the top of unit B at the upstream site indicates that glacial erosion also took place.

The rhythmically bedded silts and clays (unit D) overlying the till are interpreted as deep-water sediments deposited in an ice-proximal environment. These fine-grained, laminated to bedded sediments are characteristic of low-energy, deep-water deposits. The draping of an underlying surface (till) by silts and clays indicates that the sediments originate from suspension settling (Powell, 1981; Stewart, 1988). The interfingering of the till and the rhythmites suggests that the ice retreated in contact with the sea. The sediment source may have been highly sediment-charged subglacial meltwater streams which upon entering deep water produced laminated mud lithofacies, the proximity of deposition depending on turbulence and density differences within the water column (Powell, 1981). Turbid plumes occur when subglacial or englacial meltwater exits a glacier, enters denser sea water and rises to the surface, transporting sediment by overflows or interflows (Fig. 5.24) (Miall, 1984; Lemmen, 1988; Stewart, 1988). Lemmen (1988) and Stewart (1988) identified turbid plume

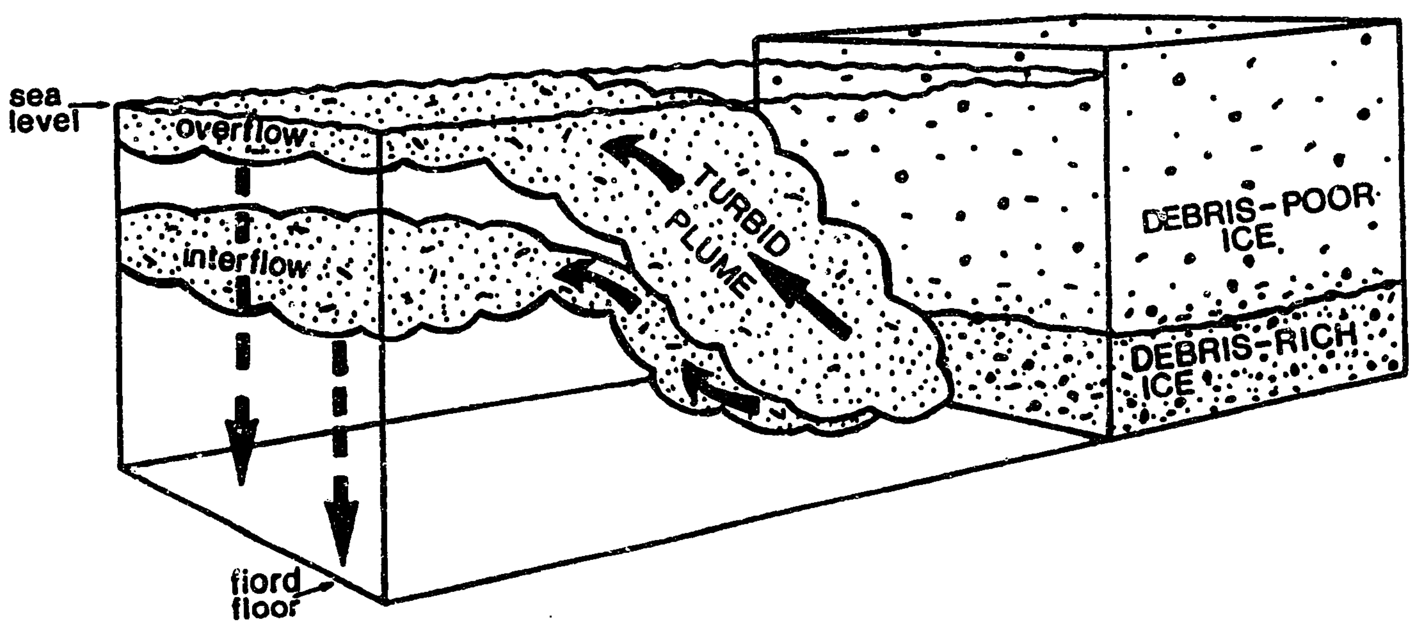

Fig. 5.24. Shematic diagram of turbid plume transport and suspension settling at the margin of a tidewater glacier (from Lemmen, 1988).

deposits on northernmost Ellesmere Island. Turbid plume deposits are of paleoclimatic significance for they indicate that warm-based, tidewater glaciers existed in the high arctic in the past, whereas present-day glaciers are sub-polar.

The transition between the underlying darker, barren silts and clays into the lighter, fossiliferous silts may represent a change from anaerobic to aerobic conditions. On northeastern Ellesmere Island, England (1983) and Retelle (1983) observed that marine silts which predate inital emergence (and therefore deglaciation) contain sparse fauna compared to younger deposits and that beaches at marine limit are also poorly developed compared to those at lower elevations. This was attributed to a shift from landfast permanent sea ice to seasonal sea ice (following deglaciation) which favoured marine faunal colonization and beach development. In the Foster Creek area, it is clear that the dark silts and clays were deposited following initial deglaciation; however, the transition to light, fossiliferous sediments may be related to a shift to a seasonal sea ice cover. The shell fragment from this boundary of dark and light silts and clays provides both an age of this event and a minimum date on deglaciation of 7570 ± 60 BP.

The gently dipping beds of rounded cobbles, gravel, and sand (unit E) overlying unit D (Fig. 5.16) are interpreted as foreset beds of a prograding delta deposited into a relative sea level at least 90 m asl. The delta surfaces (80 - 90 m and 49 m asl) are well below local marine limit (131 m asl), therefore they must postdate it. The absence of topset beds suggests that an erosional surface was formed as relative sea level dropped. Deltaic bottomsets which are often associated with foreset beds were not observed, however, fossiliferous silts and clays occurred at the top of unit D.

