QUATERNARY GEOMORPHOLOGY, GLACIAL HISTORY AND RELATIVE SEA LEVEL CHANGE IN OUTER NACHVAK FIORD, NORTHERN LABRADOR

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QUATERNARY GEOMORPHOLOGY, GLACIAL HISTORY AND RELATIVE SEA LEVEL CHANGE IN OUTER NACHVAK FIORD, NORTHERN LABRADOR

BY

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

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"The grandest scenery of Labrador begins with the ragged shapes of the Kiglapait near Port Manvers, continues nobly with the groups at Cape Mugford, and culminates in a magnificent climax at Nachvak.

Nowhere can the student of geology and of the causes of scenery better study his problems than here where nature lays bare the manner of her working, without concealing its stages and effects under the hindering veil of snow or of vegetation. Clear and sharp stand out the evidences and nature of glacial action ages ago, and of the alternate sinking and rising of the land with reference to the level of the sea. Beautifully plain appear the effects of denudation by frost and water, as seen in the shattered summits, the forms of the mountain masses, the accumulations of talus, and the great curves and complexity of the valleys. Standing on one of these heights, one can almost trace out the whole history of the diversification of the original simple ridges into the intricate system of varied peaks and valleys that now exist.

Evidences of the former glaciations of this country are abundant."

E.B. Delabarre, 1902

Abstract

The Torngat Mountains of northern Labrador portray a complex history of regional and local ice movements during the Wisconsinan. Research undertaken in outer Nachvak Fiord, central Torngat Mountains, provides the basis for a morphostratigraphic and lithostratigraphic framework. The application of relative and absolute dating techniques allows reconstruction of a tentative glacial and relative sea level chronology for the area.

The *Nachvak* glacial phase represents the last regional ice advance into the outer Nachvak Fiord area. During this event, tributary valleys were backfilled by fiord ice as far east as the fiord threshold, whereas south of the fiord, ice terminated at the western end of Adams Lake. This ice advance is distinguished from previous glacial events on the basis of its drift properties. The earlier *Adams Lake* glacial phase is considered to be local in origin. The eastern limit of this advance, as defined by the maximum extent of local lithological and geochemical drift characteristics, was Valley of the Flies terminal moraine. The *MI* glacial phase represents an extensive, older regional ice advance and is characterized by the upper limit of till and moraine in the study area.

Acoustic stratigraphy of Nachvak Fiord and Adams Lake sediments supplements the terrestrial glacial geology of the region. Acoustic units are interpreted as representing sedimentary facies related to one or more glacial - deglacial cycles, in both the fiord and lake basins. Tentative models of deposition require a grounded to partially-grounded outlet glacier in the fiord during the last regional ice advance, and a significant ice margin at the western end of Adams Lake.

Analysis of raised marine evidence reveals eleven shorelines, ranging in elevation from 9 m to 73 m above present sea level. The extent and geometry of the shorelines are the bases of a model of relative sea level change during deglaciation from the Nachvak phase maximum. This permits the identification of the Shoal Water Cove and Tessersoak glacial readvances, and the

Kogarsok and Townley Head glacial still-stands which appear consistent with the morphological evidence of moraines in the region. Locations of major loading centers which affected the Nachvak Fiord region, are based on the assumption of synchronous shoreline formation in the outer fiord area. Radiocarbon-dated shells from sediments associated with a raised marine shoreline suggest that approximately 30 m of emergence has occurred since 9 ka BP.

A tentative chronology of glacial events is reconstructed from radiocarbon dates, amino acid ratios, and soil development rates. The Nachvak glacial maximum is tentatively dated at 22 $\pm$2 ka BP. It is estimated that the Adams Lake glacial phase occurred during the period 29-50 ka BP. The MI glacial phase is tentatively dated at $\geq$ 75 ka BP, based on comparisons with similar regional ice advances on Baffin Island.

Correlations with local chronologies from northern Labrador are proposed, and critically examined in the context of previous models of glaciation. The Early Wisconsinan in northern Labrador was likely characterized by an extensive advance of Laurentide ice which may have extended onto the Labrador Shelf. The Middle Wisconsinan is suggested to have been a period of local cirque glacier expansion, with associated high relative sea levels. The extent of Late Wisconsinan ice may have been topographically controlled in northern Labrador. In the Nachvak Fiord region ice was restricted to the fiord basin, whereas in areas north and south of the Torngat Mountains the ice advance may have been more extensive than during the Early Wisconsinan. The deglacial chronology proposed for Nachvak Fiord suggests that a stepwise recession of Laurentide ice characterized the late-glacial period.

Acknowledgements

This thesis is the product of ideas which have evolved and matured with the help of many people. Dr. Robert J. Rogerson provided advice, encouragement, and friendship during my three years at Memorial University. Whether through subconscious effort, or constant irritation, he has allowed me the space to learn and create in my own manner, yet he has drawn and guided my thoughts with imperceptible supervision. I hope this thesis fulfills his expectations.

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During my two fields seasons I had the assistance of Laurence Davis (1984), Lloyd St. Croix (1984), Craig Cummings (1985) and Robert Rowsell (1985). Their enthusiasm and dedication are responsible for the extensive data sets employed in this thesis. Their resilience in the face of danger and diminishing chocolate supplies is to be commended. Their humour and patience made awkward situations memorable. I would also like to thank Caren Garrity, Alison Dyer and others who contributed to the atmosphere of base camp during their stay with us.

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The flying skills of Peter Horsman (Golfe Helicopters) deserve special attention, as they were responsible for memorable moments high above the Torngat Mountains. I would also like to take this opportunity to thank Mrs. Minda Rogerson for welcoming me into her home and for constantly testing my ability to eat her pies. Such good memories kept me alive on cold, wet, hungry days in northern Labrador.

Last of all, I would like to thank my family who must, from time to time, question my sanity but nevertheless have supported me to the end. This is as much theirs as mine.

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Glossary of placenames

The following is a list of official placenames from the Nachvak Fiord region which appear in the text. The Inuttut names were originally translated by Rev. A. Stecker of the Moravian Society (Daly, 1902). Other landforms were named after members of a Harvard scientific expedition who visited Nachvak Fiord during the summer of 1900.

Adams Lake	Mr. Huntingdon Adams, expedition organizer.
Bigelow Bay	Mr. H.B. Bigelow, Harvard undergraduate.
Delabarre Bay	Prof. E.B. Delabarre, Brown University.
Ivitak Cove	'red-coloured', land with a reddish colour; red ochre; bricks.
Kammarsuit Mtn.	'wall', referring to the shape of the mountain.
Kogarsok	'small brook'.
Koktortoaluk	'big waterfall'.
McCornick River	Mr. L.B. McCornick, Harvard undergraduate.
Nachvak Fiord	'found, found at last'. The first Eskimo, coming from the west in search of the salt water, cried out "Nachvak" when he reached the head of the fiord (Daly, 1902).
Naksaluk Cove	'great valley'.
Palmer River	Mr. H.W. Palmer, Harvard undergraduate.
Tallek Arm	'arm', referring to the shape of the bay.
Tessersoak	'large pond', referring to Nachvak Lake.
Tasiuyak Arm	'like a pond'.
Tinutyarvik Cove	a basin at the head of a cove, barred off by a continuous wall-like shoal, so that at low tide fish are trapped as in a weir.
Torngat Mtns	home of the 'bad spirits'.

The following unofficial placenames are used in the text for ease of reference.

Kogarsok Brook Cove	Gloom Cove	Bigelow valley
Green Point	Gurnot Lake	Valley of the Flies
West Cove	Blue Pond	Kammarsuit valley west
Bay Cove	Naksaluk valley	Kammarsuit valley east
Shrule Cove	Tinutyarvik valley	Mount Memorial
Ivitin Cove	Adams Lake valley	Gulch peninsula

Chapter 1
Introduction

The late Quaternary history of northern Labrador is largely unknown. Over the last century, elaborate models of glaciation have been proposed but none has gained universal acceptance as the most likely reconstruction which explains the reported field observations. Although such debate is encouraged by the nature of scientific research, the presentation of intellectually-attractive theories based on limited and equivocal evidence distorts the preferred evolution of objective scientific reasoning. These considerations prompted Johnson (1985) to remark that "more detailed local studies .... are needed to insure that ideas put forward about northern Labrador's glacial history by early workers such as Coleman, Daly, Tanner and Ives are put into perspective".

The Torngat Mountains have been the primary source of our present knowledge on Pleistocene glaciation in northern Labrador. The importance of this mountain range in the reconstruction of former ice-sheet configurations has been recognised since the earliest expeditions along the Labrador coast, but not until recently has detailed field research uncovered the wealth of information relating to local and regional glacial events.

Seismic reflection profiling of Labrador Shelf sediments has provided an opportunity to study land-sea interactions during expansion and contraction of the Laurentide ice sheet. A major controversy concerning the maximum extent of ice during the last glaciation should be resolved from evidence relating to seaward ice margins recorded in Labrador shelf sediments, and vertical limits to ice passing through the mountains.

The timing of glacial episodes in northern Labrador has proved problematic due to the paucity of radiocarbon-dated material and the inherent restrictions associated with the development of absolute chronologies from relative-age frameworks. Therefore morphochronological inferences on a local scale are critical to the construction of a regional chronology which may be further developed through applications of robust relative dating techniques.

This study forms an integral part of an on-going project to resolve the glacial history of the Nachvak Fiord area in the central Torngat Mountains (Fig. 1-1). This area is particularly relevant to regional glaciation because the fiord trough provided an important outlet for Laurentide ice flowing from the Labrador-Ungava dispersal centre to the Labrador Sea. In addition, the Selamiut Range, south of the fiord, contains the largest of present day Labrador glaciers and therefore may have hosted large local ice masses during continental glaciation (Fig. 1-1).

1.1. Objectives

The outer Nachvak Fiord area (Fig. 1-1) has been the source of infrequent, isolated observations on glacial history and therefore the main objective of this study is to record and interpret glacial landforms and surficial sediments at a 1:25,000 scale. On the basis of landform associations and morphostratigraphy, a tentative morphochronology of glacial events is proposed. Analysis of glacial drift characteristics complemented by a morphostratigraphic framework is used to determine local and regional ice movements. The application of relative and absolute dating techniques provides a framework for a local chronology which is critically examined in the context of chronologies proposed for the adjacent Selamiut Range and the Iron Strand/Kangalaksiorvik Fiord region, 50 km north of the study area.

Measurement and modelling of raised marine shorelines document relative sea level change in the area which, in association with landform and sedimentary evidence, provides information on the sequence and style of deglaciation along the fiord. Extrapolation and integration of a local glacial and deglacial chronology from the outer Nachvak Fiord area constitutes a test for recent models of glaciation proposed for northern Labrador.

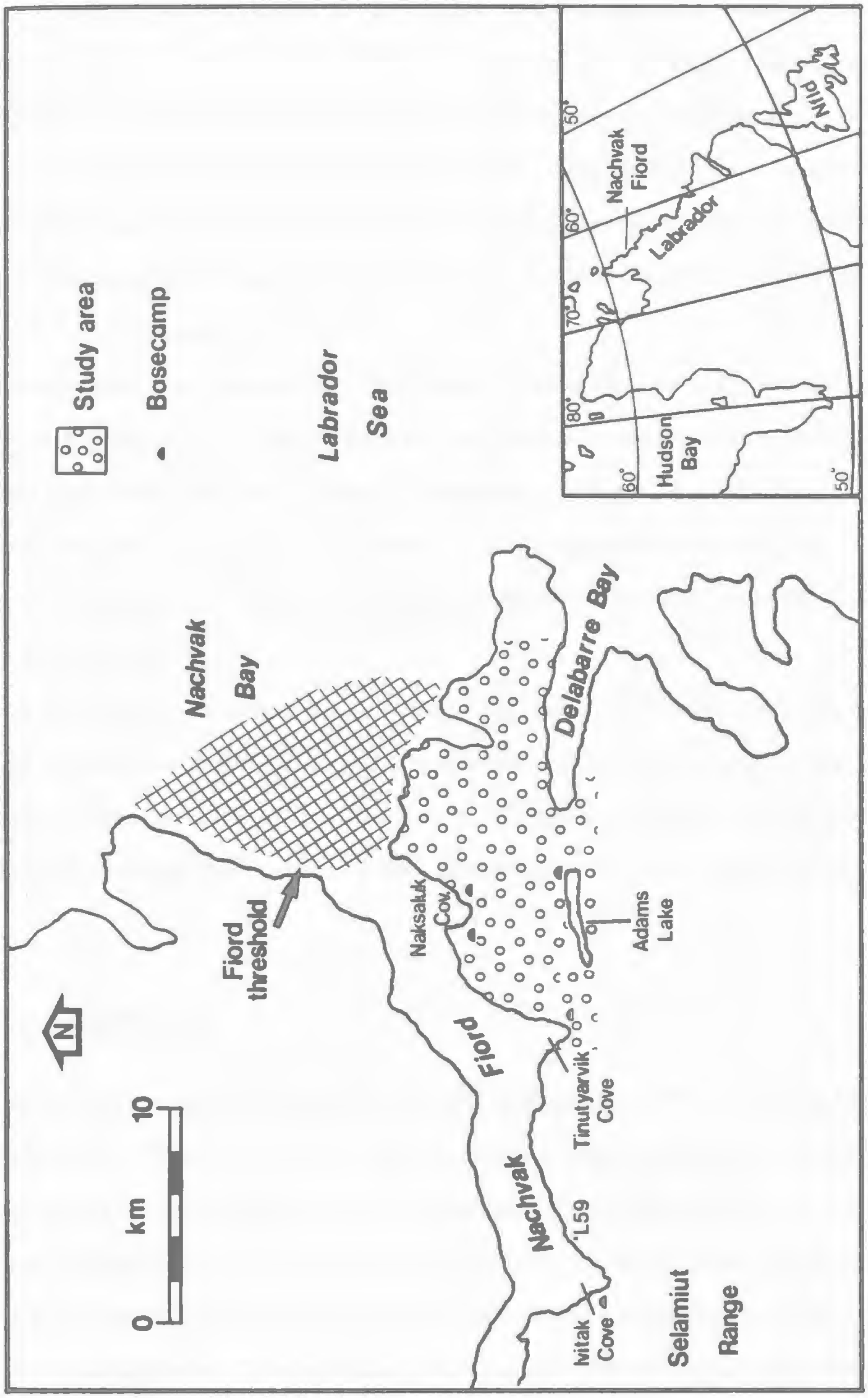

Figure 1-1: Location of the study area in the Nachvak Fiord region.

1.2. Thesis outline

Chapter 2 is designed to familarize the reader with the physical landscape of the study area and to introduce previous research which has been carried out in the general region. Although the latter is primarily a literature review, it is presented to illustrate conflicting viewpoints and apparent inconsistencies which have evolved over time. This review is not complete but serves to introduce specific topics which are more fully appraised in relevant chapters. The last section of Chapter 2 describes the significance of the study area in the context of regional stratigraphic units proposed for northern Labrador.

The multifaceted approach used in this thesis requires that each type of evidence be reviewed independently prior to discussion and formulation of general theories. Therefore glacial landforms and sediments, fiord and lacustrine sediments, and raised marine evidence are presented individually in Chapters 3, 4 and 5, respectively. Data acquisition and analysis are described for each type of evidence and tentative conclusions are discussed in the context of local chronologic and stratigraphic frameworks.

Chapter 6 employs a multiparameter dating approach to develop a relative and absolute chronology of reconstructed glacial events. Radiocarbon dates, amino acid ratios, and indexes of soil development form the basis of this chronology. Chapter 7 summarizes the glacial and relative sea level histories of outer Nachvak Fiord and discusses their regional significance in northern Labrador.

1.3. Field logistics

Field work was carried out between July 20th and August 23rd in 1984 and between July 10th and September 11th in 1985. Two Otter floatplanes were chartered in Goose Bay in 1984 to transport a party of five, including equipment and provisions, 850 km north to Nachvak Fiord. A base camp was established at Ivitak Cove, 12 km west of the study area (Fig. 1-1). This site was chosen for the benefit of the survey team who were working on the cirque glaciers in the Selamiut Range, south of Ivitak Cove. Two secondary base camps were set up at Naksaluk Cove and Tinutyarvik Cove in outer Nachvak Fiord to facilitate this research. A transportation link was maintained with the primary base camp using a rubber Zodiac, although this was dependent upon calm conditions in the fiord, and the movement of sea ice in the outer part of the fiord. A return trip was made every ten days to replenish supplies. Most of the field work was conducted out of the secondary base camps, although it was necessary to use fly-camps on the longer traverses.

Between July 27th and August 2nd 1984, the Canadian Hydrographic Service and the Atlantic Geoscience Centre carried out bathymetric, acoustic and sonar surveys of Nachvak Fiord, from on-board the *CCGS Labrador* which was anchored off Naksaluk Cove. Helicopter flights provided by the survey teams and the Coast Guard allowed extensive reconnaissance of the study area during initial field work.

In 1985, a party of three was transported from St. John's to Nachvak Fiord on the *CCGS Sir John Franklin*. Logistical support using the ship's helicopter helped to establish a base camp on the northeastern shore of Adams Lake, and to set up fuel caches in Naksaluk Cove and Delabarre Bay. A rubber Zodiac was used on Adams Lake to provide easy access to the western part of the study area. Fly-camps were used along the outer coast during extensive levelling and sampling programs.

Towards the end of August, a 3.5 kHz acoustic survey and coring program was carried out on selected lakes in the Torngat Mountains. This project was made possible through logistical support provided by R.A. Klassen, Geological Survey of Canada. A helicopter transported personnel, Zodiac and technical equipment between lake sites and established a new base camp at Naksaluk Cove. At the end of the field season a floatplane brought samples, equipment and personnel south to Goose Bay, with the exception of the author who was collected at Naksaluk Cove by the *CSS Hudson* during a coring program of Nachvak Fiord.

Chapter 2
The study area

2.1. Orientation

The study area is situated along the outer coast of Nachvak Fiord in the Torngat Mountains of northern Labrador. It is bounded by Nachvak Fiord and Nachvak Bay to the north, Tinutyarvik valley to the west, Adams Lake valley and Delabarre Bay to the south, and Valley of the Flies to the east (Figs. 2-1 & 2-2). The divide between Ungava Bay and Labrador Sea drainage, which also forms the boundary between the province of Quebec and the province of Newfoundland and Labrador, approaches within 20 km to the southwest where it passes across the highest summit in both provinces, Mount Caubvick (1738 m). A major part of the 150 km^2 study area is more than 300 m above sea level with Kammarsuit Mountain (905 m) attaining the highest elevation. The area is dissected by a series of valleys which are generally low-lying and bounded by steep slopes. The Kammarsuit Mountain range between Naksaluk valley and Valley of the Flies has an upland valley system radiating out from the centre of the range.

Tinutyarvik valley on the western margin of the study area is approximately 4 km long and is drained by a river which flows north to Nachvak Fiord. The valley extends farther south along the western flank of Quartzite Mountain to a col which descends south into Ramah Bay. Adams Lake valley is approximately 8 km in length, and is oriented in an east-west direction from Tinutyarvik valley to Delabarre Bay. The valley is only 0.5 km wide at its western end but east of Adams Lake widens to 2 km. The drainage divide between Tinutyarvik valley and Adams Lake valley is 53 m above high tide (aht) and therefore the discharge from Adams Lake (36 m aht) flows east to Delabarre Bay. Naksaluk valley runs north-south from the fiord to Adams Lake valley where it appears as a hanging valley 44 m above lake level. Gurnot Lake in Naksaluk valley is 80 m aht and drains north to Nachvak Fiord and south to Adams Lake valley. Valley of

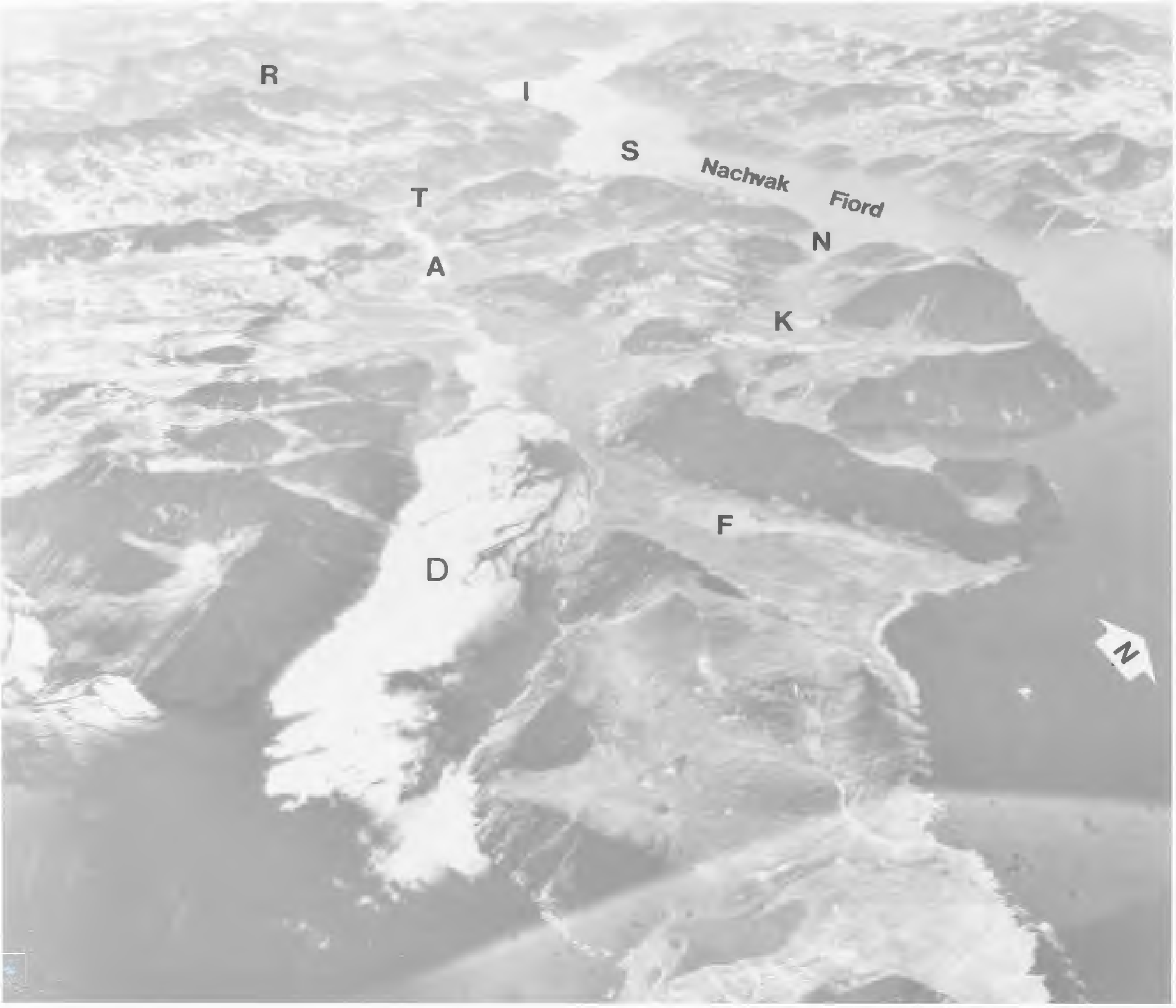

Figure 2-1: Oblique air photograph of the study area and Nachvak Fiord. **A** Adams Lake; **D** Delabarre Bay; **F** Valley of the Flies; **I** Ivitak Cove; **K** Kammarsuit valley system; **N** Naksaluk Cove; **R** Selamiut Range; **S** Approximate location of Shoal Water Cove moraine; **T** Tinutyarvik valley

the Flies is the broadest valley in the study area (5 km^2) and has a floor elevation of approximately 20 m aht. It is a flat, low-lying area with little surface water (Fig. 2-3).

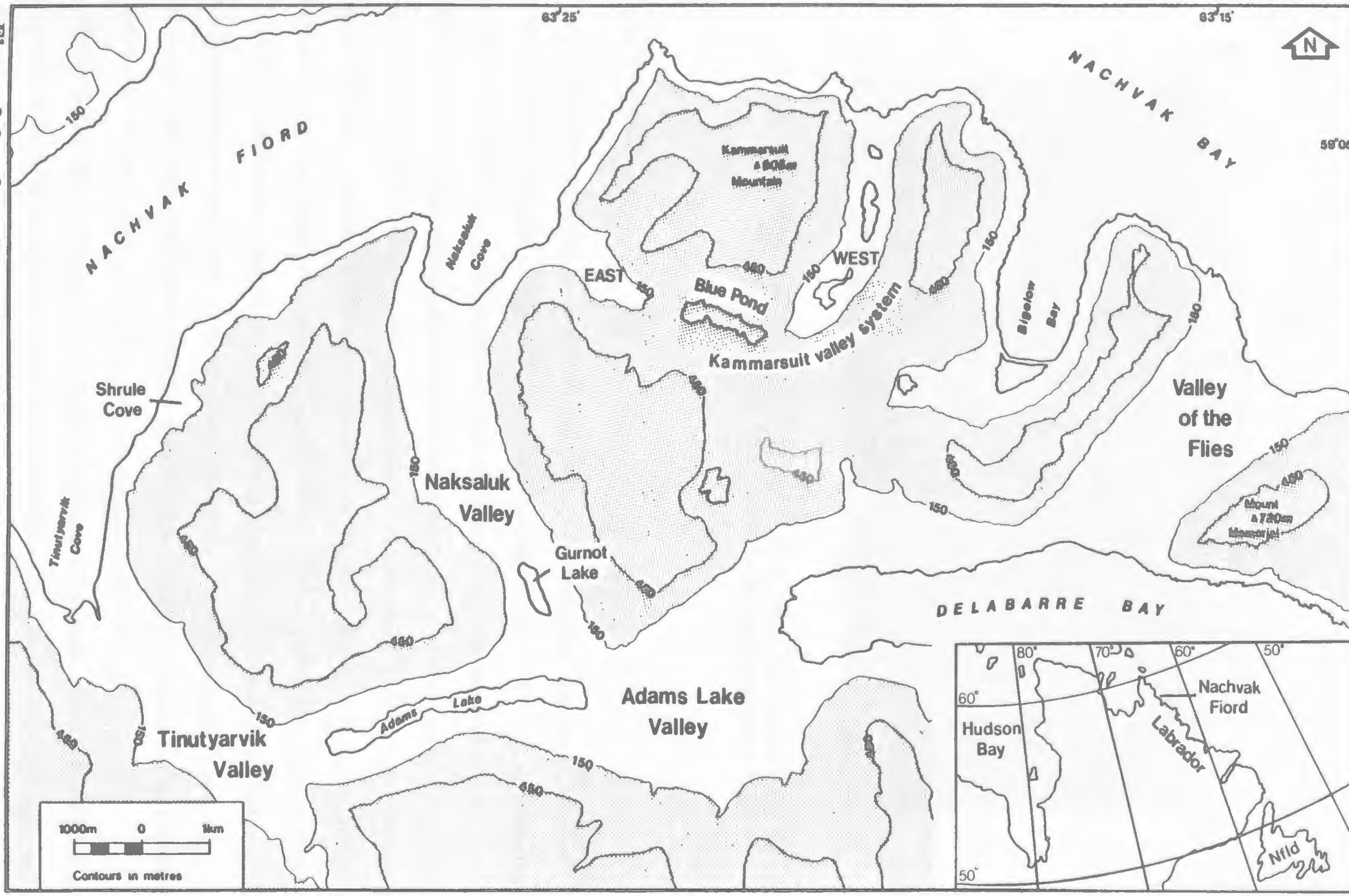

Figure 2-2: Outer Nachvak Fiord showing location of areas discussed in the text.

Figure 2-3: Oblique air photograph of Valley of the Flies, looking towards the northeast. Mount Memorial is in the right middleground.

2.2. Geology

The study area straddles the boundary between the Precambrian Nain and Churchill Structural Provinces (Morgan, 1972, 1973, 1975; Wardle, 1983) (Fig. 2-4). East of Naksaluk Cove, the Nain Province, in upper amphibolite to granulite facies, consists of a complex of Archaean tonalitic to granodioritic gneisses with minor remnants of supracrustals. The Churchill Province, to the west, is predominantly composed of Archaean rocks variably reworked during the Hudsonian Orogeny. The position of the Nain-Churchill boundary has been defined, on structural grounds, as the eastern margin of granulite facies gneiss by Greene (1970) and Taylor (1979). Morgan (1975) and Wardle (1983) redefined the boundary on the premise that it is not a single tectonic feature but rather a broad zone of crustal reworking. Wardle introduced a third structural zone, the Churchill Border Zone, in order to characterize the effects of this reworking, and to demonstrate the gradational nature of the boundary (Fig. 2-4).

The Nain Province is characterized by granitoid gneiss of plutonic origin. Wardle referred to this gneiss as Nachvak gneiss and recognized the inclusion of numerous relicts of early mafic

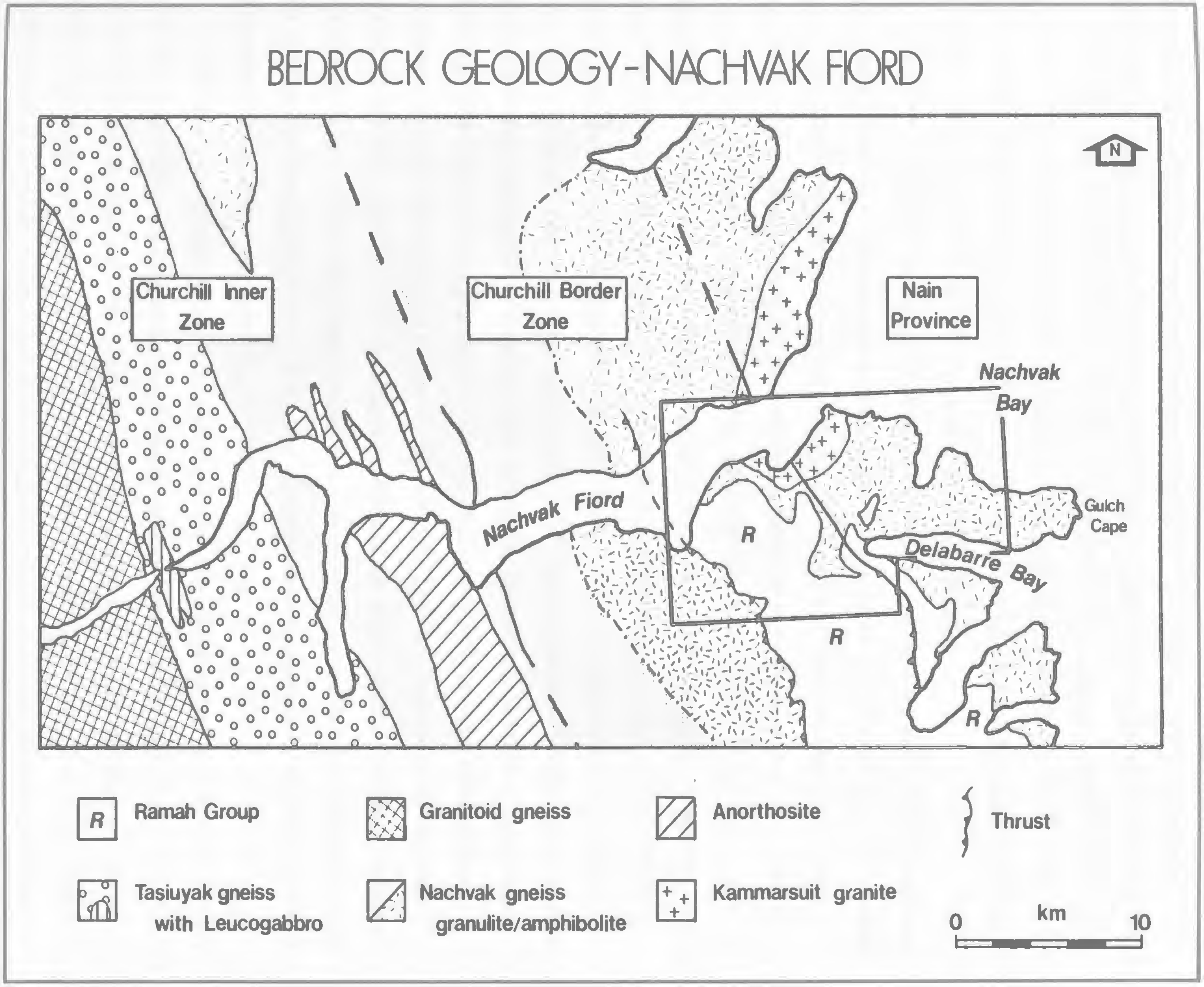

Figure 2-4: Generalized geology of the Nachvak Fiord region (*after* Wardle, 1983).

and supercrustal gneiss. Evidence of granulite grade metamorphism and regional deformation is well-preserved within the gneiss and has been dated at 2.8 Ga (Morgan, 1979). This date represents a minimum age for Nachvak gneiss. There is a general north-south trend to structures in the gneiss with steep to moderate westerly dips. Intrusion of a strongly foliated megacrystic granite, named Kammarsuit granite, occurred 2.7 Ga (Collerson *et al.*, 1982). A late dyke swarm with an east-west trend truncates structures in the host Nachvak gneiss and Kammarsuit granite. Effects of Hudsonian orogenesis are minimal in the Nain Province.

In the Churchill Border Zone, Nachvak gneiss differs from that of the Nain Province in the greater development of sheared, straightened and mylonitic gneiss and is overlain unconformably or along a thrust by the lower Proterozoic Ramah Group. Prior to deposition of the Ramah Group and the Hudsonian Orogeny, an east-west trending dyke swarm intruded the Archaean basement. The Ramah Group forms a linear north-trending fold belt within the zone, and is thrust at a low angle over the Archaean basement along faults which steepen to the west. It is a sedimentary succession consisting of a lower, shallow water siliciclastic sequence overlain by a deep water argillite-carbonate sequence (Morgan, 1975; Knight and Morgan, 1981). Extensive cleavage, folding and thrusting in these sequences attest to the homogeneous deformation which occurred during the Hudsonian Orogeny. The underlying Archaean gneisses were also locally reworked and sheared during this event. Morgan (1975), as a result of these observations, suggested that the Ramah Group lay within the Churchill Province, and therefore the Nain-Churchill boundary must lie along the eastern margin of the group. Wardle (1983) described a retrograde isograd which divides the border zone gneisses into a granulite facies in the west and an amphibolite to greenschist facies in the east. This division corresponds to the Nain-Churchill boundary as defined by Taylor (1979).

The Archaean gneisses of the Churchill Inner Zone, which extends west of Ivitak Cove, are also referred to as Nachvak gneiss but are mainly in granulite facies. Ductile north-south trending shear zones and (semi)brittle, east-directed, west-dipping thrusts are characteristic structural features of the gneiss in this zone. The western limit of Nachvak gneiss is formed by a sheared contact with a strongly-lineated blastomylonitic gneiss which was named Tasiuyak gneiss by Wardle (1983). Hudsonian deformation increased in intensity from east to west towards the contact with the Tasiuyak gneiss and has resulted in pervasively straightened and mylonitized gneisses. Linear bodies of deformed anorthosite occur in the eastern Churchill Province and are thought to be Archaean in age (Wardle, 1983). The regional post-tectonic dyke swarm that is evident throughout the Nachvak gneiss has been partially to completely transposed into the Hudsonian regional fabric in the Churchill Inner Zone.

2.3. Physiography

The peninsula of northern Labrador is dominated by the Torngat Mountains which have been described as "the highest and most rugged area of the eastern mainland of Canada" (Bostock, 1970, p. 15). The mountains are essentially the dissected rim of a differentially uplifted plateau which may be of pre-Paleozoic age (Ambrose, 1964). During the upper Cretaceous and early Tertiary, Labrador Sea rifting occurred leaving the steep slopes of northern Labrador and Greenland separated by a depressed basin (Bird, 1972). In the Tertiary period fast-flowing rivers eroded deep valleys along the Labrador coast which were subsequently occupied by outlet glaciers during the Pleistocene and modified to fiords. The straight east-west orientation of many Labrador fiords and valleys suggests control by structure, possibly transverse faults formed during Labrador Sea opening. West of the drainage divide, elevations are much lower and gradually decrease towards Ungava Bay.

Pleistocene glaciation was undoubtedly responsible for the most striking modifications to the upland plateau. The large U-shaped troughs which held continental ice streaming to the Labrador Sea from ice centres farther west are deeply incised into the palaeolandscape. Cirques, arêtes and horns also attest to the prolonged action of local 'Alpine' glacial erosion and periglacial activity.

Evans (1984) recognised three broad physiographic regions from the adjacent Selamiut Range, south of Nachvak Fiord. This landscape classification can also be used to describe the present study area, although the areal extent of each region differs between the two locations.

The fretted mountain summits and felsenmeer plateaus form the dominant landscape in the Selamiut Range, but along the outer coast this physiographic unit is limited due to the lower relief of the area. Hence upland cirques and lower valley systems predominate, restricting felsenmeer spreads and fretted summits to isolated areas, for example, Kammarsuit Mountain (Fig. 2-5). Cirque basins are relict features in the study area (Fig. 2-5), unlike those in the Selamiut Range which host the only glaciers in continental North America east of the Rocky Mountains

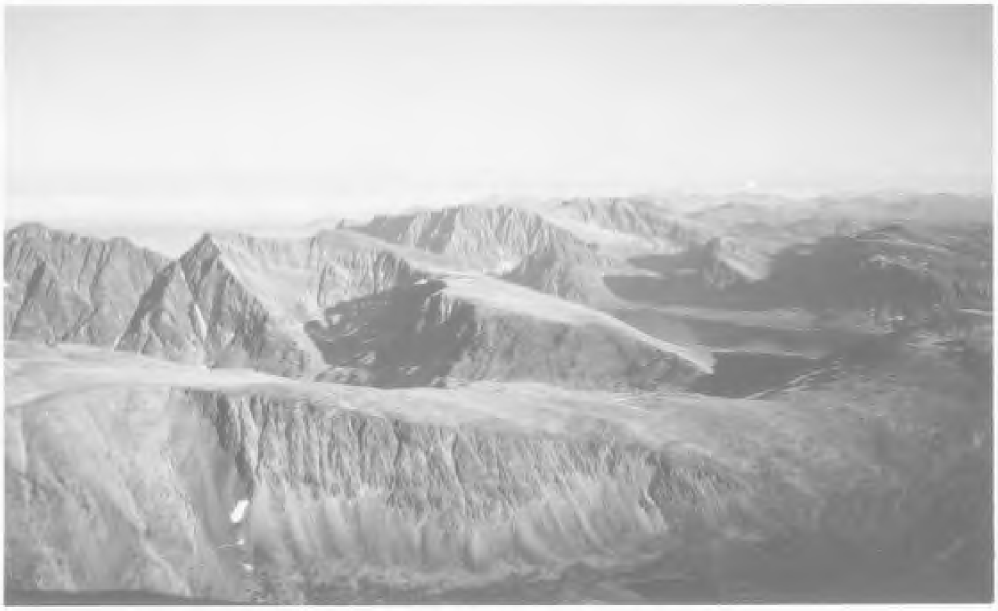

Figure 2-5: Photograph looking southeast along the Labrador coast from Kammarsuit Mountain. Note the plateau remnants, fretted summits and upland cirque in middleground.

(Rogerson, 1986). Clark (1984) calculated a mean altitude of 533 m $\pm$206 m for the floors of abandoned cirques in northernmost Labrador, and suggested that topography was a major influence on cirque basin altitude. All the cirque basins in the study area have a lower altitude than the mean value computed by Clark (1984) and the majority of the cirques have outlet valleys which are truncated by lower valley systems.

Adams Lake valley, which is oriented across the north-south structural trend of the region, has been modified by distributary valley glaciers moving southeastwards from the fiord through Tinutyarvik valley and east towards Delabarre Bay. Nachvak Fiord, which is also oriented in a general east-west direction, may represent a deeply modified pre-Pleistocene drainage system, subsequently exploited by outlet glaciers, or by selective linear flow within a continental ice sheet. The Kammarsuit valley system, to the east of Naksaluk valley, appears to result from local cirque glacier activity alone.