The small, raised marine delta and the wave-washed surface on the bedrock knoll at 130 - 131 m asl must postdate the valley glaciation and thereby represent a deglacial marine limit of 131 m asl. This falls within the gradient of marine limit observed in Canon Fiord (Hodgson, 1985; England, 1990). If the glacier retreated in contact with a 131 m sea, the silts and clays of unit D (at 30 - 60 m asl) were deposited in 70 to 100 m of water; this supports their interpretation as deep-water rhythmites.

5.6 Site 2: Description

Site 2 is ca. 2 km upvalley from the fiord coast at the apex of the raised delta (Figs. 5.1, 5.3, 5.25). The five units exposed are described below; their stratigraphic positions are illustrated in Figures 5.15, 5.26 and 5.27.

Unit A' - Unit A' consists of steeply dipping beds (ca. 15 - 20^0) of poorly sorted, matrix-supported gravels and cobbles interbedded with diamicton (Fig. 5.26). This 1.5 m thick unit is overlain unconformably by unit B' at 36 m asl. It is a limited exposure, ca. 2m x 2 m.

Unit B' - Unit B' is interbedded with unit C', and consists of interbedded sands

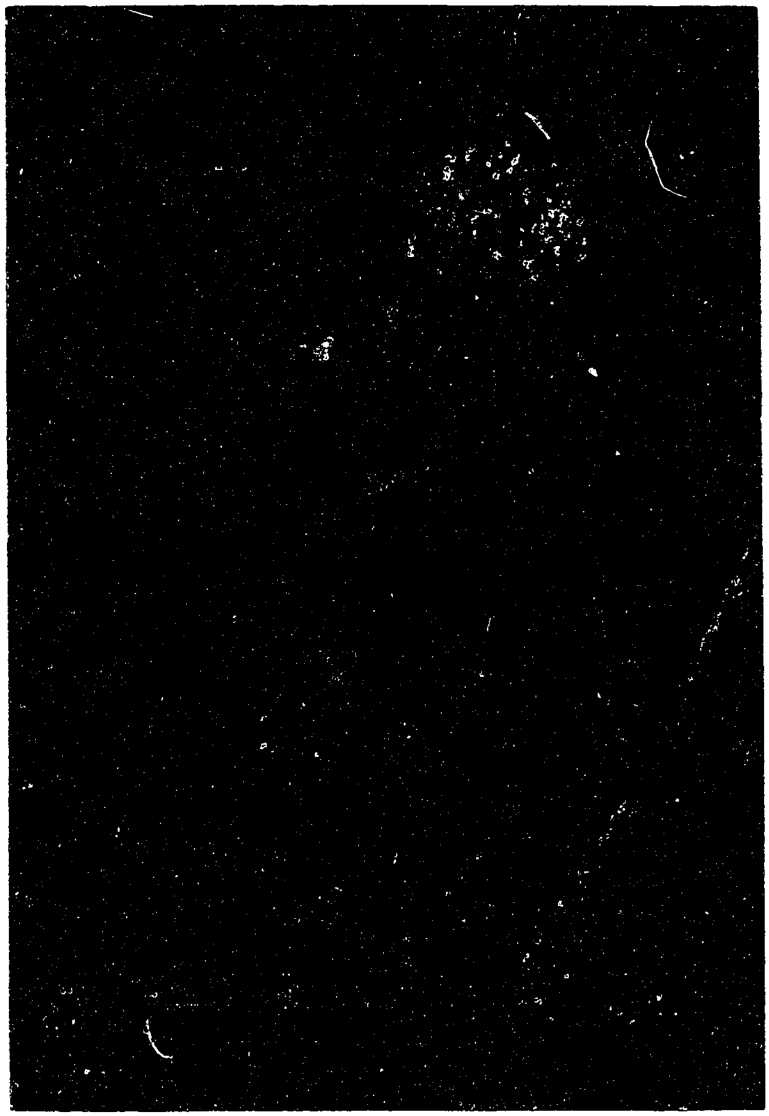

Fig. 5.25. A view of section 2 (arrow) looking upvalley. Note the talus-covered gravel terrace at the top of the section. The clay and silt mounds in the middleground make up unit D.