2.4. Previous research

The glacial landscape of northern Labrador was the subject of some of the earliest observations on Pleistocene glaciation in the eastern Canadian Arctic (Bell, 1885; Packard, 1891, Low, 1895; Daly, 1902). Despite this pioneer research, large gaps exist in our knowledge of the glacial history of the region. In Canada's contribution to the *IGCP Project 24* (Fulton, 1984) - a document which presented the current understanding of the Quaternary stratigraphy and history of Canada in 1984 - the following statement introduced the section on northern Labrador; "The glacial chronology of this region is poorly understood..." (Andrews and Miller, 1984). Many factors have contributed to this situation, but primarily it is the result of too few researchers working in a vast, remote and mountainous region. Prior to the last few decades much of the research was reconnaissance in nature, resulting in qualitative observations which are difficult to compare between study areas. Broad generalizations, based on equivocal data, have nurtured conflicting viewpoints which have alternately gained popularity in the literature. Further complexities have been introduced due to the extrapolation of chronologies from Baffin Island which are rarely tested through local, intensive research in northern Labrador. Consequently, little is known exactly about the Quaternary history of the region and new initiatives tend to be overshadowed by the presumptuous, yet intellectually attractive, debates of the past.

This section outlines, in chronological order, the research contributions that have formulated current theories on the glacial history of northern Labrador. These contributions can be subdivided into three specific topics or debates: Pre-Wisconsinan and Wisconsinan glaciations; Weathering zones and glaciation; Late Quaternary chronologies in northern Labrador.

2.4.1. Pre-Wisconsinan and Wisconsinan glaciations

The late 19th century geographers and geologists who explored the coastal areas of northern Labrador were convinced that regional ice was restricted in extent during the last glaciation. Both Bell (1885) and Low (1896) concluded that the upper elevations of the Torngat Mountains were never glaciated, whereas Daly (1902) suggested that the last regional ice stream was restricted to elevations of less than 650 m above sea level. Below this altitude, he recorded abundant "marks of former glacial activity", while above it these marks were absent. Daly (1902) argued that the presence of felsenmeer on mountain summits indicated that these areas existed as nunataks above the limit of regional and local ice masses. Inferred from Daly's commentary is the suggestion that the Wisconsinan maximum ice limit was 650 m above present sea level in the Nachvak Fiord area (Daly, 1902, pp. 245-251). Coleman (1921) agreed with the interpretations of Bell, Low, and Daly, but suggested that the unglaciated zone may extend at least 80 km inland at Nachvak Fiord. He attributed the Wisconsinan tills that he observed in the valleys of the fiord to deposition by local glaciers, and suggested that the unweathered character of some of the deposits indicated local Late Wisconsinan ice advances ["toward the end of the Glacial period"] (Coleman, 1921, p. 26).

The suggestion that large areas of terrain remained unglaciated during the Wisconsinan period was subsequently supported by the biogeographical research of Fernald (1925). He applied the 'Nunatak Hypothesis' of Blytt (1876) to the flora of Labrador and concluded that the disjunct distribution of certain plant species on the high mountain summits could be explained if these areas had remained as nunataks throughout the last glaciation.

Odell (1933, 1938) visited the Torngat Mountains on a scientific expedition with Dr. Ambrose Forbes in 1931. On the basis of limited observations, Odell concluded "that during the greatest advance of the ice from the Labradorean center, the Torngat and Kaumajet ranges were both entirely submerged" (1938, p. 206). Odell's evidence for total ice inundation of these areas was the presence of "ice-polished rocks" at 1500 m above sea level (asl) on one summit in the Torngat Mountains. However, he also disclosed that "the somewhat weathered condition of this rock surface at the time showed that it might possibly be due to an ice advance earlier than the last part of the Pleistocene epoch, commonly referred to as the Wisconsin" (1938, p. 205).

Odell (1938) further suggested that the extensive felsenmeer on the mountain summits could be accounted for "since the lapse of the glacial epoch", and that the disjunct distributions of certain species on these summits may reflect colonization after the ice receded. Tanner (1944) publicly supported Odell's hypothesis following a discussion in Odell's home in Cambridge, England in 1937, and on his own subsequent visit to Labrador.

The popularity of Odell's 'maximalist viewpoint'[1] was enhanced by the subsequent support of the greatly-respected Richard F. Flint who developed a theory for the growth of the Laurentide

[1]'Maximalist viewpoint' - a term used by Ives (1978) to describe the hypothesis of a last ice sheet which covered the Torngat Mountains and extended to the edge of the continental shelf.

Ice Sheet. Flint (1943) proposed that the ice sheet originated as valley glaciers in the mountains of Labrador and Baffin Island, and expanded through nourishment by warm, moist air masses. He suggested that ice extended out into the Atlantic Ocean on the eastern side of the ice divide, presumably overtopping many of the mountain summits. Although this theory was instrumental in consolidating the 'maximalist viewpoint', it could not be "fully support[ed] ... with field data" (p. 357). Despite this weakness, the idea of an all-encompassing Wisconsinan ice sheet became entrenched in the literature through the monumental publications of Flint (1947, 1959, 1971).

2.4.2. Weathering zone debate

During the middle and late 1950s, McGill University initiated an intensive study of the glacial geomorphology of northern Labrador and parts of Baffin Island. The major result of this research was the recognition of vertical differences in the degree of weathering and preservation of glacial landforms[2]. J.D Ives published a series of papers between 1957 and 1963, outlining the results of fieldwork conducted in Labrador (Ives, 1957, 1958*a*,*b*, 1960, 1963). Although Ives (1957) initially believed that continental ice completely inundated the Torngat Mountains at the Wisconsinan maximum, he modified this idea with the introduction of the concept of "weathering zones" (Fig. 2-6). Weathering zones have been defined as "units of the land surface which are distinguishable from each other on the basis of the distinct weathering features that record different lengths of time through which they were formed" (Dyke, 1977, p. 40).

The upper weathering zone, termed the Torngat zone, was characterized by mountain-top detritus or felsenmeer, and contained possible glacial erratics (1958*a*, p. 53). A lower zone, termed the Koroksoak zone, consisted of glacially scoured bedrock with perched boulders and sections of kame terraces and lateral moraines. A subzone within this zone, later termed the Saglek zone (Löken, 1961; Ives, 1963), was completely masked with glacial and fluvio-glacial deposits and had a broad, distinct kame terrace and lateral moraine complex at its upper level. Ives (1958*a*) concluded that the Torngat zone was glaciated in the pre-Wisconsin and that the Wisconsinan maximum was represented by the upper boundary of the Koroksoak zone. The major trimline within this zone was considered to be a recessional stage from the Wisconsinan maximum.

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[2]It should be noted that some of the most meticulous and intensive research on vertical weathering zones in the Torngat Mountains had been undertaken by Daly (1902, pp. 245-251).

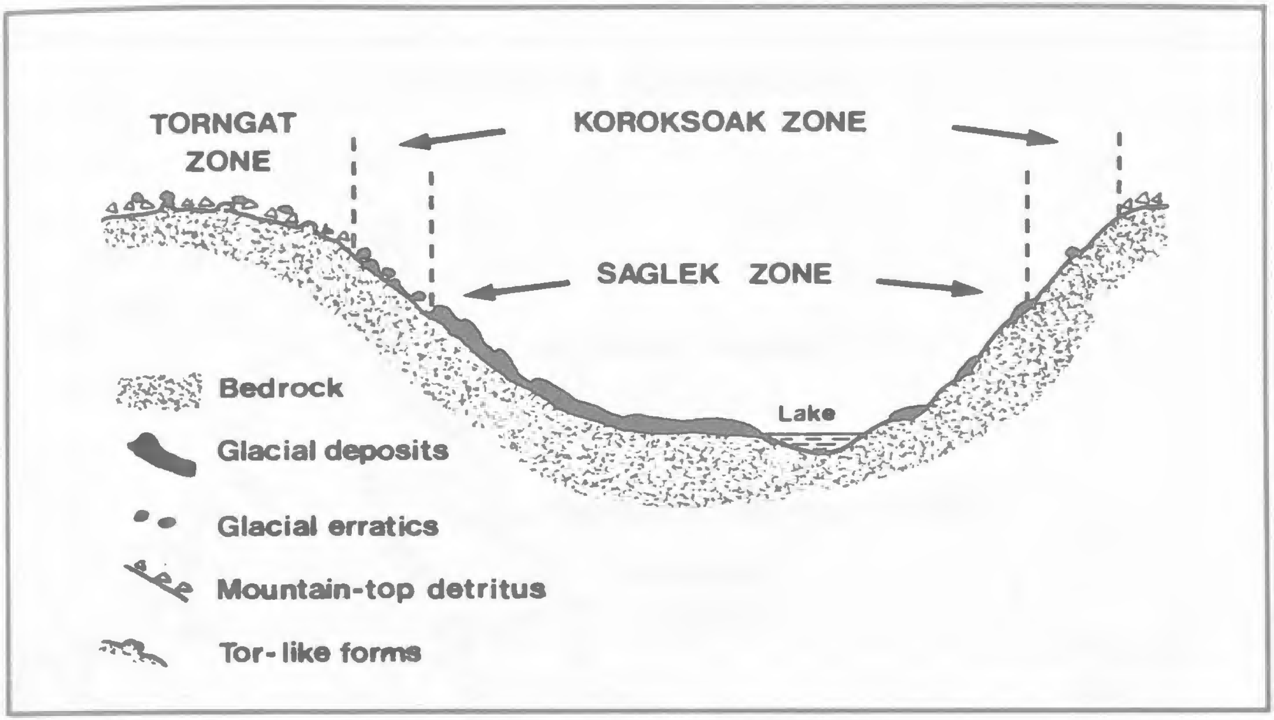

Figure 2-6: Schematic cross-section of glacial trough in the southern Torngat Mountains with location and character of weathering zones as originally described by Ives (1958*a*, 1963).

Löken (1961, 1962*a*) recognized three weathering zones "identical with the vertical zonation found by Ives" (1961, p. 192), in the Ryans Bay and Eclipse Channel areas (Fig. 2-7). He redefined the lowest zone [Saglek zone] as a separate zone from the Koroksoak zone and proposed that the three distinct levels which slope seaward, represent glacial stages. Both Löken (1962*a*) and Ives (1963) suggested that the Saglek zone was glaciated during the Saglek glaciation which was believed to be the 'classical' Wisconsin or Late Wisconsinan maximum. The Koroksoak zone was considered to be 'pre-classical' Wisconsin, but a pre-Wisconsinan age was thought improbable. Löken (1962*a*) questioned the identification, by Ives, of possible erratics in the Torngat zone, suggesting that they may have been weathered-out inclusions. Löken implied that this zone may never have been glaciated, but suggested "that if it were, it must have been in pre-Sangamon time or earlier" (1962*a*, p. 113). Johnson (1969) observed three weathering zones in the central Nain-Okak Bay region which he equated with the zones described by Ives. However, he concluded that the Koroksoak zone represented the highest stand of glacial ice during the last glaciation and that the Saglek level was a recessional stage from this maximum.

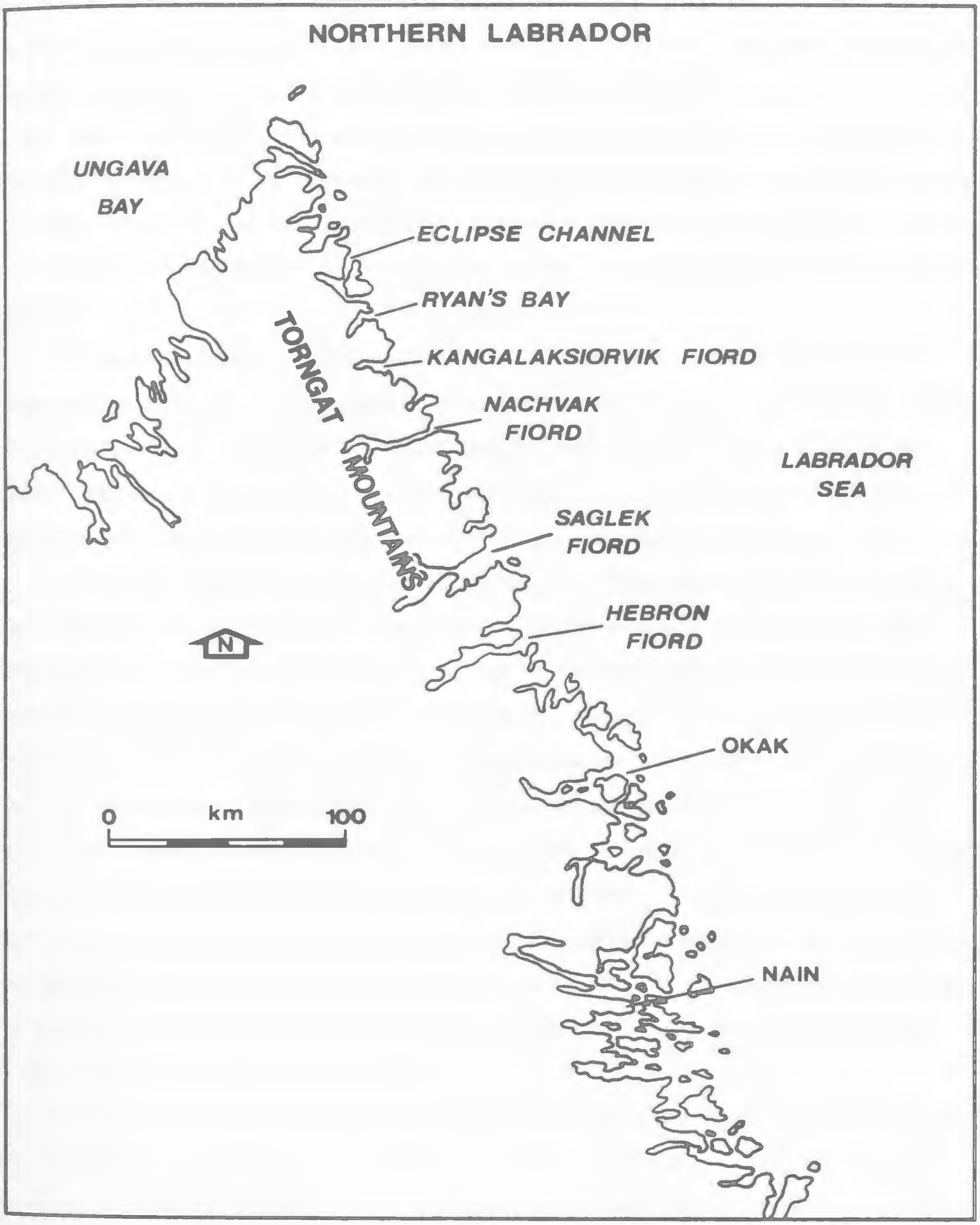

Figure 2-7: Location of study areas in northern Labrador discussed in the text.

By the early 1960s, the 'minimalist viewpoint' regarding Wisconsinan glaciation in Labrador had been re-established. Although Ives provided a major impetus for this 'wind of change', the work of Löken (1961, 1962*a*), Andrews (1963), Tomlinson (1958, 1963) and later, Johnson (1969) demonstrated that the concept could be applied throughout northern Labrador. During the middle and late 1960s there was a rapid development of research programs on Baffin Island at the expense of studies in northern Labrador, and hence subsequent modifications of the weathering zone concept were applied to Labrador from observations made in Baffin Island. Correlations between these two regions were based on similar degrees of weathering and vertical separation of the zones.

Pheasant and Andrews (1973) and Boyer and Pheasant (1974) recorded three vertical weathering zones in eastern Baffin Island and suggested that these zones may represent regional stratigraphic units, and therefore they could form the framework for a regional Quaternary history. Pheasant (1971) observed that the zones slope seaward and that these slopes are compatible with those of large outlet glaciers with basal shear stresses between 0.5 and 1.0 bar.

Attempts to date the observed weathering zones in relative and absolute terms began in the late 1960s and early 1970s on Baffin Island. Löken (1966) recognized that the weathering characteristics of the Koroksoak zone were similar to those of a bedrock knoll which had stood as a nunatak during a glacial phase dated at >54 ka BP. Andrews (1974) reported an Uranium series date of 137 ±10 ka BP from fossiliferous sediments associated with the outermost moraine that forms the boundary between the Saglek and Koroksoak zone equivalents in outer Narpaing Fiord, Baffin Island. Consequently it was suggested that the Saglek glaciation and the youngest Zone (III) of Pheasant (1971) embraces the Wisconsinan period. The age of the Koroksoak glaciation (Zone II, Pheasant, 1971), based on the extrapolation of various weathering parameters, was tentatively estimated at between 200 ka and 500 ka BP. The abundance of clay minerals and the increased percentage of ferric oxide in deposits from the Torngat zone indicated a possible age of 400 ka to 1000 ka BP (Andrews, 1974).

The interpretation of the uppermost Torngat zone, which Ives (1963) had argued was at one time glaciated, was modified by Ives (1974, 1975, 1978) and Ives *et al.* (1976), through the recognition of a fourth weathering zone, the Komaktorvik zone (Fig. 2-8). This zone consisted of scattered erratics lying on a surface that was intensely weathered, whereas the higher Torngat

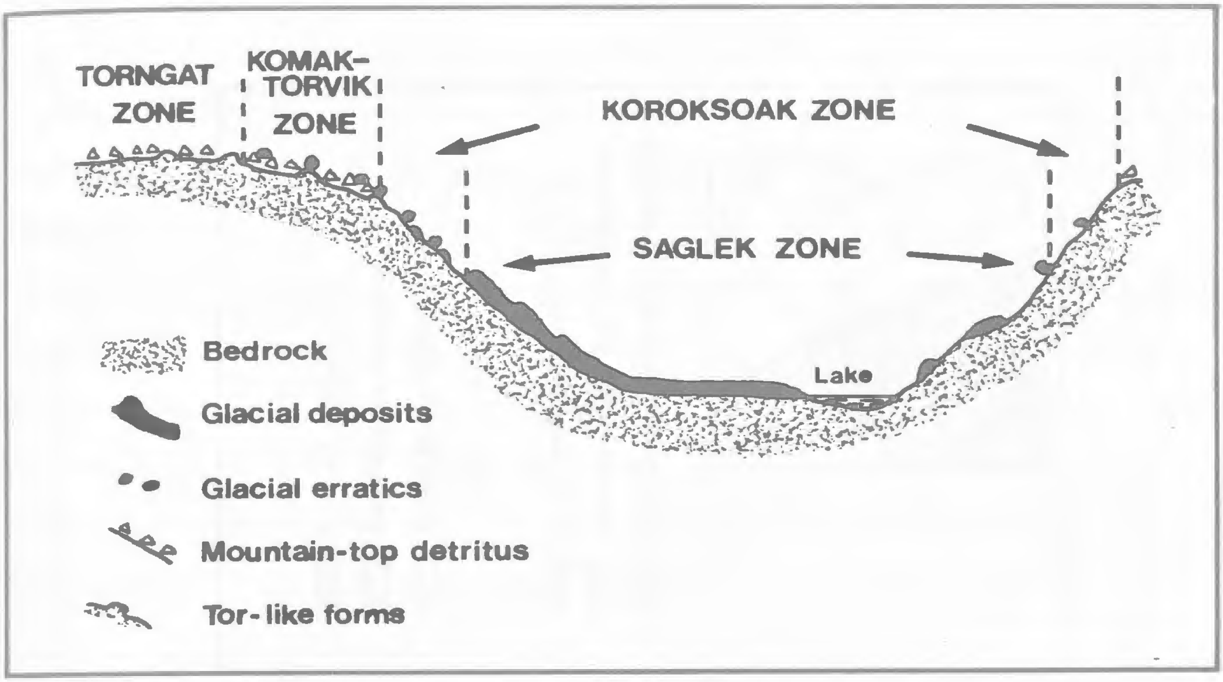

Figure 2-8: Schematic cross-section of glacial trough in the central Torngat Mountains, showing the significance of the Komaktorvik weathering zone (*after* Ives, 1978).

zone was marked by extensive mountain-top detritus and the presence of tors. The Komaktorvik zone encompassed a collection of summits which received erratics from an extensive pre-Wisconsinan glaciation (also predating the Koroksoak glaciation), while the Torngat zone was thought to be unglaciated (Fig. 2-9).

During the mid 1970s, the development of theoretical ice models based on assumed maximal ice cover overshadowed previous research supporting the weathering zone concept and a restricted Wisconsinan ice advance. Sugden (1974, 1976, 1977, 1978) proposed a model of selective linear erosion for the Laurentide ice sheet, a theory first developed by Tarr (1900) to explain the form of mountain summits in Maine. Sugden's model provided for the overdeepening of glacial valleys by active warm-based ice and the protective covering of high mountain summits and plateaus by cold-based ice. The latter is thought to be non-erosive since a frozen ice/rock boundary is an order of magnitude stronger than ice itself. The latter condition could preserve the felsenmeer observed on the mountain summits, but could also allow the emplacement of erratics at these elevations. The boundaries between weathering zones were interpreted as the points of

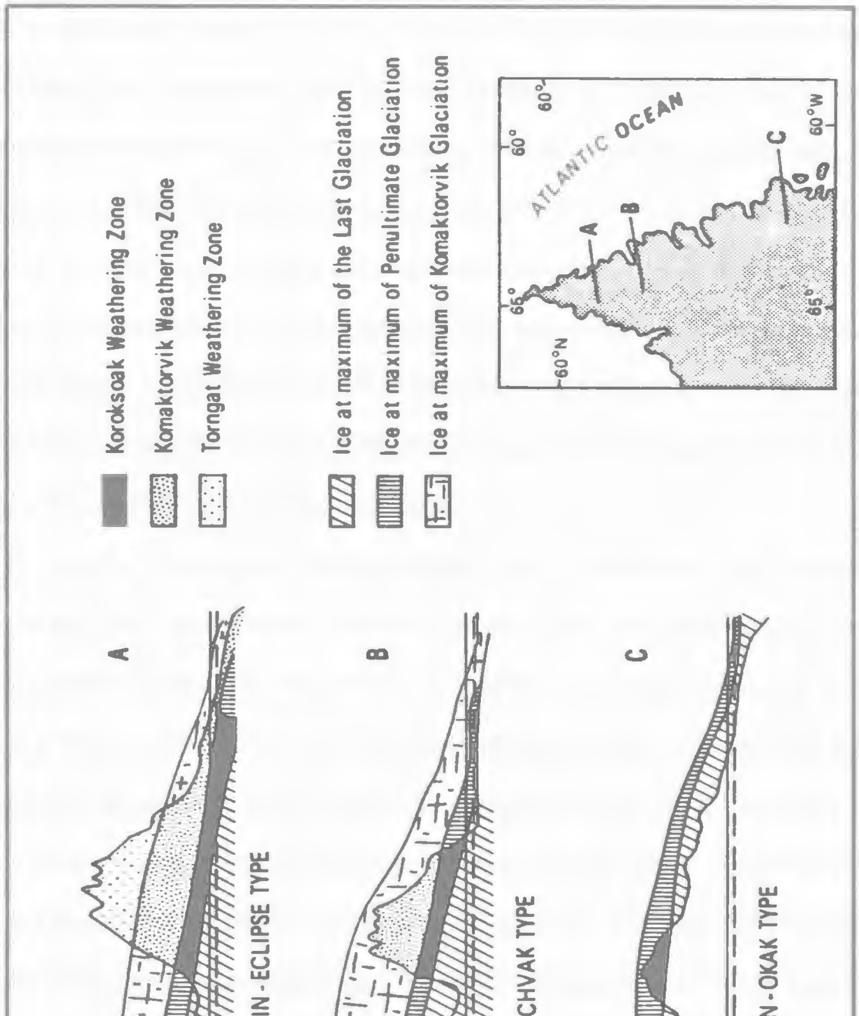

Figure 2-9: Application of the weathering zone concept from observations made at three locations along the Labrador coast (*after* Ives, 1975).

transformation from cold-based to warm-based glacier ice. Support for Sugden's theory was demonstrated by Sugden and Watts (1977) in eastern Baffin Island where some tors in upland areas had 'roche moutonnée' forms and glacial erratics in these areas were unweathered. They concluded that the Torngat zone was probably inundated by Laurentide ice during its maximum.

Hughes *et al.* (1981) and Mayewski *et al.* (1981) favoured Sugden's model and suggested that weathering zones represented erosion or protection by differing subglacial thermal regimes. They argued that the interpretation of weathering zones as ice limits requires the identification of ice marginal features, such as moraines, at the boundaries between zones. Ives (1978) responded to this theoretical model by suggesting that it was feasible, provided that postglacial development of felsenmeer was not required. In addition, he felt that the sharp boundary between the Komaktorvik and Koroksoak zones in the southern Torngat Mountains may be too distinct to be caused by a difference in subglacial thermal conditions.

Gangloff (1983) may be interpreted as supporting the 'maximum Wisconsinan' viewpoint. He suggested that tors and tor-like forms, observed in the Torngat Mountains, are not indicative of unglaciated areas, pointing out their similarity to forms close to present sea level in areas clearly covered by Late Wisconsinan ice. He further suggested that weathering features on bedrock do not necessarily date from interstadial or interglacial periods. He also suggested that felsenmeer is glacial till reworked by periglacial processes, on the basis of identifying sand-sized micro-erratics in both till and felsenmeer. However, he gave no account of the complex bedrock in the area, nor a definition of erratic material. These conclusions are based largely on evidence of X-ray diffraction of clay minerals in felsenmeer and till samples which he notes are similar, with slightly better clay mineral development in tills at low altitude. The samples however, were not glycolated, and it is possible that apparent chlorite peaks are actually glacial flour of quartz composition.

Dating of weathering zones in northern Labrador through association with fossiliferous sediments has proved tentative due to the restrictive limit of radiocarbon age determinations. Ives (1977) reported a ^{14}C date of $42730 \pm^{6680}_{9770}$ years BP (DIC-517) from a shelly diamicton (origin uncertain), located beyond the outer limits of the Saglek zone in the Iron Strand area of northern Labrador (Fig. 2-7). He considered this date to be a minimum date for the deposit (Iron Strand drift), and consequently a minimum age for the Koroksoak weathering zone. He suggested that the deposition of Iron Strand drift predated the Wisconsinan maximum. Short (1981) obtained a date of 18210 ±1900 BP from basal sediments in a lake dammed by the Saglek Moraine, north of Saglek Fiord (boundary between the Saglek and Koroksoak weathering zones). Ives (1981) commented that due to the very small sample (0.105 gm carbon), this value is considered to be a minimum date for the Late Wisconsinan maximum and a minimum age for the Saglek weathering zone. Implicit in this commentary is the assumption that the Late Wisconsinan maximum represents the maximum extent of ice during the Wisconsinan period.

Clark (1984) who worked in the Iron Strand/Kangalaksiorvik Fiord region, also proposed that the area occupied by Iron Strand drift correlated with the Koroksoak weathering zone. On the basis of ^{14}C dates and aminostratigraphical correlation with Baffin Island, Clark concluded that this glacial event occurred during the middle to early Wisconsinan period. Two Loon drift was interpreted to represent the Saglek weathering zone and was suggested to be Late Wisconsin in age.

Clark's assignation of Wisconsinan ages to both the Saglek and Koroksoak weathering zones represents a return to Ives' (1958*a*, 1963) interpretation of these regional chronostratigraphic units and their relative chronology. Clark's conclusions suggest that the Koroksoak glaciation represents the Wisconsinan maximum, and that this event occurred in the middle to early part of the period. By implication, the Late Wisconsinan ice advance was less extensive, but was not regarded as a recessional stage of the Wisconsinan maximum.

Evans and Rogerson (1986) reported a tentative chronology of glacial events in the Nachvak Fiord area of northern Labrador. Although they made no direct correlation with the weathering zones described for northern Labrador, they suggested that the Wisconsinan ice maximum (Ivitak phase) occurred during the early to middle Wisconsinan period, followed by a more restricted ice advance (Nachvak phase) during the Late Wisconsin.

Clark and Josenhans (1986) and Josenhans *et al.* (1986) proposed extensive Wisconsinan ice advances onto the Labrador Shelf. Iron Strand drift which lies within the Koroksoak weathering zone and which is believed to be Early to Middle Wisconsinan in age, was tentatively correlated with a till unit (unit 3a) that extends almost 200 km to the Labrador Shelf edge (Clark and Josenhans, 1986, p. 175). Another till unit (unit 3b) which overlies unit 3a on the shelf, has a less extensive distribution (30-50 km off the coast of the Torngat Mountains) and is confined to water depths greater than 160 m. This till unit is correlated with glacial sediments that compose the Saglek weathering zone, regarded by Clark and Josenhans (1986) as Late Wisconsinan.

Figure 2-10 is an attempt to diagrammatically present the evolution of the weathering zone concept as a chronostratigraphic framework for Quaternary glacial events in the eastern Canadian Arctic.

2.4.3. Late Quaternary chronologies

This section provides a more detailed account of localized studies in the Torngat Mountains, and attempts to draw comparisons from the various chronologies constructed for the individual areas. Reference to these studies will be made throughout the thesis and therefore the following is a summary of the pertinent discussions and arguments.

Since Ives' (1958*a*,*b*) initial descriptions of the Saglek weathering zone and the Saglek lateral moraine-kame terrace system in the Nakvak Lake area, north of Saglek Fiord, extensive reconnaissance has proven that Saglek level equivalents are widespread throughout the Torngat Mountains. Löken (1962*a*) recognized a Saglek level, separating the Saglek and Koroksoak weathering zones in the northern Torngat Mountains (Fig. 2-7). However, he noted that no lateral moraine-kame terrace system coincided with this level, as earlier described by Ives. Andrews (1963) recognized a Saglek level in the northern Nain-Okak area (Fig. 2-7) which was marked by "massive lateral moraines, sometimes 15 m high" (p. 161). Andrews further suggested that on the basis of air photograph interpretation and field observation, this level can be traced northward to the Saglek level type site in the southern Torngat Mountains. Johnson (1969) recorded "a well defined moraine system ...[which] appears to be equivalent to those marking the maximum extent of the Saglek Glaciation" (p. 391) in the central Nain-Okak Bay region. This moraine system was located at the boundary between the Saglek and Koroksoak zones.

Clark and Josenhans (1986) mapped over 100 Saglek Moraine segments and recessional moraines throughout the Torngat Mountains. They described these moraines as having "well-preserved, massive, high-crested morphologies that have undergone little post-depositional modification" (p. 172). In addition they described the soils on these moraines as having characteristically thin cambic B horizons ($<$10 cm thick) whereas soils on moraines within the Koroksoak weathering zone have $\geq$ 36 cm thickness of cambic B horizon. Clark and Josenhans

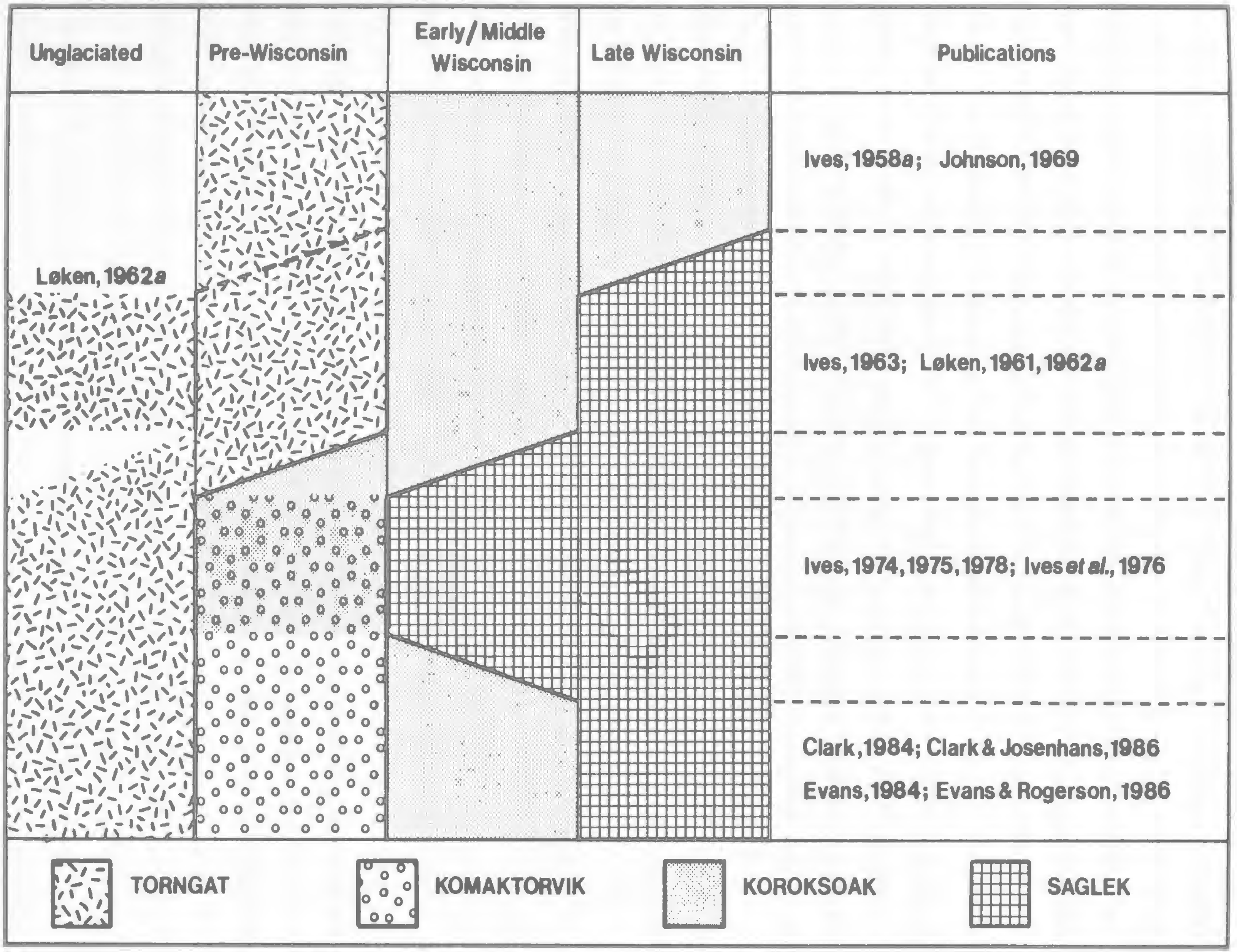

Figure 2-10: Evolution of the weathering zone concept and age interpretations of each zone.

(1986) suggested that the Saglek Moraines in the Torngat Mountains were deposited during the maximum extent of Late Wisconsinan Laurentide ice. Evans and Rogerson (1986) identified a moraine system which they considered to represent the Late Wisconsinan maximum in Nachvak Fiord. This moraine system is lower in altitude compared to the Late Wisconsinan equivalents of Clark and Josenhans (1986) in the same area, and soils were found to have thicker cambic B horizons. These latter data contradict the basic premise of the weathering zone concept which requires that the degree of weathering increase across weathering zone boundaries. There is obviously a clear discrepancy between the two chronostratigraphic frameworks for this area.

Results of these recent studies in the Torngat Mountains dispute the hypotheses of previous researchers who concluded that the Wisconsinan maximum occurred 14-18 ka BP (Ives *et al.*, 1976) and $<$25 ka BP (Andrews, 1963). Both Clark (1984) and Evans and Rogerson (1986) have estimated that the Wisconsinan maximum dates to the early or middle Wisconsinan period, 33-50 ka BP and $\geq$ 40 ka BP respectively. However their estimates for the Late Wisconsinan maximum concur with those of Ives (1976) and Andrews (1963) cited above, that is, $>$10 ka BP and 17-24 ka BP respectively. Clark and Josenhans (1986) and Josenhans *et al.* (1986) date their Late Wisconsinan maximum at 20 ka BP.

The eastward or seaward limit of the Late Wisconsinan ice advance in the Torngat Mountains is considered to be roughly along the outer coast with minor ice shelves occurring near the mouths of large valleys and fiords (Ives, 1976, 1978; Vilks and Mudie, 1978; Vilks, 1980; Josenhans, 1983; Clark, 1984; Johnson, 1985; Evans and Rogerson, 1986). This hypothesis is based on the pattern and magnitude of postglacial uplift along the outer coast (Evans and Rogerson, 1986), the identification of undisturbed pre-Late Wisconsinan sediments distal to the proposed ice limits (Ives, 1977; Clark, 1984; Johnson, 1985), and the recognition of thick accumulations of stratified sediment on the Labrador Shelf which may date to 40-80 ka BP (Vilks and Mudie, 1978). Vilks and Mudie also suggested, mainly on the basis of paleontological evidence, that sea ice cover in the Late Wisconsinan marine environment was similar to the present day, implying that the Late Wisconsinan ice sheet did not extend onto the shelf. They correlated three floral episodes, identified from palynological investigations of shelf sediments, with similar episodes recorded from Labrador lake sediments. The oldest tundra episode in the lake cores was recorded down to the top of highly inorganic sediments dated at 10.3 ka BP, whereas the lower part of the marine stratigraphy registers the continuous presence of sedge-shrub-tundra vegetation to at least 21 ka BP (Vilks and Mudie, 1978).

Fillon and Harmes (1982) described an extensive ice cover over the mainland of Labrador extending to the shelf edge as recently as the Late Wisconsin. Clark and Josenhans (1986) and Josenhans *et al.* (1986) have recently modified their earlier conclusions and have suggested that the Late Wisconsinan ice maximum is located 30-50 km off the coast of the Torngat Mountains. The latter hypothesis is dependant on the correlation of Late Wisconsinan glacial and postglacial sediment units on land with acoustic units identified on the Labrador Shelf. This correlation is based on "(1) acoustic stratigraphic continuity from fiord basins onto the shelf, (2) similarities in lithologic properties of the units, and (3) ages of the units suggested by radiocarbon dates" (Clark and Josenhans, 1986, p. 175). The continuity in acoustic stratigraphical records was not illustrated by Clark and Josenhans (1986) and, as Josenhans *et al.* (1986) have implied, the shallow inner shelf area causes a hiatus in the lateral tracing of acoustic units. The similarity in lithological composition between till units on land and on the shelf, as observed by Clark and Josenhans (1986), supports previous hypotheses that Laurentide ice extended through the fiords from an ice dispersal centre on the interior plateau of Nouveau Quebec-Ungava (Ives, 1958*a,b*; Löken, 1962*a*; Andrews, 1963; Johnson, 1969; Clark, 1984). However, this characteristic alone cannot discern Late Wisconsinan tills from older tills. The correlation of postglacial raised marine sediments on land with a stratified silt unit on the shelf is more significant because they contain similar ice-rafted carbonate lithologies (Clark and Josenhans, 1986). However, this probable correlation does not mean that the correlation of underlying sediments can be subsequently inferred. Radiocarbon dates from units on the Labrador Shelf have been exhaustively discussed (Fillon *et al.*, 1981*a*) and it was concluded that total organic matter dates were locally unreliable due to contamination by reworked terrestrial or marine carbon. Josenhans *et al.* (1986) proposed a general date of 20 ka BP for deposition and deglaciation of their extensive Late Wisconsinan till on the shelf and suggested that this date was open to interpretation, due to the inaccuracies in dating. Clark *et al.* (1985) reported free aIle/Ile ratios of 0.15 $\pm$.02 for *Hiatella arctica* shell fragments recovered from grab samples of the upper shelf till unit which they interpreted to be Late Wisconsinan in age. According to the aminostratigraphies constructed for northern Labrador (Clark, 1984) and Baffin Island (Miller, 1985), these ratios suggest an age of $>$28 ka BP and $>$40 ka BP, respectively.

Detailed studies of late-glacial and postglacial environments in northern Labrador have resulted in a number of localized chronologies which are predominantly based on relative frameworks, due to the paucity of dateable material in the region.

Löken (1962*b*) carried out a comprehensive study of recessional moraines and associated sea level stands in northernmost Labrador. He correlated the Noodleook readvance phase with distinct moraines along the outer part of Ryans Bay, and deduced from morphostratigraphical evidence that sea level stood at 41-56 m asl (Strandline [Sl] 1) during this ice advance. Moraines formed at Two Loon Lake were also associated with a readvance phase, and a suggested sea level stand of 31-45 m asl (Sl-2) was reconstructed for the area. The Kangalaksiorvik readvance phase was associated with a 28-34 m (Sl-3) sea level which Löken recognised as a transgressive event, and which was tentatively dated at 9 ka BP. He correlated the formation of moraines at the southern end of Two Loon valley and the Sheppard moraines along the Ungava Bay drainage divide with this readvance phase. A second marine transgressive event was recognized at 15m (Sl-4) above present sea level which he tentatively correlated with the 4700 year old Tapes transgression of Scandinavia. Various sea level stands were recorded below an elevation of 10 m above sea level. Shorelines Sl-1 to Sl-3 have progressively lower slopes and tilt up towards the SSW. Sl-4 is a horizontal strandline. Löken (1962*b*) drew isobases, on what he interpreted to be contemporaneous strandlines, to illustrate the late-glacial and postglacial emergence of the region during the final retreat of the Wisconsinan ice sheet.