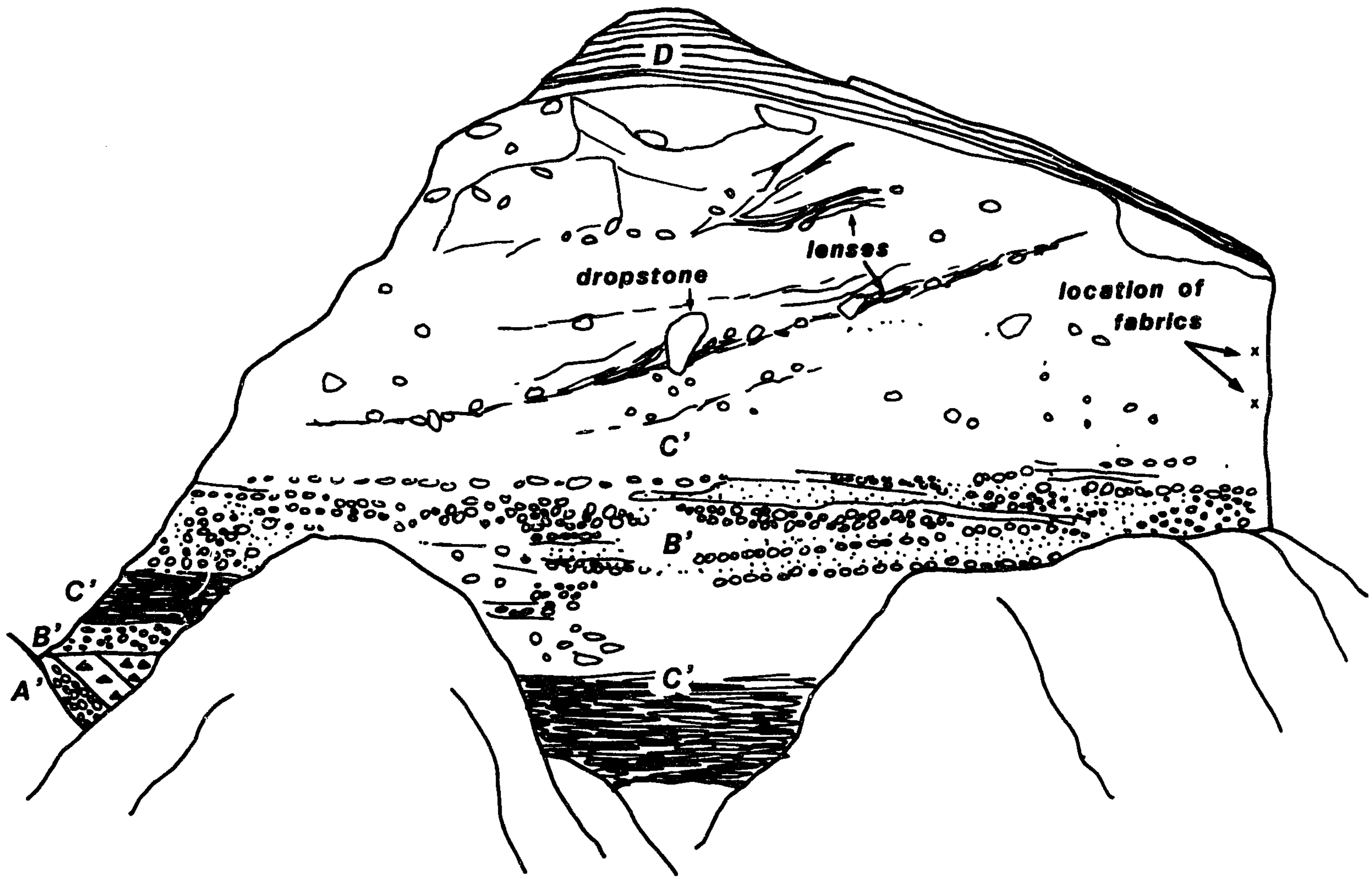

Fig. 5.26. A schematic diagram of the lower part of site 2, showing units A', B', C', and D.

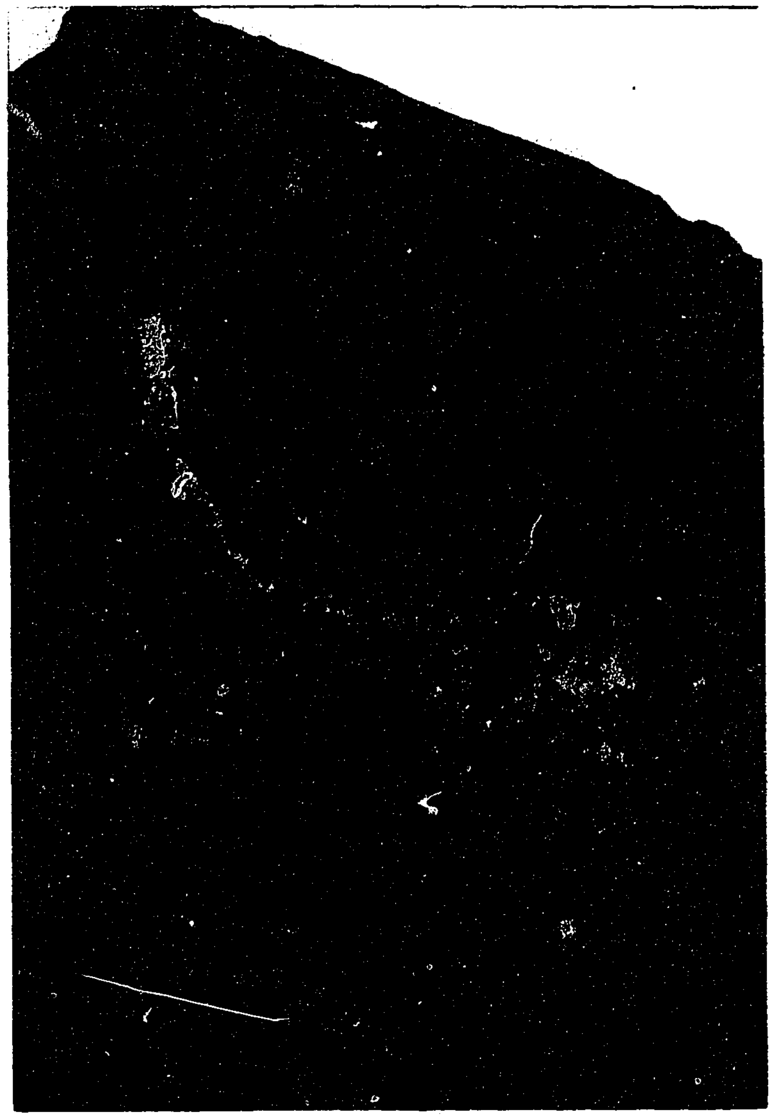

Fig. 5.27. The lower units at site 2. The dark area at the base recently slumped, exposing frozen sediments. Note the 2 m thick unit of gently dipping, interbedded sands and matrix-supported gravels (behind David Bechtal) which grades up into diamicton. About 1.5 m above, sand and silt beds within the diamicton dip upvalley. The ice-rafted boulder within these sands and silts is circled. These units are interpreted as subaquatic, ice-contact deposits, similar to those described by Stewart (1988), which are interbedded with till and ice-rafted material.