Both Andrews (1963) and Johnson (1969) constructed local chronologies for late-glacial events in the Nain-Okak area. Attempts to correlate the details of their chronologies with studies carried out further north in the Torngat Mountains would result in an oversimplification of the complex interactions which existed during the retreat of Late Wisconsinan ice along the Labrador coast. However, the critical importance of their research to this present study is the recognition of a fluctuating late-glacial ice margin, retreating in association with high relative sea levels.

Clark (1984) reported ^{14}C dates of 8700 ±470 years BP (L-642) for deglaciation of outer Kangalaksiorvik Fiord and 8240 ±300 years BP (GSC-1560) for deglaciation west of the fiord head. A marine limit of 56 m above sea level was recorded in both Eclipse Channel and Kangalaksiorvik Fiord and was considered to have been formed immediately following deglaciation from the maximum ice position. This conflicts with Löken's (1962*b*) chronology which related a similar relative sea level to the earlier Noodleook phase. Local ice caps and cirque glaciers persisted after recession of Laurentide outlet glaciers and their associated drift sheets were dated at 9 ka BP and 2.5 ka BP (Clark, 1984).

Evans and Rogerson (1986) described three local glacial advances which occurred during the late-glacial and postglacial period in Nachvak Fiord. Superguksoak phase I is tentatively dated at 5-12 ka BP and has been correlated with a relative sea level of 33 m above present. This level represents the marine limit along the central portion of Nachvak Fiord. Superguksoak phase II cannot be clearly distinguished from phase I on the basis of pedogenic weathering criteria, however it was tentatively dated at 3 ka BP using lichenometry. Superguksoak phase III was assigned an age of $<$1.5 ka, also on the basis of lichenometry. Evans and Rogerson suggested that phases I and II may correlate with the Kangalaksiorvik readvance phase of Löken (1962*b*) and/or the outer moraines recorded by McCoy (1983) in the Cirque Lake basin, south of Nachvak Fiord. McCoy (1983) dated ten recessional moraines; the outermost one was tentatively suggested to represent the Late Wisconsinan maximum, dated at $\gg$4 ka BP by lichenometry. Various other phases dated between $<$150 years BP to $>$2.8 ka BP. All of the phases dated by lichenometry should be subject to some doubt since Rogerson *et al.* (1986*b*) report erratic lichen growth rates from 'growth stations' established five years earlier by McCoy (1983).

2.5. Context

The study area is situated in a coastal location where, according to Ives (1978), the Komaktorvik and Koroksoak weathering zones should reach sea level (Fig. 2-9). Although Ives (1978) proposed that the Komaktorvik glaciation (pre-Wisconsin) overtopped all the coastal summits in the Nachvak Fiord region, Coleman (1921) reported that erratics from west of the sedimentary Ramah Group are not present above 650 m in the study area (p. 35), thereby suggesting that the Torngat weathering zone exists above this altitude. Along the outer part of Nachvak Fiord, the presence of the Ramah Group provides conditions under which it should be possible to identify erratics from the Archaean formations farther west (Fig. 2-4). Such observations might provide conclusive evidence whether or not the coastal summits ever were glaciated.

The delineation of the Koroksoak weathering zone in the study area would allow comparison with the Iron Strand region of northernmost Labrador which is the only known coastal site in Labrador where this weathering zone has been given a limiting age by radiocarbon dates. Furthermore, the existence of a pre-Late Wisconsinan weathering zone near sea level would dispute the ice model of Clark and Josenhans (1986) which considers the study area to be at least 30 km west of the ice terminus and inundated by a 200-300 m thick ice sheet during the Late Wisconsinan.

The coastal location of the study area also provides an opportunity to record and analyse raised marine shorelines which can be compared with the shoreline elevations farther west in the fiord (Evans and Rogerson, 1986). Clark (1984) suggested that late-glacial isobases may run parallel to the northern Labrador coastline. Such an hypothesis may be tested by observations in outer Nachvak Fiord.

Chapter 3
Glacial landforms and sediments

3.1. Introduction

This chapter describes the form and character of terrestrial evidence relating to glaciation in the study area. Such evidence can yield important information concerning the origin, extent and style of glaciation. Geomorphic and geologic relationships provide a necessary framework for chronology reconstruction.

Prior to fieldwork, topographical maps and air photographs were used to demarcate landforms and to classify surficial sediment units. Groundtruthing and sampling of these were carried out during the field season, and when possible, air reconnaissance flights were used to verify information from inaccessible locations, particularly the upland areas.

3.2. Glacial landforms

This section provides a general overview of the glacial landscape in outer Nachvak Fiord and describes the location and morphology of glacial landforms identified in the area.

3.2.1. Glacial landscape

Two discrete components make up the glacial landscape of the study area. The higher of the two is regarded as a palaeolandscape, and is most clearly recognised on Gulch peninsula (Fig. 2-1). Here the remnants of truncated valleys and cirques occur with bottoms between 150 m and 300 m asl. South of Adams Lake valley is an area (*ca.* 13 km^2) of relatively flat, possibly areally-scoured terrain at 300-400 m asl. Eight cirques are identified in this palaeolandscape, all of which have a rock basin elevation $\geq$ 275 m asl, have truncated outlet valleys, and all but one are oriented in an east or northeast direction.

The lower component has been incised into the palaeolandscape and truncates many of the high level glacial valleys and cirques. Selective erosion of the palaeolandscape may have been associated with lines of weakness due to faulting, or it may have resulted from overdeepening of an ancient drainage system in the area. Adams Lake valley was modified by a distributary valley glacier moving southeastwards from the fiord through Tinutyarvik valley and east towards Delabarre Bay, across the north-south structural trend of the outer Nachvak Fiord region. Naksaluk valley and Valley of the Flies were probably also excavated by regional ice exploiting significant faults in the bedrock. The Kammarsuit valley system, to the east of Naksaluk valley is possibly the result of local ice activity alone. Four small cirques are identified in this landscape, each having an outlet valley which feeds down into the main glacial valley system. These cirques are oriented to the north (3) and to the southwest (1).

The age of these major erosional events is unknown. However, the suggestion that an ancient glaciation sculptured a landscape, much older than the existing one, has several important implications for palaeoenvironmental reconstruction in the study area. It is proposed that only those features which are exclusive to the most recent glacial modification of the landscape can be attributed to the last glaciation of the area. Therefore the presence of glacial erratics on the high mountain summits can only suggest that glaciation once occurred at this elevation, and not necessarily during the evolution of the present glaciated landscape. Furthermore, once these major troughs had been cut, thinner ice masses of regional or local origin would tend to flow along them and consequently, the existing landscape may have hosted a long history of glaciation.

3.2.2. Striations

Evidence of glacier movement is found in the form of striations which are sparsely preserved in the outer Nachvak Fiord area (Fig. 3-1). Generally they indicate an easterly flow of ice across the study area with local ice movement occurring in the Kammarsuit valley system. The orientation of striations confirms that Tinutyarvik and Adams Lake valleys acted as a distributary outlet for fiord ice. In Adams Lake valley lunate fractures occur in association with striations (Fig. 3-2).

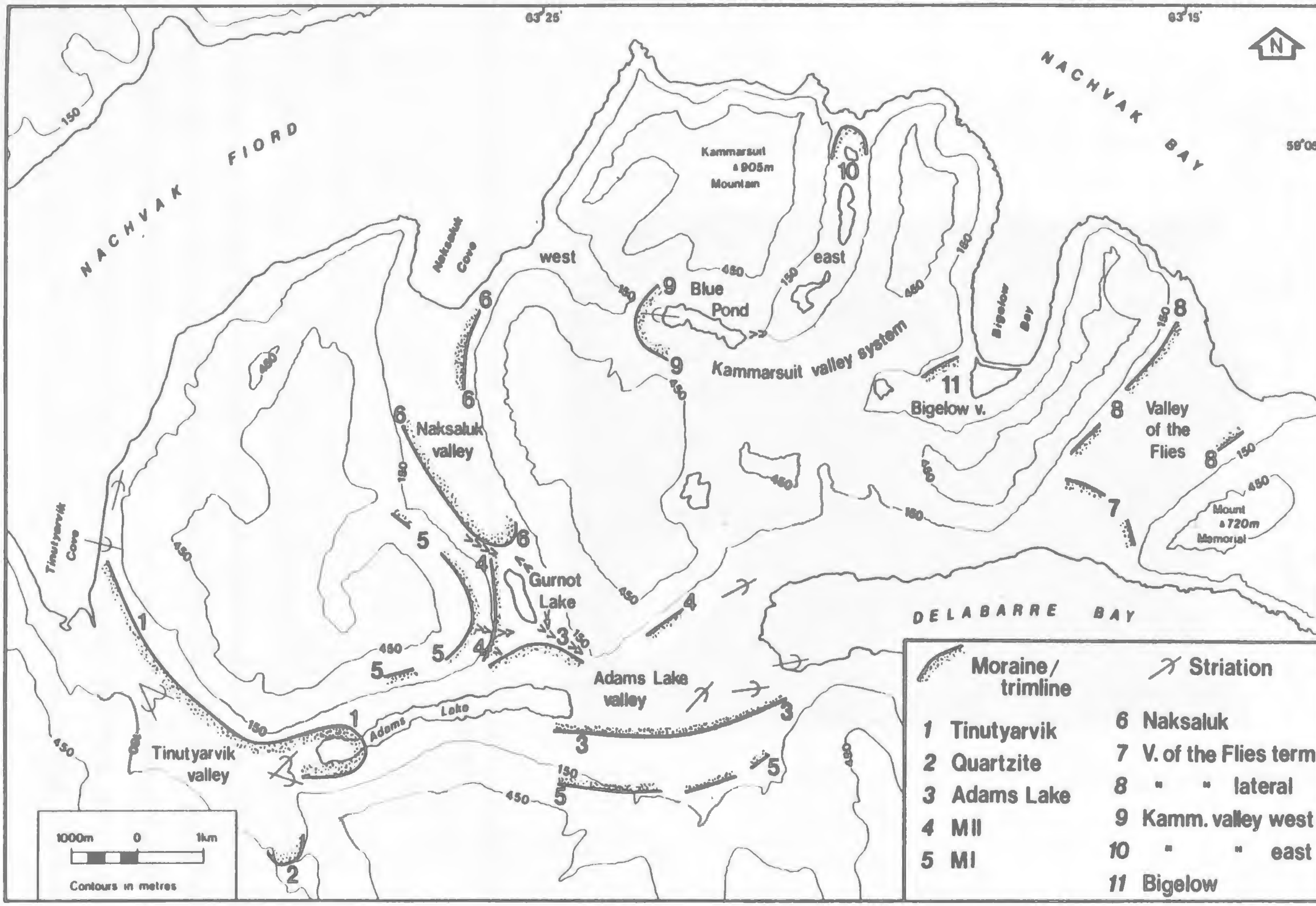

Figure 3-1: Location of striations, moraines, meltwater channels and former glacial lakes in the study area.

Figure 3-2: Photograph of striations and lunate fractures oriented northeast in Adams Lake valley. Mount Memorial is visible in the background.

3.2.3. Moraines and meltwater channels

Figure 3-1 shows the locations of moraines and associated landforms in the study area. At the head of Tinutyarvik valley is a large end moraine (Quartzite moraine) at 130 m aht[3] (Fig. 3-3). Incised into the moraine is a deep gully which presently allows drainage from farther south to flow into Nachvak Fiord, through Tinutyarvik valley. Along the eastern slopes of Tinutyarvik valley and at the western end of Adams Lake valley is Tinutyarvik moraine which appears to cut across Adams Lake as an end moraine. This moraine attains its maximum elevation of 106 m aht

[3]The elevation of this and other moraines is the average elevation of the moraine crest as measured by altimeter, expressed in metres above high tide (m aht).

in Adams Lake valley; however, the section in eastern Tinutyarvik valley has experienced extensive downslope displacement due to snowpatch development and mass movement (Figs. 3-3 & 3-4). Tinutyarvik moraine cannot be traced east from Tinutyarvik Cove along the fiord coastline which may be due to the steepness of the fiord walls and to mass movement of material down into the fiord basin.

At the southeastern end of Adams Lake a moraine segment, called Adams Lake moraine, is oriented parallel to the lake, at 78 m aht (Fig. 3-4). The subdued remnants of this moraine can be traced along the southern slopes of Adams Lake valley as far as Delabarre Bay where it was measured by level at 70 m aht. On the northern slopes of the valley is a corresponding moraine at 85 m aht which backfills the entrance to Naksaluk valley. This moraine has low relief and appears to have been planed along its lower central section.

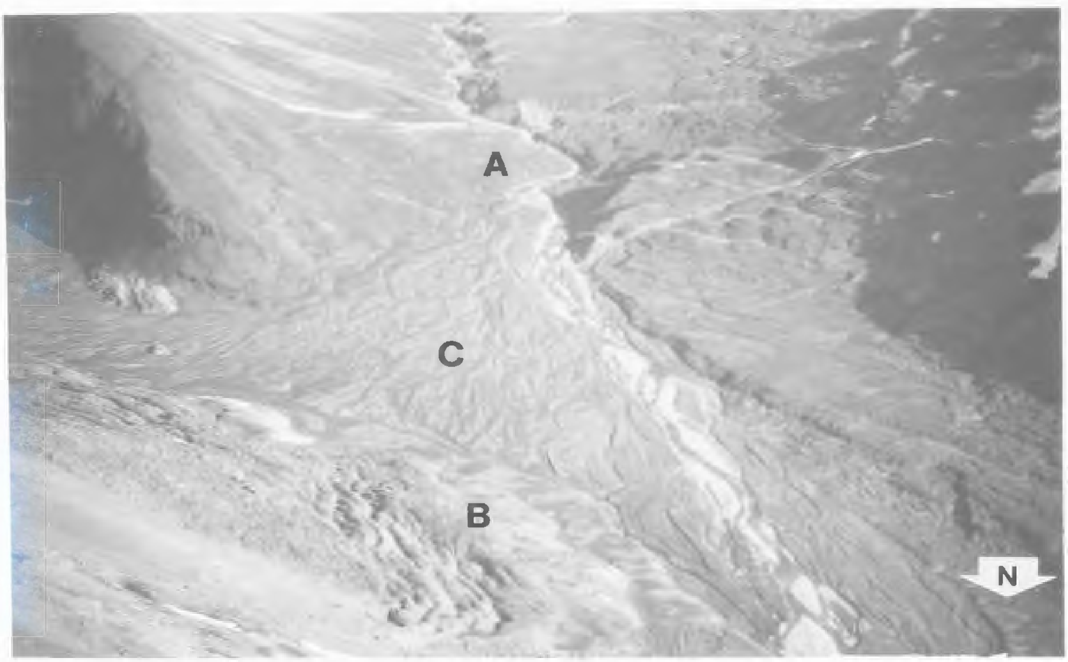

Figure 3-3: Photograph of southern Tinutyarvik valley. **A** Quartzite moraine; **B** Tinutyarvik moraine, displaying the effects of solifluction and mass movement; **C** Alluvial fan associated with the draining of a lake dammed by Quartzite moraine.

Directly above Adams Lake moraine on the southern slopes of the valley is a second lateral moraine which descends from 180 m aht to 140 m aht towards the east. This moraine is poorly preserved and is extensively shrouded by local scree accumulations on the steeper slopes (Fig. 3-4). A corresponding moraine segment is located on the northern slopes of the valley, above Adams

Lake and is considered to have a similar elevation based on abneymeter observations. The upper surface of a drift above Adams Lake, sheet which wraps around the southwestern shoulder of Naksaluk valley has an elevation of 170 m aht. This trimline (MI) extends into Naksaluk valley but cannot be traced along the steep slopes of the valley walls. The drift sheet appears to terminate abruptly *ca.* 500 m north of Gurnot Lake. Two major meltwater channels originate at this trimline and fall steeply to the valley floor (Fig. 3-1).

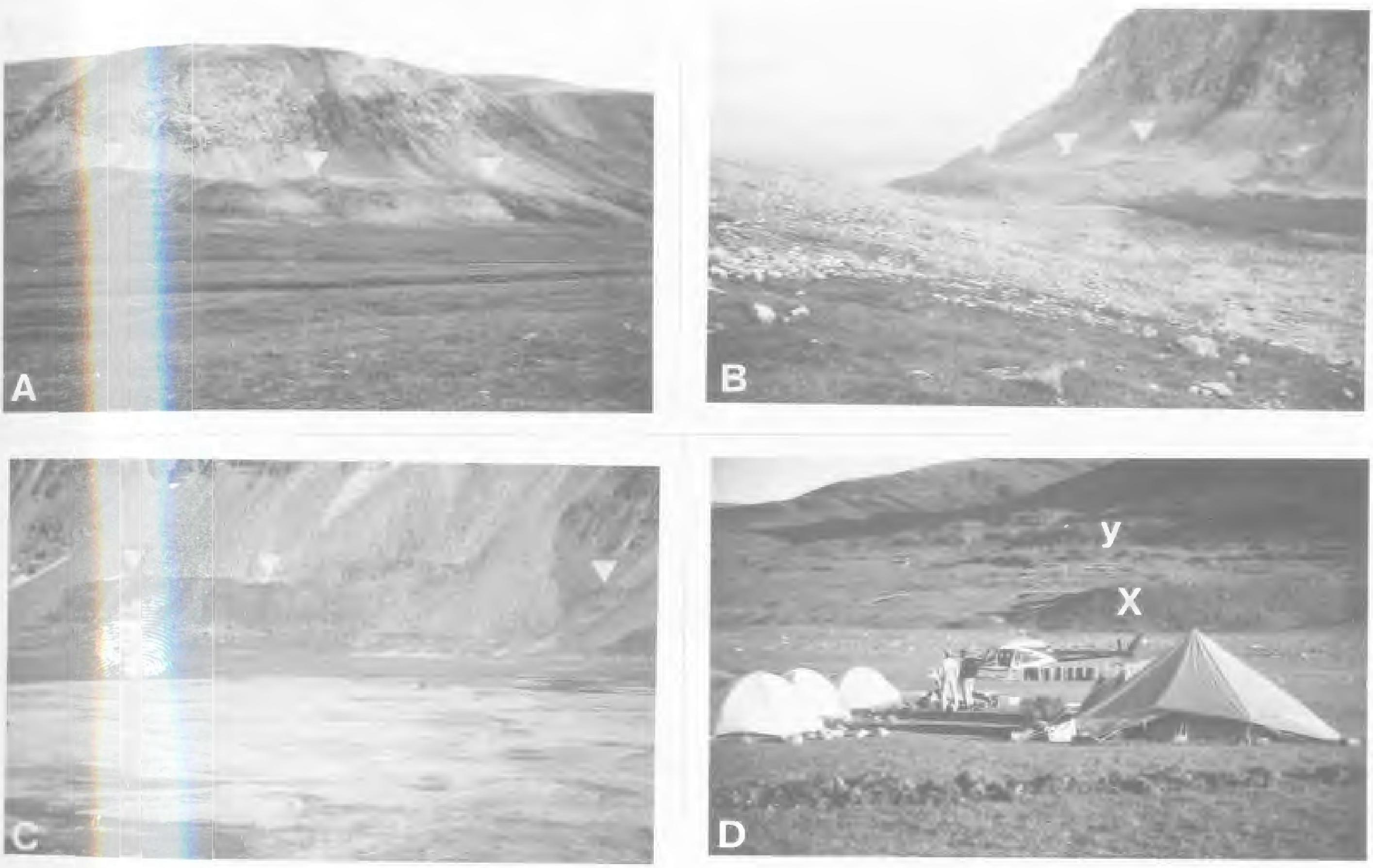

Figure 3-4: Photographs of **A** Tinutyarvik moraine; **B** Kammarsuit valley East moraine; **C** Valley of the Flies lateral moraine; **D** Adams Lake moraine (**x**), MI moraine in Adams Lake valley (**y**).

A lower trimline (MII) was recorded at 120 m aht on the western slopes of Naksaluk valley. This level decreases in altitude to the north where it is truncated by a lower moraine system north of Gurnot Lake. Associated with MII trimline is a meltwater channel which drops steeply from 120 m aht to 83 m aht, the elevation of Adams Lake moraine in this area (Fig. 3-1). On the eastern slope of Naksaluk valley, above Gurnot Lake is a meltwater channel which slopes down to the north and to the south from a maximum elevation of 120 m aht. The channel terminates at

87 m aht to the north of Gurnot Lake and converges in the south with outlet channels that drained into Adams Lake valley (Fig. 3-1). The meltwater channels associated with MI trimline are partially filled with material below the level of MII moraine. Possibly related to MII moraine is another moraine segment situated on the northern slopes of Adams Lake valley at an elevation of 116 m $\pm$10 m aht. The morphology of this moraine has probably been severely altered as it is a site of active nivation, solifluction and scree accumulation.

Figure 3-5: Photograph of Naksaluk moraine looking northeast towards Nachvak Fiord. Note also in the distance a segment of a complex moraine system which is interpreted by Clark and Josenhans (1986) as representing the Late Wisconsinan maximum of ice on the northern fiord margin. It is *ca.* 300 m asl, and in detail is found to comprise two moraines, the distinct upper one formed against ice coming out of the cirque, and the lower formed at the true fiord ice margin.

In the northern part of Naksaluk valley, a 1.5 km long moraine, the Naksaluk moraine, abutts against the western slopes of the valley and terminates north of Gurnot Lake where it truncates the MII trimline (Fig. 3-5). Its elevation is 80 m aht, and corresponding moraine segments are found at the same level on the eastern slopes of the valley. The morphology of this moraine is characterized by both a steep proximal (northeastern) slope which suggests deposition from a glacier flowing into the valley, south from Nachvak Fiord, and the occurrence of kettle-holes against its distal edge.

Valley of the Flies contains two moraine systems. Two segments of a subdued terminal moraine are oriented east-west across the valley floor at an elevation of 22 m aht. The form of this moraine suggests that it was deposited from an ice source in southern Valley of the Flies. The second moraine system consists of two lateral moraines on either side of the valley at 115 m aht. It is probable that solifluction and rock fall from the adjacent valley walls have considerably altered the morphology of these lateral moraines (Fig. 3-4). Within the Kammarsuit valley system, two terminal moraine complexes were recorded. Kammarsuit valley west has a terminal moraine straddling the valley at 180 m aht which causes the drainage from Blue Pond to flow south and east, through Kammarsuit valley east, to Nachvak Fiord. This moraine is associated with a lateral moraine which was observed on the valley slopes above Blue Pond. The mouth of Kammarsuit valley east is blocked by a moraine with a distinct, well-preserved morphology (Fig. 3-4) which is associated with a lateral moraine located on the western slopes of the valley. Elevations for the crests of the two moraines are 135 m and 160 m aht, respectively. The morphology of moraines in the Kammarsuit valley system suggests that local glaciers, originating in the mountains and flowing towards the fiord, were responsible for their deposition. A 105 m aht lateral moraine segment was recorded on the northern slopes of Bigelow valley.

3.2.4. Glacial lakes

Evidence for four glacial lake systems was recorded in the study area. A former proglacial, or moraine-dammed lake behind Quartzite moraine is suggested by lake shorelines south of the moraine, at 120 m aht, and its drainage is suggested by a large alluvial fan which appears to originate from a deep gulley through the moraine (Figs. 3-1 & 3-6).

In southern Naksaluk valley two lake shorelines occur in the vicinity of Gurnot Lake (79 m aht). The highest shoreline, at 87 m aht, extends along the eastern and western margins of the lake but does not occur to the north or south, as the land elevation is $\leq$ 86 m aht. It is suggested that this lake was dammed in the south by ice in Adams Lake valley. It is uncertain what may have supported this high lake level further north as a subsequent ice advance from the fiord into Naksaluk valley (depositing Naksaluk moraine) truncated and removed morainal material (MII drift) which may have dammed the lake and controlled the lake level elevation.

This 87 m lake shoreline was formed subsequent to the deposition of Adams Lake moraine in northern Adams Lake valley as the surface of this moraine is planed in many locations. Evidence suggesting lake discharge indicates that the lake emptied through two outlet channels which converge and flow into Adams Lake valley. One of the outlets occurs along the eastern margin of the lake, at 85 m aht, the other occurs in the southeast corner of the lake and presently contains the main discharge from Gurnot Lake. South of the existing lake margin is an area of fossil patterned ground containing sorted polygons, 2-3 m in diameter which may have formed on the bed of the former ice-marginal lake.

The lower lake shoreline, at 82 m aht, encompasses Gurnot Lake to the south and extends north along the sides of Naksaluk valley. The northern extent of the shoreline is obscured by talus. It is suggested that this lake was supported by ice in Naksaluk valley and that it existed after the deposition of Naksaluk moraine as southern segments of the moraine appear to be planed. The main outlet for this 82 m aht lake appears to be in the southeast along the present Gurnot Lake discharge channel.

A series of glacial lake systems existed in Kammarsuit valley west and Kammarsuit valley east, as evidenced by elevated lake shorelines (Fig. 3-6). Blue Pond (165 m aht) in Kammarsuit valley west previously had a water level of 175 m aht and discharged into the fiord through a gully cut in the end moraine, northwest of the lake. The bed of this channel was measured at 175 m aht suggesting that it was responsible for controlling the level of the lake. It is proposed that this high lake level was ice-supported southeast of Blue Pond, but subsequently emptied into Kammarsuit valley east, through an overdeepened channel (165 m aht), as the ice margin receded west into the mountain cirques.

In Kammarsuit valley east three sets of lake shorelines were recorded at 135 m, 130 m and 125 m aht and are regarded as evidence for a former proglacial lake dammed by the end moraine at the mouth of the valley. The highest level was controlled by the upper surface of the moraine which appears to be water-planed along its central portion. Subsequent lower lake levels represent discharge that formed a deeply incised gully in the moraine.

3.2.5. Discussion

On the basis of the above description of glacial landforms, several general comments can be made concerning the glacial history of the study area. The highest observed moraines in the area, excluding the Kammarsuit valley system, occur in Adams Lake valley and in Naksaluk valley (MI) at 180 m aht. Correlation between these moraines and the two lateral moraine segments (115 m aht) in Valley of the Flies is tentatively suggested, based on a calculated ice gradient of 6 m/km between the two areas. There is no indication of a correlative moraine in Tinutyarvik valley nor along the fiord coastline; however it is unlikely that high morainal material would be preserved on the steep slopes of the valley and fiord. It is proposed that these high-level moraines represent a significant regional ice advance, during which the main valleys of the study area acted as distributary outlets for fiord ice. The moraines located in Kammarsuit valley west (180 m aht) and Kammarsuit valley east (135 m aht) are considered to be local in origin because of their relationship to lateral moraines up-valley, and therefore they probably postdate this extensive regional ice advance.

A second lateral moraine occurs in southern Naksaluk valley (MII) at 120 m aht, and is correlated with a moraine segment at 116 m ±10 m aht in Adams Lake valley. This glacial phase may be either a readvance of regional ice following deglaciation from MI, or a separate ice advance postdating the MI glacial phase. MII is not considered to be a recessional moraine from MI as meltwater channels feeding down to the valley floor are partially filled with MII drift and this could only occur during a subsequent ice advance into the valley.

The lower lateral moraines identified in Adams Lake valley (Adams Lake moraine, 70-80 m aht) may represent a recessional stage from the MII glacial phase with an ice limit located at the subdued terminal moraine in Valley of the Flies (22 m aht). There is no obvious geomorphic evidence to suggest that the Adams Lake moraine represents an independent ice advance.

The cross-cutting relationship of MII moraine by Naksaluk moraine, north of Gurnot Lake indicates an ice advance from the fiord which postdates the MII phase. Tinutyarvik moraine in Tinutyarvik valley may represent deposition from a regional ice source which advanced from the fiord into the western end of Adams Lake valley. However, it is possible that Tinutyarvik moraine represents a recessional stage from the Adams Lake moraine phase. Neither alternative can be proven on the basis of geomorphic evidence alone.

The general framework of glacial events outlined above is tentative and serves only to provide a broad morphochronology which can be elaborated through other evidence of glaciation.

3.3. Surficial sediment units

This section provides a brief description of the major surficial units of the study area. Sediment unit classification is adopted from Dyke *et al.* (1982), and includes the following: bedrock, residuum, colluvium, raised marine and lacustrine deposits, active and inactive alluvium, till and moraine. Figure 3-6 generalizes the distribution of three of these units in the study area. This map is necessarily preliminary, reflecting the reconnaissance nature of data collection. The steep slopes marked on Figure 3-6 indicate the probable boundary between the bedrock and residuum units and the colluvium unit which is generally restricted to the base of these slopes. The approximate distribution of till and moraine can be inferred from Figure 3-1. All other surficial sediment units are distributed along the lower slopes and valley floors and are undifferentiated on the diagram, with the exception of raised marine and lacustrine deposits which are represented by areas inundated by the sea or previously occupied by glacial lakes.

3.3.1. Bedrock

The predominance of bedrock as a surficial unit in the study area is attributed to the recent glacial history of the area. Ice scouring in the valleys and cirque basins has resulted in precipitous walls and cliffs which are too steep for accumulation of surface sediment. Minor occurrences of bedrock, which outcrop in valley bottoms and on lower slopes, are the result of marine and fluvial processes removing the surface sediment veneer. Bedrock characteristics are described in the geology section of Chapter 2 (see Section 2.2).

3.3.2. Residuum

This surficial unit delineates the distribution of felsenmeer, glacial erratics and grus located on the upland plateaus and summits in the study area. It may represent areas that existed as nunataks during recent glaciation because it is situated above the highest moraines in the study area. The presence of erratics among the felsenmeer indicates that these areas were once inundated by regional ice. On the flat summits and plateaus, extensive periglacial activity has formed sorted and unsorted stone circles, turf-fronted lobes and altiplanation terraces.

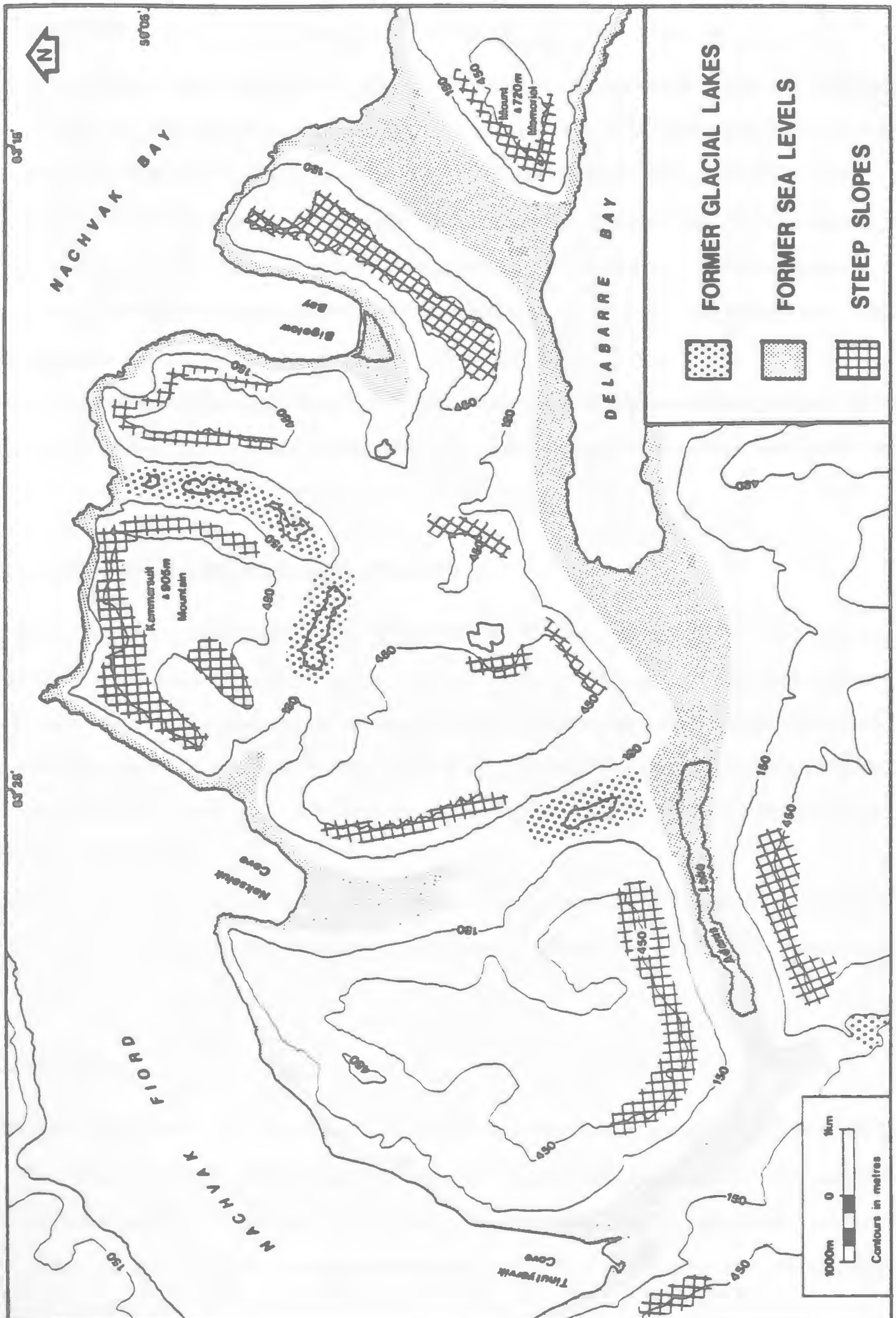

Figure 3-6: Generalized distribution of steep slopes, and areas formerly inundated by the sea or glacial lakes.

3.3.3. Colluvium

Along the lower slopes of steep bedrock walls, talus accumulation is extensive. Material consists of large angular blocks and coarse gravel with sand which is fall-sorted along the slope profile. Although the low-angled sections of many of these talus slopes appear to be relict features (showing colonization by vegetation), recent slurry flow tracks are evident on the steeper slopes, and minor rock falls were observed during the summer months. A large scale rock avalanche occurred in Kammarsuit valley east in relatively recent time, spreading large amounts of material across the valley floor between the two larger lakes.

Many lateral moraines in the study area have been extensively modified by highly active rock-glacierized screes and by the development of protalus ramparts along the base of the steeper slopes.

3.3.4. Raised marine and lacustrine deposits

Raised marine deposits in the study area occur as beach sands and gravel and as a thin veneer of sand mantling drift sediments. A raised gravel/sand spit occurs along the western slopes of Valley of the Flies and numerous beach ridges straddle the valley slopes and floors throughout the region. In southern Valley of the Flies glaciomarine sediments were observed underlying the marine cover unit. The distribution of raised marine deposits is discussed more extensively in Chapter 5.

Lacustrine sediments deposited in former ice- or moraine-dammed lakes were recorded as terrace remnants and isolated beach ridges. No section exposing these sediments was observed in the study area.

3.3.5. Alluvium

Inactive alluvium occurs in southern Tinutyarvik valley and in the central part of Naksaluk valley. The formation of an alluvial fan that occupies upper Tinutyarvik valley and extends east into Adams Lake valley is attributed to the emptying of a proglacial lake dammed by Quartzite moraine. Numerous abandoned channels dissect the fan surface and expose the coarse nature of the outwash gravels.

Outwash deposits located in Naksaluk valley occur below Naksaluk moraine and may have been deposited through discharge from marginal channels above the moraine. Meltwater from cirques, which are located above the valley to the west, may also have contributed coarse material through erosion and slopewash of the drift sheet which once mantled the upper valley slopes.

Active alluvium is confined to the seasonally flooded valley bottoms in the study area.

3.3.6. Till and moraine

This unit is confined to the valley bottoms and lower slopes where it is delimited by moraines or distinct trimlines. Till modification by marine and lacustrine processes has occurred in areas inundated by the sea and in areas previously occupied by glacial lakes. On the distal (northern) side of the terminal moraine in Valley of the Flies, the till surface has been modified by periglacial activity, displaying fossil sorted polygons up to 3 m in diameter (Fig. 3-7).

Figure 3-7: Photograph showing large fossil sorted polygons in northern Valley of the Flies.

The following sections describe till characteristics observed in the study area and the results of lithological and geochemical analysis of sediment units.

3.4. Sediment description and stratigraphy

Surficial sediment units exposed in stratigraphic sections are described from field observations. Interpretation of the depositional history of sediments is deduced from this information and tentative correlations between units from different geographic locations are suggested based on similarities in texture, sedimentology and stratigraphic position.

3.4.1. Tinutyarvik drift

Tinutyarvik drift sheet is situated in Tinutyarvik valley and Adams Lake valley and is bounded by Quartzite and Tinutyarvik moraines. Sediments comprising this drift are predominantly till. River-eroded sections in the central part of Tinutyarvik valley revealed a maximum of 4 m of fissile, clay-rich till with well rounded clasts (Fig. 3-8). These characteristics are typical of all till samples collected from the drift sheet. At some locations a thin mantle (30-50 cm thick) of sand was observed overlying the till unit. The sand is typically a yellowish brown, fine to medium grained, bedded unit with occurrences of silt and coarse sand lenses. This unit may represent beach sands deposited in a shallow, nearshore, marine environment.

Figure 3-8: Photograph of till section in Tinutyarvik valley displaying a fissile structure with well rounded clasts.

3.4.2. Naksaluk drift

Naksaluk drift sheet comprises the sediments bounded by Naksaluk moraine in Naksaluk valley, and primarily consists of till and outwash deposits. Till exposures revealed variable thicknesses of a clayey-silt unit with rounded to sub-rounded clasts. No sections were evident in the outwash deposits. South of Naksaluk Cove, at approximately 4 m aht, till is overlain by a 25 cm thick unit of sand which displays similar characteristics to the sand unit overlying Tinutyarvik drift, and is interpreted to be marine in origin.

3.4.3. Adams Lake drift

This drift sheet occupies Adams Lake valley east of Tinutyarvik moraine, up to the level of Adams Lake moraine. Till samples consisted of a sandy deposit with sub-angular clasts. Two river-eroded sections were recorded in sediments which overlie the till in this drift sheet and are illustrated in Figure 3-9.

Site AA is situated at an elevation of 22.5m ±0.8 m aht, below a raised shoreline segment on the southern slopes of Adams Lake valley. The section revealed $\geq$ 0.5 m of grey clayey sediment containing whole and fragmented shells [unit AA3], overlying 30 cm of sand [unit AA2], overlying a light grey muddy unit with occasional shells [unit AA1] (Fig. 3-10). This stratigraphy may represent a marine transgressive sequence with fossiliferous, marine clays overlying beach sands. A barnacle fragment (*Balanus balanus*) was identified from unit AA3 and *Mya truncata* and *Chlamys* shells were collected from surface sediments near to the section.

Further north in Adams Lake valley at 8-10 m aht two sections were excavated along a braided river channel (Site AB). Both sections consisted of sand [unit AB2] overlying a bluish-grey muddy unit [unit AB1]. The base of the sections were defined by the upper surface of a greyish-brown (10 YR 3/2) diamict. Unit AB1 contained a diverse assemblage of shells consisting of gastropod, scallop and barnacle fragments, and several whole *Gemma gemma*, *Aporrhais occidentalis*, *Mya truncata* and *Hiatella arctica* shells (Fig. 3-11). This unit is interpreted to be a nearshore marine deposit overlain by beach sands.

This site requires further excavation as some fossil shells were observed on the surface of the diamict but it is uncertain whether or not they were derived from this unit. Sampling of the diamict at this site was prevented due to the frozen condition of the sediment.