and matrix-supported limestone and dolomite gravels and cobbles that dip gently downvalley (ca. 5^0) (Figs. 5.26, 5.27). The sands are massive and thinly bedded, while the matrix-supported gravels and cobbles are poorly sorted. This unit is quite recessive - for example, fresh colluvium is observed immediately downslope of unit B' in Figure 5.27. A subhorizontal bed of diamicton-supported cobbles marks the boundary of units B' and C' at 38 m asl (Figs. 5.26, 5.27).

Unit C' - The lower beds of unit C' resemble the diamicton at site 1. Unit C' is characterized by massive, dark grey silt containing carbonate and striated limestone and dolomite clasts. Two fabrics (of 35 clasts each) measured within unit C' at the western end of the exposure indicate a preferred east-west clast orientation parallel with the valley and an average downvalley dip of 20^0 (Fig. 5.26, 5.28). In the uppermost 6 m of unit C', the diamicton exhibits faint bedding marked by lenses of sand and silt which dip upvalley and exhibit angular unconformity (Figs. 5.26, 5.27). The sand and silt lenses thin upvalley into beds of cobbles (Figs. 5.26, 5.27). One boulder (ca. 40 cm x 60 cm) within the sand and silt lens, oriented with its long axis downwards, both deforms and is overlain by the sands and silts (Figs. 5.26, 5.27). The diamicton unit is conformably overlain by, and grades upwards into, unit D.

Unit D - Unit D consists of laminated to bedded silts and clays which can be traced laterally downvalley to unit D at site 1. The exposure of unit D is ca. 33

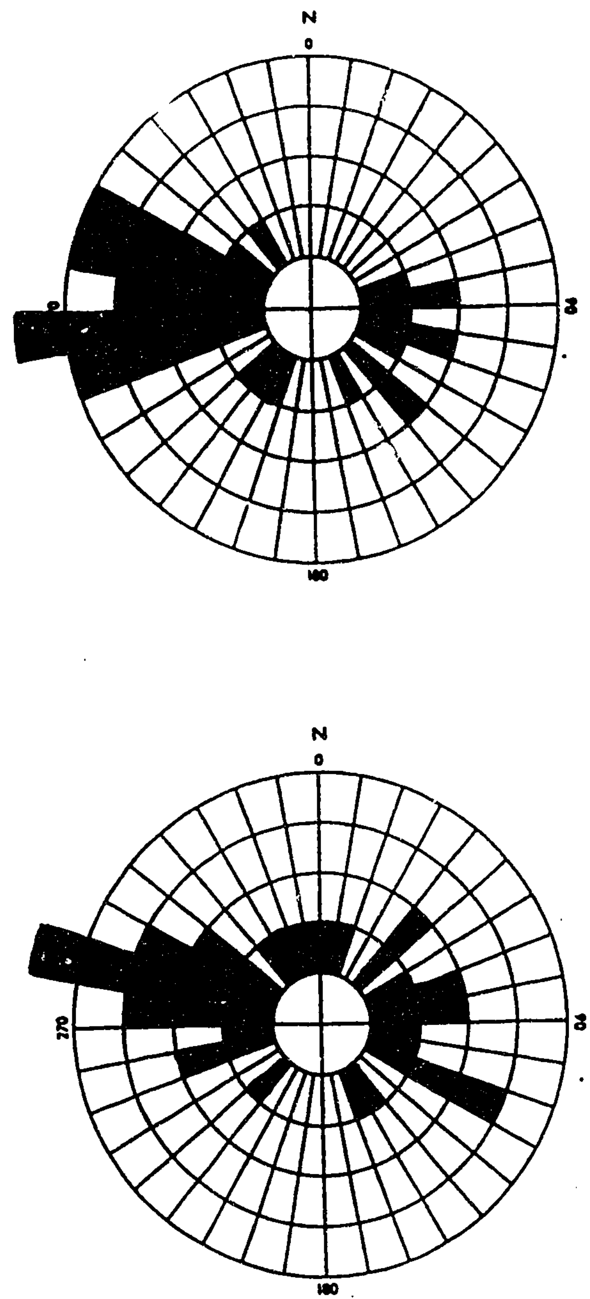

Fig. 5.28. Results of two fabrics conducted in lower beds of unit C' at the western end of site 2. They show a preferred east-west orientation of clasts which had a mean upvalley dip of 20^0.

m thick, although the upper contact is obscured by colluvium from the overlying unit,E, at 78 m asl (Fig. 5.25). Although the silts and clays outcropping at site 2 are exclusively dark grey and barren, the light grey silts and clays outcrop ca. 100 m downvalley at 85 - 94 m asl where they are fossiliferous. Mya truncata found in growth position at this exposure provided a date of 8020 ± 90 BP (GSC-4706).

Unit E - Unit E at site 2 resembles unit E at site 1, consisting of loose, rounded cobbles, gravel and sand. Unlike site 1, no bedding was observed, possibly due to greater accumulations of colluvium. The upper boundary of unit E forms a terrace which is covered with large, angular boulders from an adjacent talus slope and grades from 112 to 120 m asl upvalley (Fig. 5.3, 5.25).

5.7 Site 2: Interpretation

The unit of dipping beds of matrix-supported gravels and cobbles interbedded with diamicton at the base of the section (unit A', Fig. 5.26) is interpreted as a subaquatic debris flow onto a proglacial slope from an englacial stream. The occurrence of diamicton interbeds suggests deposition in an ice-proximal environment, while the matrix-supported gravels and cobbles are characteristic of sub-aquatic outwash (Stewart, 1988). The steep dip of the beds also indicates a debris flow. Although this unit resembles dipping deltaic foreset beds, such as unit A at site 1, the sedimentology is not characteristic of foresets, because the unit contains diamicton interbeds. Interpretations of this unit are tentative because the exposure of unit A' is minimal.

The matrix-supported gravels of unit B' are interpreted as subaquatic outwash derived from the base of the former valley glacier, similar to those described by Stewart (1988). Matrix-supported gravels indicate a density sufficient to prevent coarse particles from sinking or sufficient clast-to-clast collision to keep the particles in suspension, representing a highly concentrated slurry or debris flow (Stewart, 1988). The low angle of bedding in unit B' suggests that it represents a slurry where outwash entered the marine environment. The bed of diamicton-supported clasts at the boundary of units B' and C' indicates a gradational change of processes. Like unit D, site 1, this deposit is significant because it suggests deposition by a warm-based tidewater glacier.

The unit of interbedded diamicton, sand, and silt (unit C'), is interpreted as consisting of glacial and ice-proximal glaciomarine deposits. The diamicton closely resembles unit C at site 1 (till), and fabrics within the lower beds of unit C' indicate a downvalley clast orientation and upvalley dip suggesting glacial deposition. However, the primary bedding evident in unit C' (Figs. 5.26, 5.27) indicates that glacial deposition was not the only process at work. Sand and silt lenses within unit C', containing dropstones which both deform and are draped by the sand and silt, may represent ice-rafting at the front of a glacier in a subaquatic environment. The interbedding of units C' and B' (subaquatic outwash) is interpreted as deposition at a fluctuating ice margin within a marine environment.

The diamicton at site 2 is overlain conformably by laminated to thinly bedded silts and clays (unit D) which extend laterally downvalley to unit D at site 1. Unit D is interpreted as deposition from ice-proximal, subaquatic, turbid plumes which are overlain by rounded cobbles, gravel, and sand (unit E) interpreted as foresets of a deglacial delta, as at site 1. The upper elevation of unit E indicates that this deglacial delta was deposited in a relative sea level at least 120 m asl, at least 8,020 $\pm$ 90 BP, the ^{14}C age of shells within unit D.

5.8 Summary

The overall sequence of events reconstructed from deposits at sites 1 and 2 is as follows. A delta prograded into a relative sea level at least 20 m asl in front of an advancing valley glacier in Foster Creek. The valley glacier then overrode this delta, depositing till, en route to the fiord where it likely calved in the glacioisostatically raised sea. The ice then retreated upvalley in contact with the sea which transgressed the land to 131 m asl. Till, subaquatic outwash, and ice-rafted material were deposited into the sea at the margin of a fluctuating ice front. Where subglacial streams entered the sea water, turbid plumes formed, providing sediment which settled out of suspension to deposit deep-water rhythmites, draping the till. Marine shells found in growth position within these deep-water sediments provide a minimum deglacial date of 8020 $\pm$ 90 BP. Once the valley glacier retreated above relative sea level, glaciofluvial sediments were deposited in a Gilbert-type delta which prograded over the deep-water rhythmites towards the fiord. As relative sea level dropped, the deltaic sediments were eroded and redeopsited, forming surfaces at 90 - 80 m and eventually 49 m asl (Fig. 5.29).

Fig. 5.29. Remnants of glaciomarine deltas along lower Foster Creek. The delta surfaces descend from 120 m asl behind the viewer to 49 m asl at the surface beyond the tents in the left centre of photo. View is to the NW across Canon Fiord to Fosheim Peninsula.

CHAPTER 6

Conclusions

Glacial geology of central Canon Fiord suggests that the last glaciation was marked by expanded plateau ice caps which were coeval with a relative sea level at 119 to 131 m (Fig. 6.1). During this interval, the paleoglaciation level was approximately 420 m asl. In Foster Creek, the Agassiz Ice Cap advanced 8 km downvalley and contacted the sea, resembling many contemporary calving glaciers on northern Ellesmere Island (Fig. 6.2). When the ice reached the lower valley, the sea was already at least 20 m asl. Subsequently, the glacier retreated upvalley in contact with a 131 m sea level, depositing deep water rhythmites interpreted as turbid plume deposits. These sediments may represent deposition by a warm-based tidewater glacier which indicates a different glacial regime than that of the subpolar glaciers today (Lemmen, 1988; Stewart, 1988). Marine fauna within the deep-water rhythmites provide a minimum age for the deglaciation of lower Foster Creek (8020 ± 90 BP, GSC-4706).