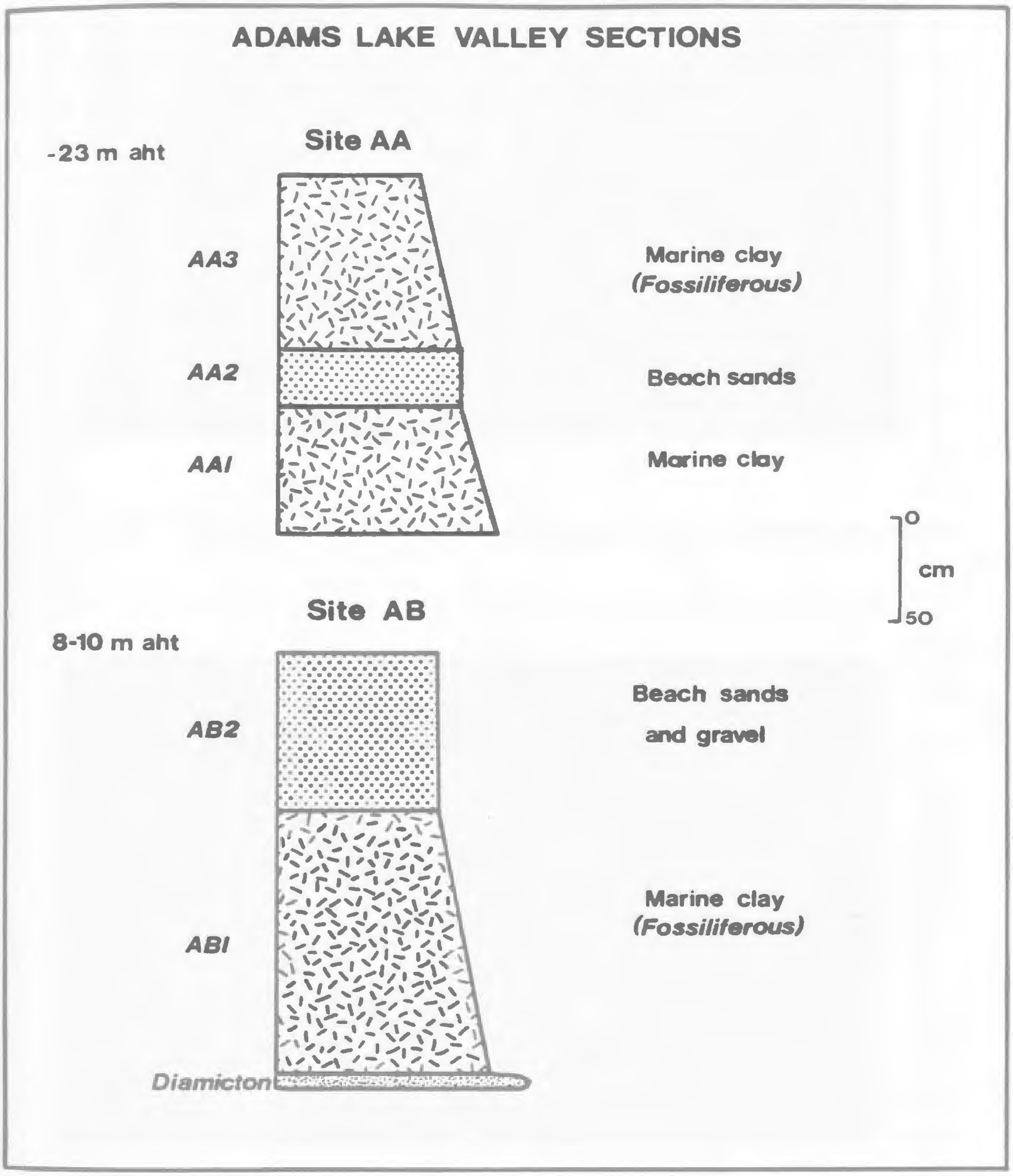

Figure 3-9: Stratigraphic section and description of units exposed at sites AA and AB in Adams Lake valley.

Figure 3-10: Photograph of stratigraphic section at site AA in Adams Lake valley.

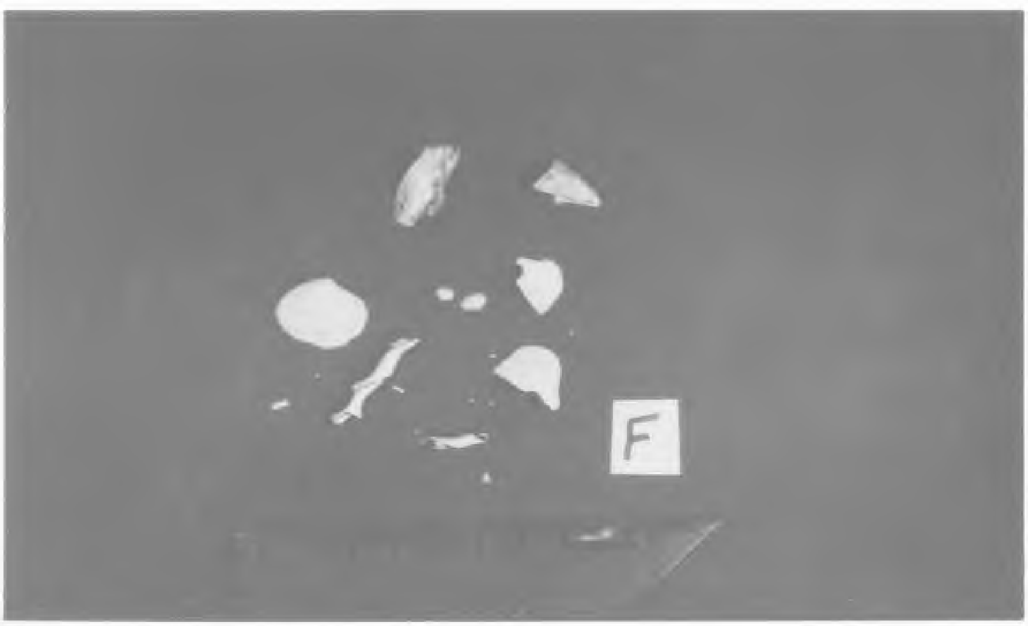

Figure 3-11: Photograph of shell sample from unit AB1 in Adams lake valley.

3.4.4. Valley of the Flies drift

This drift sheet occurs in southern Valley of the Flies where it is bounded to the north by Valley of the Flies terminal moraine. Sections through this drift are exposed along the southern shore of the valley as shoreline cliffs. Extensive slumping of material prevents a complete section being exposed, however excavations were conducted in order to delineate the extent of individual units and to determine the nature of the boundaries between units. Figure 3-12 provides an interpreted stratigraphic section through the deposits in this area.

Site VA consists of three units. At the base of the section is a dark brown (10 YR 3/3), clay-rich diamict with frequent large boulders and cobbles [unit VA1] (Fig. 3-13). This unit contained large amounts of fossil shells including *Hiatella arctica, Mya truncata, Hemithyris psittacea* and barnacle fragments (Fig. 3-14). The size of the largest *Hiatella arctica* valves were 4.9 cm X 2 cm and 3 cm X 1.6 cm X 2.4 cm in height. They were ≤ 6 mm thick at the hinge and/or thickened end. Most valves retained fairly good external ornamentation and a little internal lustre. No periostracum was evident.

Overlying the diamict was a 40 cm thick dark yellowish brown (10 YR 4/6), clast supported, shelly gravel unit [unit VA2]. This unit contained massive robust shells consisting of *Hiatella arctica, Mya truncata, Hemithyris psittacea,* a *Balanus* species and unidentified pelecypod and limpet fragments. The shells had similar characteristics to those from the underlying diamict

The uppermost unit in this section consisted of a >60 cm thick coarse clastic unit, fining upwards to a dark (5 Y 2.5/2) clayey matrix with sub-angular to sub-rounded clasts. Some shell fragments were observed along the diffuse boundary with unit VA2.

The section at site VB revealed >2 m of non-fossiliferous greyish brown (10 YR 3/2), clay-rich diamict with sub-rounded boulders and cobbles.

Site VC is located at the thickest section of the wave-cut cliff, approximately 375 m west of site VB. At the base of the section, 1.5 m of massive, medium grained sand [unit VC1] was exposed below a diamictic unit [unit VC2]. The dark olive grey (5 Y 3/2) sand is generally well sorted with occasional occurrences of granules and small pebbles. A weakly sharp to diffuse boundary separates the sand from the overlying diamicton which reaches 8.5 m in thickness, and comprises a clay-rich matrix with sub-rounded clasts. It is generally dark grey (10 YR 4/1 - variable) in colour but also contains light brown lenses of fine to medium sand.

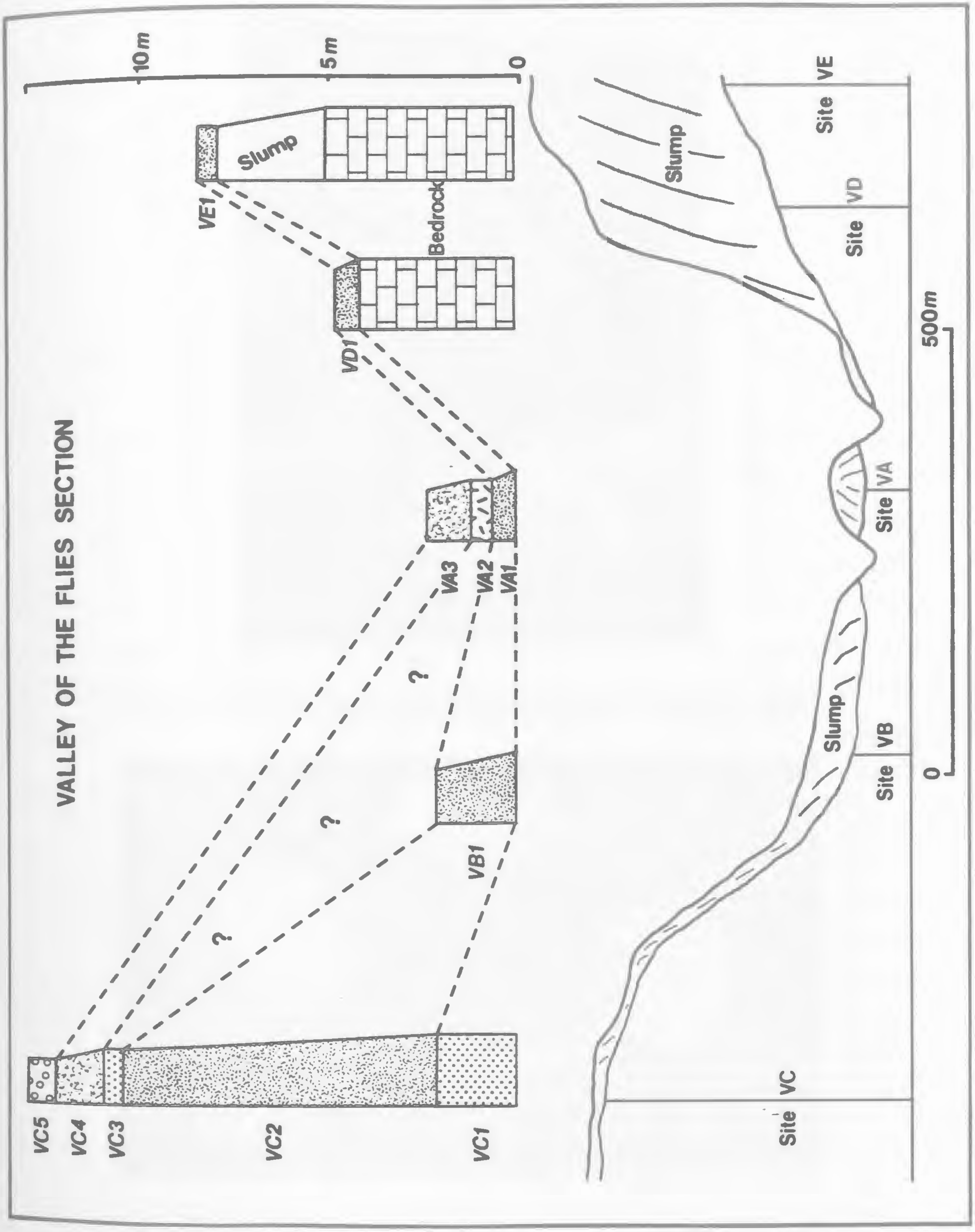

Figure 3-12: Stratigraphic section and description of units exposed along the cliffs in southern Valley of the Flies. See text for discussion of units.

Figure 3-13: Photograph of site VA in southern Valley of the Flies.

Figure 3-14: Photograph of shell sample from unit VA1 in southern Valley of the Flies.

Unit VC3 is a dark greyish brown to dark yellowish brown (10 YR 4/2 to 10 YR 4/6) fossiliferous sand. A fining upwards sequence from coarse to fine grained sand was observed in this unit. The fossil shell fauna included fragments of *Hiatella arctica* and *Mya truncata.*

Unit VC4 shows a sharp and distinct boundary with the underlying sand unit. It consists of a clast supported matrix with abundant cobbles and boulders fining up to a shell supported matrix with coarse sand and pebbles. It is approximately 1.3 m thick and varies in colour up-unit from dark greyish brown to dark yellowish brown. *Mya truncata* shell fragments were identified as one species from the fossil shell fauna in this unit.

Unit VC5 is the uppermost unit in the section, consisting of >2 m of sand with sub-rounded to rounded clasts.

Site VD revealed >80 cm of fossiliferous diamict [unit VD1] overlying bedrock. The base of the diamict is approximately 4 m aht. Most of the fossil shells were found near the diamict/bedrock boundary, although small fragments were observed above this level. The shell sample collected from unit VD1 consisted of large robust *Mya truncata* and *Hiatella arctica* fragments up to 12 mm thick in the hinge area. No periostracum, pitting nor incrustations were observed, but some internal lustre was preserved. Although the highly fragmented nature of the shell sample indicates transport, exterior ornamentation has not been removed in most cases (W. Blake Jr., personal communication, 1984).

Site VE is located 85 m east of site VD and reveals a fossiliferous, dark greyish brown (10 YR 3/2) diamict [unit VE1] lying on bedrock at 5 m aht. This unit also contains robust shell fragments of *Mya truncata* and *Hiatella arctica.*

Units within the section from Valley of the Flies can be correlated on the basis of texture and stratigraphic position (Fig. 3-12). Units VA1, VB1, VC2, VD1 and VE1 are all diamictons with similar textural and colour characteristics, and therefore may have been deposited from the same source. Clark (1984) tentatively proposed a glaciomarine origin for a diamicton exposed along the cliff section at Iron Strand, 50 km north of the study area. His diamicton had similar characteristics and depositional setting as the one described from southern Valley of the Flies, and therefore it may be tentatively concluded that the latter also has a glaciomarine origin. Unit VC1 which underlies the diamicton is interpreted to be marine sand based on its similarity with an identical unit underlying Clark's diamicton. He compared this unit with sediment sampled from the Labrador Shelf.

Unit VA2 overlies the diamicton in the lowest part of the section and is interpreted to be reworked and winnowed diamictic sediment. This conclusion is reached on the basis of the similar faunal assemblages in both units. At site VC the diamicton is overlain by a sand unit which is considered to be marine in origin. The fossil shells collected from this unit are thin and fragile in contrast to the thick, robust shells from the diamicton and are therefore not considered to be reworked from it. The fining upwards sequence within the sand unit may represent either deposition into shallowing water depths or a decrease in sediment influx or energy regime.

Units VA3 and VC4 are tentatively interpreted to be subaqueous debris flow units which exhibit fining upwards sequences of coarse clastic deposits and increasing clay content up-unit. This proposed debris flow may have occurred in association with the recession of glacial ice, or it may have been derived from the surrounding slopes during emergence of the site. These hypotheses could be tested by comparing the amino acid ratios from shells in the diamicton with those sampled from the flow unit. The sand unit that overlies the coarse clastic unit at site VC is interpreted to be a beach deposit or nearshore marine deposit based on its texture and its similarity to sand units recorded in Tinutyarvik, Naksaluk and Adams Lake valleys.

Figure 3-12 illustrates an apparent depositional hiatus which exists at site VA. The marine sand which overlies the diamicton at site VC was not observed at site VA, but instead a shell and gravel lag deposit was recorded [unit VA2]. This site is located at the bottom of a deep gully which is cut into the section, and therefore it is suggested that the original stratigraphy has been eroded and that the marine sand and upper part of the diamicton have been removed. Unit VA2, the reworked diamict unit, is only found in section along the base of the gully and consequently may represent the residue from this largely erosive event. The subsequent debris flow mantled the entire section and therefore postdates this event.

3.4.5. MI and MII drift

MI drift occurs above MII moraine in Naksaluk valley and above *ca.* 120 m aht in Adams Lake valley. The latter boundary is not clearly defined and is extrapolated from the elevation of MII moraine in Naksaluk valley. The upper level of MI drift is marked by the MI trimline in both valleys. It is tentatively suggested that MI drift may also occur in Valley of the Flies up to the level of the lateral moraines, and distal to Valley of the Flies terminal moraine (see Section 3.2.5).

Samples of morainic material from this area consisted of sub-rounded to sub-angular clasts in a sandy matrix.

MII drift is located in an area distal to Naksaluk moraine and Adams Lake moraine and bounded by MII moraine in Naksaluk valley. It also occurs distal to Adams Lake moraine in Adams Lake valley but its upper limit is not delineated by a clear trimline. No sections were observed in this drift sheet but samples taken from the bottom of soil pits revealed a sandy till with sub-angular clasts.

3.4.6. Kammarsuit valley drift

This drift sheet occurs along the floor and lower slopes of the Kammarsuit valley system. It is bounded by local lateral and end moraines in each of the valleys. A river-eroded section through the end moraine in Kammarsuit valley east revealed a sandy deposit with sub-angular clasts. A similar deposit was exposed at the base of soil pits dug on moraine crests.

3.4.7. Discussion

Several broad conclusions can be made on the basis of the above sediments and stratigraphies. First, it appears evident that till units in Tinutyarvik and Naksaluk drift sheets have a high clay content in contrast to the sandy texture of tills from elsewhere in the study area. Second, it is tentatively suggested that the glaciomarine diamicton observed in Valley of the Flies section may extend west into Adams Lake valley where a diamict with similar characteristics was observed on the floor of the valley. Third, the presence of marine sand below the diamicton in Valley of the Flies, and the presence of marine fossils within the diamicton suggest that the ice which was responsible for its deposition advanced into a marine environment. The marine sand overlying the diamicton may represent an emergent sequence. Last, it is noted that beach sand and gravel overlie many of the stratigraphic sections; however, the range in elevations of these units above high tide suggests that they were deposited at different periods during postglacial relative sea level change.

The similarity between the stratigraphies from Valley of the Flies section and the wave-cut cliff section at Iron Strand (Clark, 1984) suggests that there may be a correlation between events in these two areas. Both sections reveal glaciomarine diamictons stratigraphically positioned between two marine sand units. Chapter 6 will examine correlations between these stratigraphies based on radiocarbon dates and on amino acid ratios from fossil shells collected from the individual units.

3.5. Drift pebble lithology

During the summers of 1984 and 1985, 59 samples containing pebbles from various drift sheets and moraines were collected from the study area for lithological examination. The purpose of this investigation was to discern ice flow patterns and ice limits in the valleys of the peninsula which, through their orientation, could have acted as distributary outlets for regional fiord ice, but may also have served as outlets for local valley glaciers during, or prior to the last glaciation.

The local geology provides an ideal opportunity to examine regional and local ice sheet movements through analysis of drift pebble lithology. The geological map of the area illustrates three broad zones based on major lithological characteristics of the bedrock (Fig. 3-15).

Zone I is composed predominantly of gneisses of Archaean age, termed Nachvak gneiss by Wardle (1983). Nachvak gneiss is typically a banded quartzofeldspathic gneiss of tonalitic to granodioritic composition, and includes relics of early supracrustal sequences. During the Hudsonian Orogeny these gneisses were overprinted by slight to moderate retrograde metamorphism. In most cases this feature renders the Nain Province gneisses readily distinguishable from rocks of the Churchill Province.

Zone II comprises quartzites, pelites, sandstones and minor limestones and dolomites of the Ramah Group.

Nachvak gneisses from the Churchill Province (Zone III) are in granulite facies throughout. They are compositionally similar to those of the Nain Province, but have experienced additional deformation and metamorphism during the Hudsonian Orogeny. Within Zone III the intensity of deformation increases westwards. Deformation in the eastern part (Churchill Border Zone), is restricted to zonal reworking in shear zones with moderate development of mylonitic fabrics, whereas rocks in the western part (Churchill Inner Zone) have been pervasively straightened and mylonitized. The strongly deformed Tasiuyak gneiss is also included in Zone III.

The distribution of glacial erratics from Zone III depicts the former presence of regional fiord ice in the study area, whereas the distribution of drift characterized by Zones I and II lithologies delimits the area of local glacier activity.