In South Bay, plateau glaciers were contemporaneous with a 119 m sea level, dated 7930 ± 20 BP (GSC-4697) by *in situ* shells. In a more extensive, earlier glaciation, a trunk glacier occupied the valley east of Wolf Ridge and flowed into a sea between 147 to ≤ 164 m asl, forming an ice shelf in South Bay (Fig. 6.3). Shell fragments retrieved from a fossiliferous diamicton likely deposited by this ice shelf provided three AMS dates all older than 29,000 BP. These are interpreted as minimum age estimates. This former sea level at 147 - 164 m asl

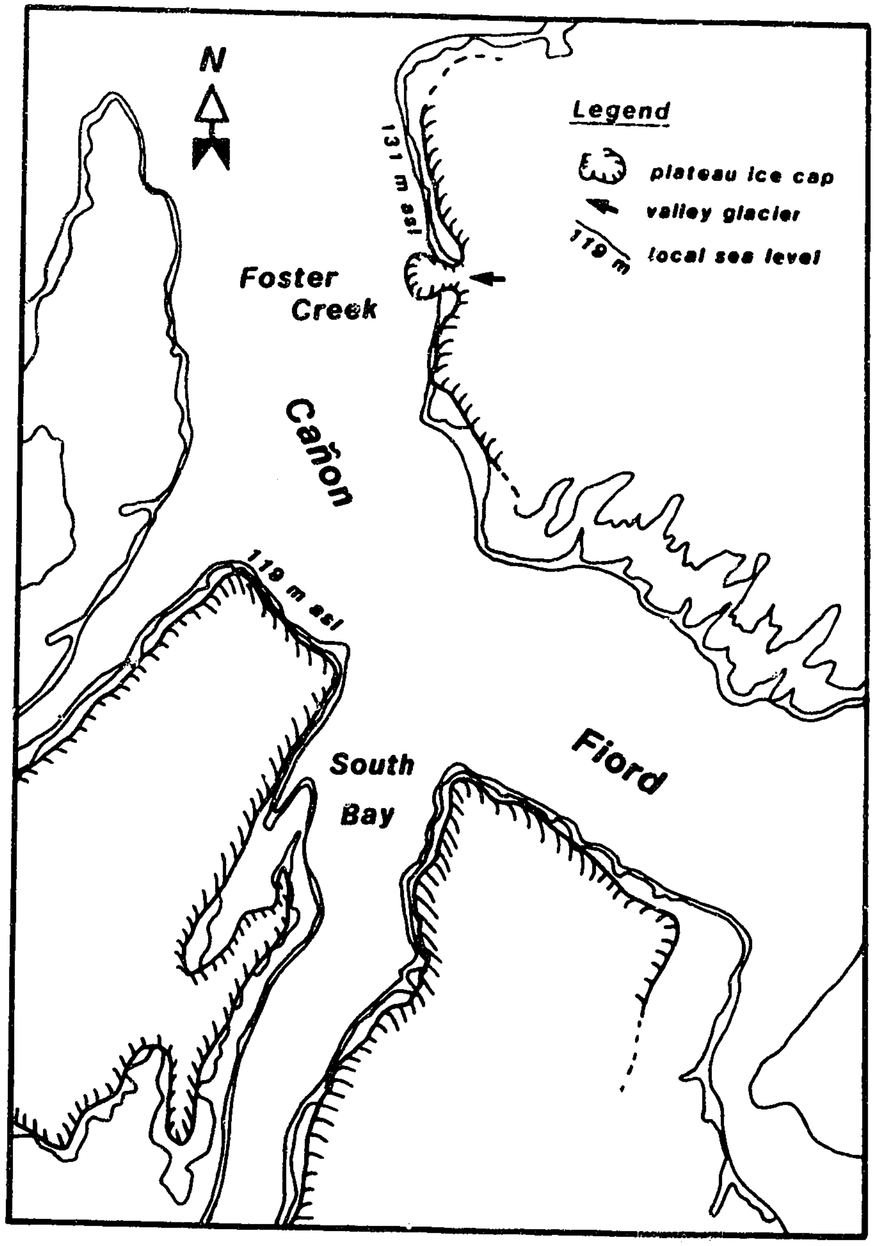

Fig. 6.1. Schematic map showing extent of the last glaciation in central Cañon Fiord with ice caps covering the plateaus and ridges adjacent to South Bay and Foster Creek.

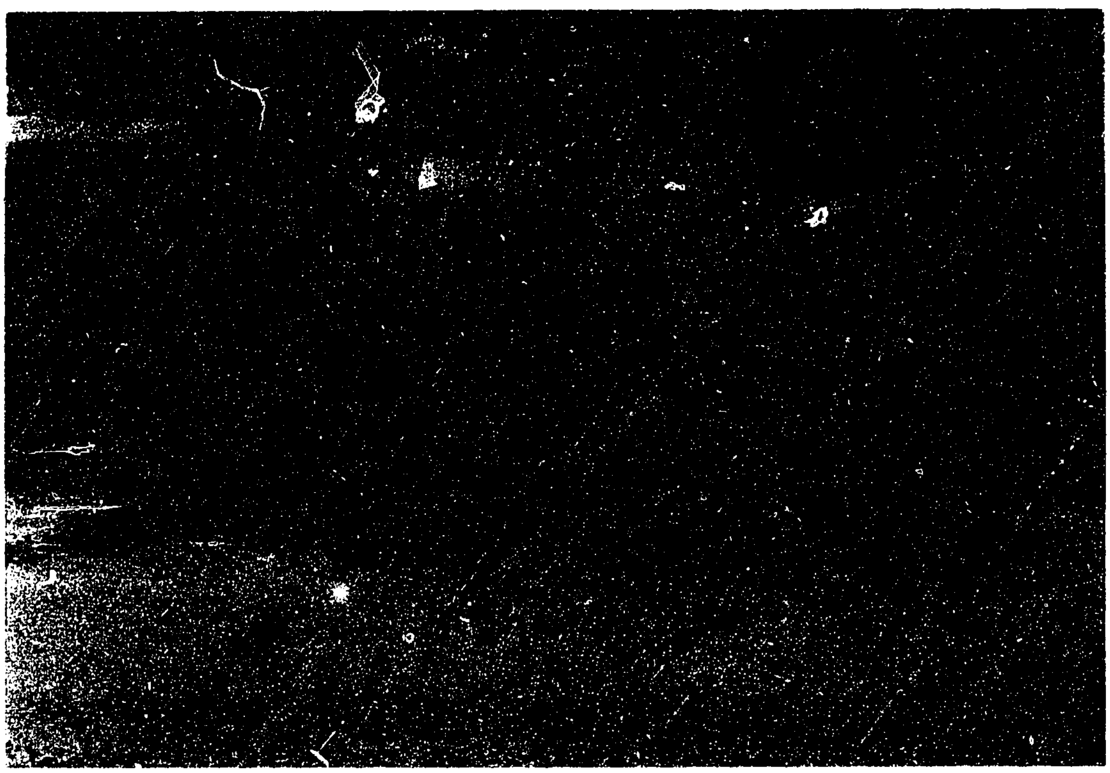

Fig. 6.2. Outlet glacier flowing into M'Clintock Inlet on northernmost Ellesmere Island.

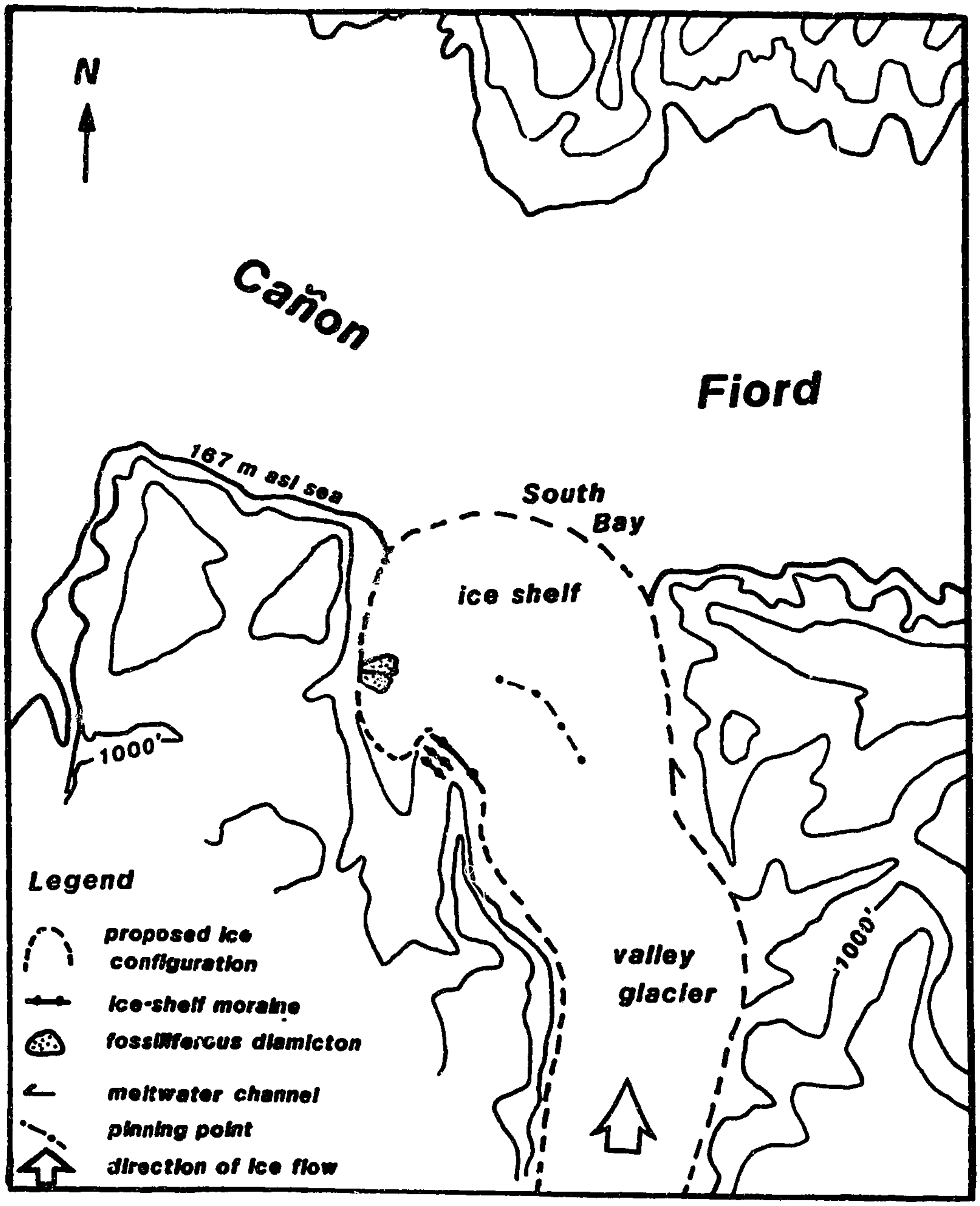

Fig. 6.3. Schematic map showing the extent of an earlier glaciation in central Cañon Fiord with a trunk glacier occupying the tributary valley to South Bay and an ice shelf in a 167 m asl sea in South Bay.

is $\geq$ 28 m above the Holocene marine limit, and it is considered to be a discrete interval of glacioisostatic loading associated with an older glaciation.

This reconstruction suggests that during the last glaciation only local plateau caps occurred contemporaneous with a 119 to 131 m sea level and that regional ice was restricted to an earlier glaciation. This reconstruction resembles the models presented by Hodgson (model C, 1985) and England (1990) which suggest that the Agassiz and central Ellesmere Island Ice Caps expanded 5 - 60 km beyond their present margins, supplemented by local ice caps on low-lying plateaus ($\geq$ 420 m asl).

The local Holocene marine limits in South Bay and Foster Creek, 119 m and 131 m asl, respectively, are consistent with the gradient of the marine limit in Canon Fiord which slopes from 148 m asl at the fiord mouth to 110 m asl near its head (England, 1990; Hodgson, 1985, respectively). In outer Canon Fiord the marine limit is considered to be of full glacial age and to have remained stable between 8400 and 7400 BP, followed by postglacial emergence. In inner Canon Fiord, the most distal, recognizable ice margin is flanked by kame deltas at 110 m which are considered contemporaneous with marine shells dated 7950 $\pm$ 130 BP. The marine limit recorded in South Bay at 119 m asl, dated at least 7930 $\pm$ 70 BP, is also considered to be distal to the last ice limit, marking the full glacial sea.

In Foster Creek, the 131 m marine limit is recorded behind the last ice limit and is therefore deglacial; however, the ice had retreated only 1 - 2 km when the small raised marine delta was deposited at least 8020 ± 90 years BP. The stratigraphic position of the marine fauna below a postglacial delta indicates that relative sea level was at least 120 m asl at the time of their deposition, and possibly 131 m asl. The retreat of ice 1 - 2 km upvalley in central Canon Fiord at ca. 8 ka is not inconsistent with the record in outer Canon and Greely Fiords where the former ice margin had retreated only a few kilometers by 7850 BP. This suggests that the glacioisostatic loading was essentially maintained until this time (England, 1990).

This pattern of emergence within Canon FIord with an upward tilt downfiord does not correspond with the distribution of ice during the last glacial maximum. The greater emergence observed in the outer fiord would suggest that the ice load was more substantial than within the inner fiord; however, the ice advanced from central Ellesmere Island and only small plateau glaciers occupied inner Fosheim Peninsula. Alternatively, this pattern of tilt downfiord may be due to the retreat of a trunk glacier occupying central Canon Fiord, with relative sea level becoming progressively lower as the ice retreated upfiord. Marine shells of Holocene age within outer Canon Fiord are as old as 8800 BP, and these pertain to the full glacial sea distal to the last ice limit (England, 1990). This contrasts with the oldest shells in the central and inner fiord at 8000 BP. Recent fieldwork within central and inner Canon Fiord suggests that a trunk glacier occupied this area during the last glaciation, possibly terminating just to the east of South Bay (England, pers. comm.). Nevertheless, within the inner half of the fiord, the marine limit, which tilts from 141 m at the Sawtooth Range (England, 1990) to 110 m near the head of the fiord (Hodgson, 1985), is essentially synchronous (8000 BP). This decline in marine limit (toward the fiord head) is difficult to explain given the configuration of the last ice limit in central Canon Fiord where ice originted from east-central Ellesmere Island. The relationship of the former glacial configuration and the pattern of postglacial emergence warrants further study; especially emphasizing what are full glacial versus deglacial marine limits. This problem is well exemplified in this study area because the 8000 BP marine limit in South Bay is distal to the last ice limit, whereas it is deglacial in Foster Creek.

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APPENDIX

Table 1

Stratification Type	Thickness
Very thickly bedded	> 1 m
Thickly bedded	30 - 100 cm
Medium Bedded	10 - 30 cm
Thinly bedded	3 - 10 cm
Very thinly bedded	1 - 3 cm
Thickly laminated	3 - 10 mm
Thinly laminated	1 - 3 mm
Very thinly laminated	< 1 mm

(After Ingram, 1954, in Stewart, 1988).

Table 2.

^{14}C Dates From Central Cañon Fiord, Ellesmere Island

Site Location	Lab. Number[a]	Material	Age (years BP)	Stratigraphy	Sample elev.	Related Sea Level	Lat. (N)	Long. (W)
1 South Bay	GSC-4697	Shells	7,930±70 BP	Marine silts	90	≥90-≤119	79°41'	81°44'
2 South Bay	TO-1200	Shells	38,100±380 BP	Surface sample	167	≥167	79°33'	81°48'
3 South Bay	TO-1201	Shells	34,950±340 BP	Surface sample	167	≥167	79°33'	81°48'
4 South Bay	TO-1198	Shells	29,380±230 BP	Surface sample	132	≥132	79°32'	81°40'
5 Foster Creek	TO-1199	Shells	7,570±60 BP	Marine silts	23	≥23-≤131	79°49'	81°33'
6 Foster Creek	GSC-4706	Shells	8,020±90 BP	Marine silts	94	≥94-≤131	79°49'	81°30'

[a]The laboratory prefix TO designates samples dated by AMS at ISOTRACE, University of Toronto. The prefix GSC designates conventional ^{14}C dates from the Geological Survey of Canada, Ottawa.