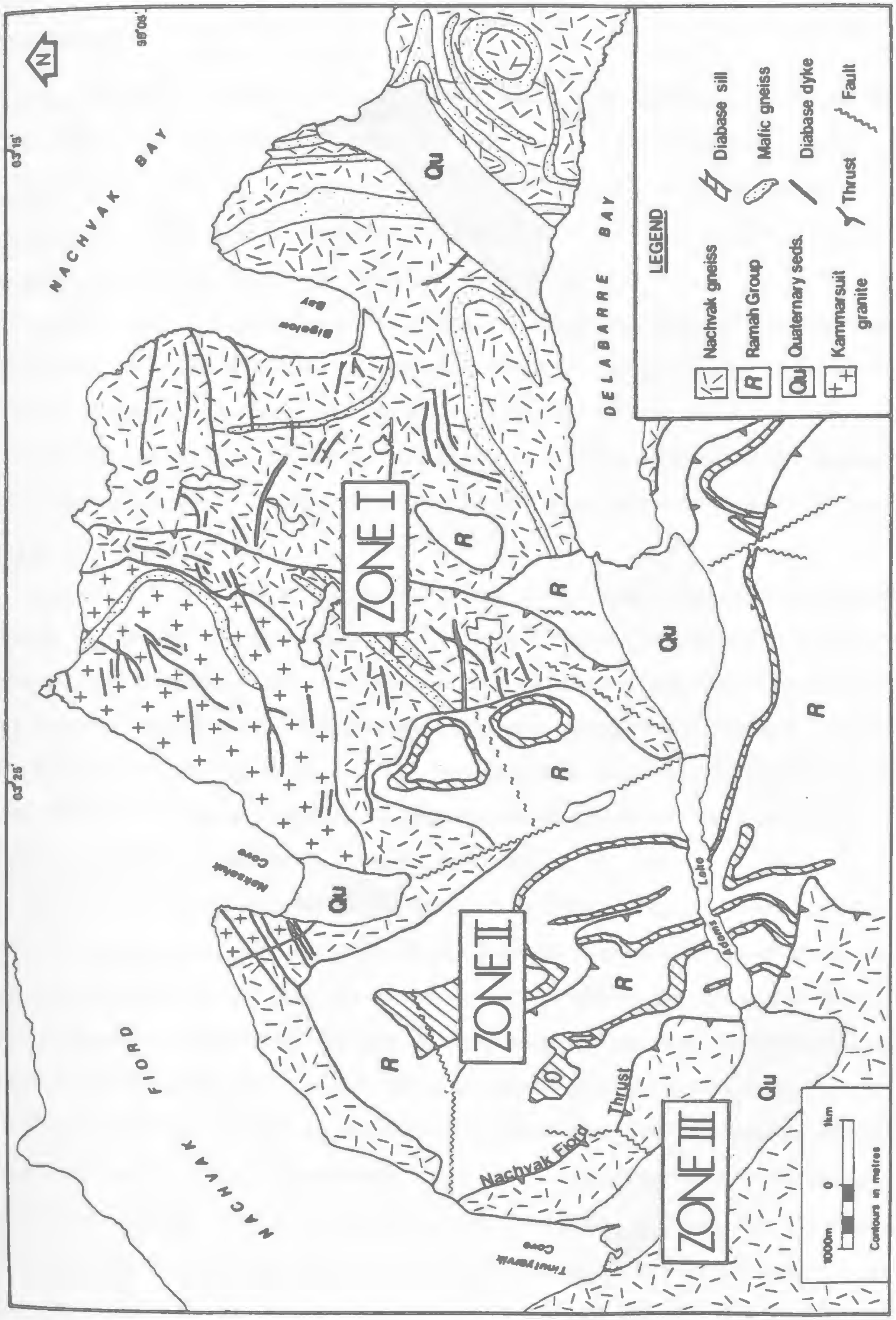

Figure 3-15: The three broad lithological zones in the study area, after Morgan (1975) and Wardle (1983).

3.5.1. Methods

Each of the 59 samples collected in the field consisted of approximately 200 pebbles and small cobbles, between 6 mm and 80 mm, gathered from exposed till sections and from the bottom of soil pits dug on moraine crests and in drift sheets (Fig. 3-16). Sites with obvious signs of solifluction or local scree accumulation were avoided. Consequently, many of the older soliflucted and inaccessible moraine segments were not sampled.

Samples sites L1-L15 are located in Tinutyarvik valley and along the shore of Tinutyarvik Cove [Tinutyarvik drift]. Samples L11-L14 were collected from the bottom of soil pits dug along the crests of Quartzite moraine, whereas samples L10 and L15 were taken from exposed till sections along the flanks of the moraine. Sample site L16 is situated east of the drainage divide between Adams Lake and Tinutyarvik valleys, but lies within the area bounded by Tinutyarvik moraine [Tinutyarvik drift].

Samples L18 and L20-L23 were collected from soil pits dug on the crest of Naksaluk moraine. Sample L17 was gathered from an exposed till section approximately 4 m aht in Naksaluk Cove [Naksaluk drift]. Sample sites L28-L35 are located in drift associated with Adams Lake moraine in Adams Lake valley [Adams Lake drift]. Samples L28, L29 and L32-L35 were collected from soil pits dug on the moraine crest. Seven samples were obtained from soil pits dug in MII drift, above Naksaluk and Adams Lake moraines; samples L19 and L24-L27 in Naksaluk valley and samples L36 and L37 in Adams Lake valley.

Sample sites L40-L56 are located in Valley of the Flies. Samples L40 and L41 were collected from soil pits dug on the crest of the high lateral moraine in the northern part of the valley [MI drift]. Samples L42 and L43 were obtained from soil pits dug on the crest of Valley of the Flies terminal moraine. Samples L44-L56 were collected from exposed drift sections along the southern edge of the valley, on the proximal side of the moraine [Valley of the Flies drift].

Sample sites L38 and L39 are located on the crests of the lateral and terminal moraines, respectively, at the mouth of Kammarsuit valley east [Kammarsuit valley drift]. Samples L57 and L58 were collected from sorted polygons at an elevation of approximately 750 m aht on the mountains to the west of Naksaluk valley (Fig. 3-17). Sample L59 was collected from a till section associated with the Nachvak glacial phase (Evans and Rogerson, 1986), in Ivitak Cove, 10 km west of Tinutyarvik Cove (Fig. 1-1).

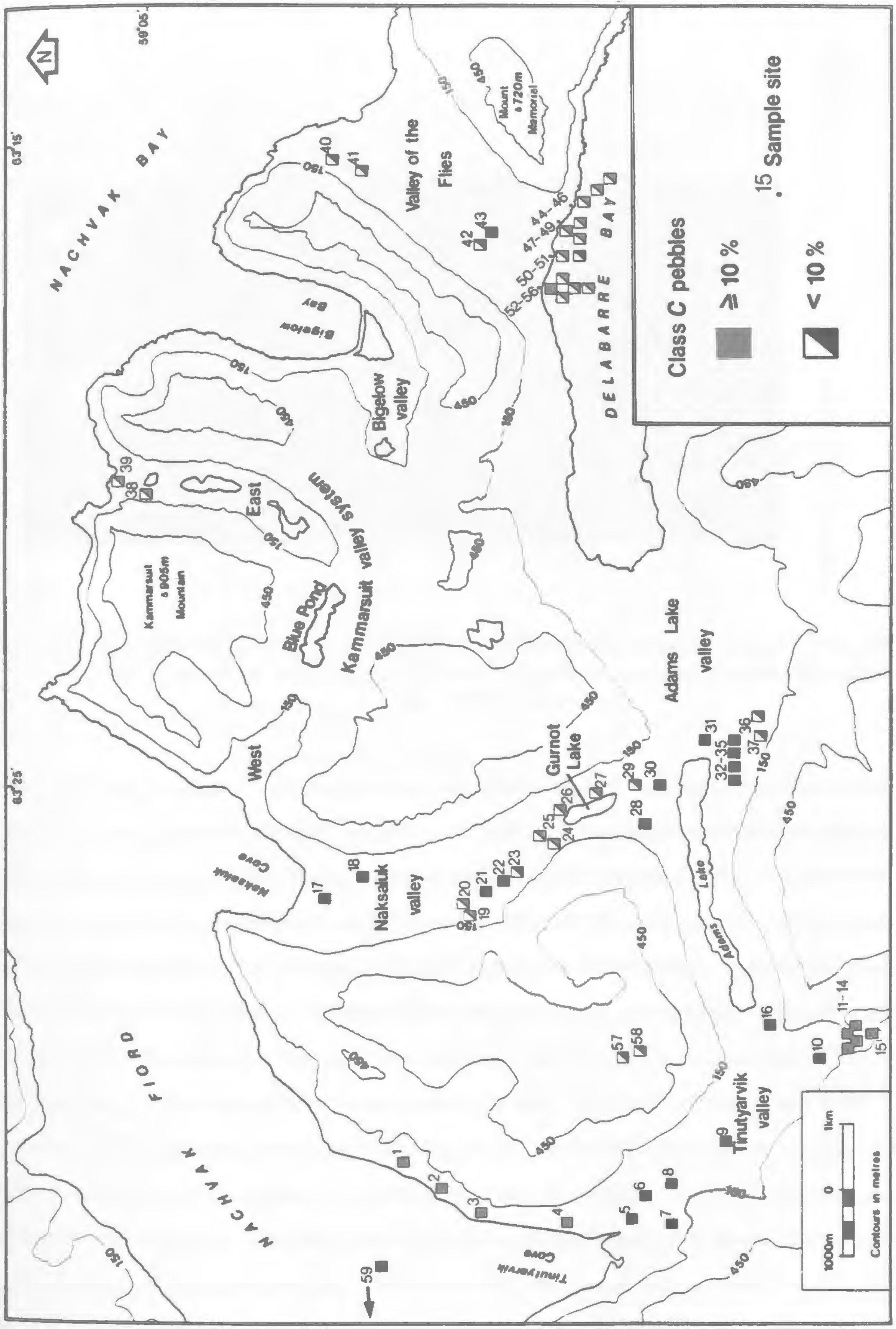

Figure 3-16: Location of drift pebble sample sites in the study area and a plot of percentage Class C pebbles in each sample.

Figure 3-17: Photograph of sample sites L57 and L58 located at approximately 750 m above sea level, north of Adams Lake valley. Samples were collected from the centre of the sorted polygons.

In the laboratory, samples were washed in a 5.6 mm sieve and each pebble/cobble classified according to five classes, A to E, defined in Table 3-1. Class A comprises pebbles that display evidence of retrogressed, amphibolite-grade mineral assemblages, typical of Nachvak gneiss in Zone I. Class B consists of pebbles from the sedimentary Ramah Group (Zone II), and includes quartzites, pelites, sandstones and dolomite. Class C comprises lithologies from Zone III which are typically highly-deformed and sheared gneisses with granulite grade mineralogy. The presence of homogeneous ductile shearing and/or garnet are characteristic of pebbles in this class. Fine to medium-grained mafic rocks are present in both the Nain and Churchill Provinces and were assigned to Class D. Similarities with Zone I lithology increase towards the eastern margin of Zone III, resulting in gneisses with amphibolite to granulite grade mineralogy, variable deformation and some evidence of retrogression. Therefore the majority of local Tinutyarvik gneiss was assigned to Class E as having an ambiguous origin. Likewise locally deformed and sheared Zone I gneiss may suggest a Class C membership, although characteristic mineral assemblages normally provided positive identification.

Table 3-1: Lithological classification and description for drift pebble samples. Refer to Figure 3-15 for location of lithology zones in study area.

Pebble Class descriptions and origins		
Class	Description	Origin
A	Gneisses with retrogressed amphibolite facies mineralogy	Zone I, area east of Naksaluk valley
B	Quartzite, pelites sandstone, dolomite	Zone II, sedimentary Ramah Group
C	Gneisses with granulite grade mineralogy, homogeneous ductile shearing, presence of garnet	Zone III, area west of Tinutyarvik valley
D	Fine to medium grained mafic rocks	All zones
E	Gneisses with amphibolite to granulite grade mineralogy variable deformation, some evidence of retrogression	Zone I & eastern Zone III, Tinutyarvik valley and east of Naksaluk valley

The study area is predominantly underlain by Zone I and II lithologies and therefore the presence[4] of Class C pebbles from Zone III indicates that tills originated west of Tinutyarvik Cove from an easterly flow of ice along Nachvak Fiord. This ice advance is referred to as *regional* ice and is interpreted to represent outlet glaciers from the Laurentide ice sheet which streamed through the major valleys of the Torngat Mountains from the interior of Labrador.

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[4]The generalized grouping of lithological types, coupled with the inherent error due to inappropriate classification does not justify specific statements based on <10% of the gneissic component of pebble samples.

3.5.2. Results

Analysis of drift pebble lithology records the distribution and abundance of each pebble class, and provides a basis for describing the extent and flow direction of ice sources responsible for deposition of glacial sediments in the study area. Results of the analysis are given in Appendix A.

Figure 3-16 is a plot of the percentage of Class C pebbles in each sample according to site location in the study area. The distribution of samples having $\geq 10\%$ Class C pebbles depicts the former presence of regional ice. This area is bounded by Valley of the Flies terminal moraine in southern Valley of the Flies, Adams Lake moraine in Adams Lake valley, Naksaluk moraine in Naksaluk valley, and Tinutyarvik and Quartzite moraines in Tinutyarvik valley [Valley of the Flies, Adams Lake, Naksaluk and Tinutyarvik drift sheets]. Sample L59 in Ivitak Cove also occurs within this distribution.

The pebble samples were also grouped according to their lithological composition by means of cluster analysis. This approach uses a hierarchial clustering procedure based upon a measure of similarity between the composition of individual samples[5]. This similarity measure was calculated using the Euclidean distance - the distance between two samples being the square root of the sum of the squared differences in values on each lithological class.

$$d_{ij} = \sqrt{\sum_{k=1}^{m} (X_{ik} - X_{jk})^2}$$

where d_{ij} is the distance between sample i and sample j, X_{ik} denotes the k^{th} class measured on sample i, X_{jk} is the k^{th} class measured on sample j, and m is the number of classes measured on each sample (Davis, 1973). Initially, measures of similarities between, or distances separating, individual samples were obtained. Then the two most similar or nearest samples were combined to form a cluster. This new cluster inherited the averaged properties of the samples that composed it. The similarities or distances between this new cluster and all other samples were recomputed. As before, the two most similar or nearest samples/clusters were combined to form a new cluster. This process was repeated until all samples were combined in one cluster.

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[5]Class E pebbles could not be positively identified according to the criteria designated for the lithological zones in the study area and therefore this class was excluded from the clustering procedure.

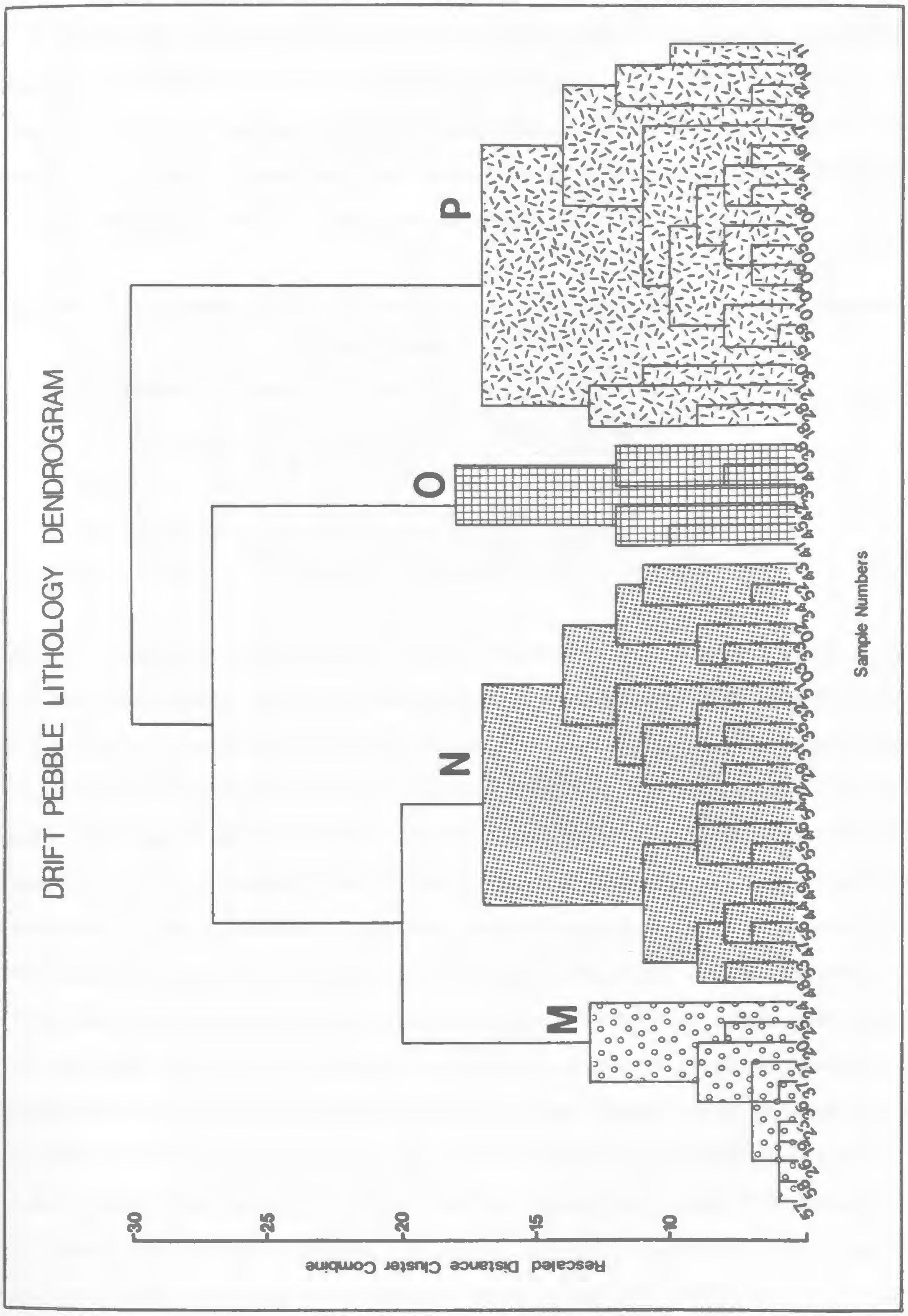

Figure 3-18: Dendrogram of cluster samples, scaled using the agglomeration measures of the clusters.

Figure 3-18 is a dendrogram of the cluster solution which is scaled using the agglomeration measures of the clusters. Although the samples that comprise each of the clusters were grouped objectively using the Euclidean distance formula, the separation of these four clusters, as representing significantly dissimilar groups of samples, was carried out through visual inspection of the dendrogram. Table 3-2 gives the mean percentages of pebble classes in each of the Clusters *M, N, O* and *P*, and Figure 3-19 shows the general pattern of cluster distribution.

Table 3-2: Mean values (%) for lithological composition by pebble class in each cluster.

Mean Cluster Values (%)					
Cluster	Class A	Class B	Class C	Class D	Samples
M	00.0	94.1	01.6	03.4	11
N	19.4	53.8	11.5	08.6	22
O	59.5	13.8	02.4	23.0	6
P	00.0	07.2	64.8	16.4	20

Cluster *M* comprises mainly samples from sites located beyond the limit of regional ice in lithological Zone II [MII drift]. These samples are composed predominantly of Class B pebbles from Zone II. Cluster *O* comprises mainly samples from sites located beyond the limit of regional ice in lithological Zone I [Kammarsuit valley drift and MI drift in Valley of the Flies]. Samples consist mainly of Classes A and B pebbles from Zones I and II respectively, and Class D pebbles.

Cluster *P* samples are located within an area bounded by Naksaluk moraine in Naksaluk valley, Tinutyarvik moraine in Adams Lake valley, and Tinutyarvik and Quartzite moraines in Tinutyarvik valley [Naksaluk and Tinutyarvik drift sheets]. Geographical anomalies within Cluster *P* distribution are samples L28 and L30 which were collected from Adams Lake moraine in northern Adams Lake valley. The lithological compositions of these two samples cannot be used to characterise this area according to this classification scheme. Sample site L59, located in Ivitak Cove, is included in Cluster *P* membership. The general distribution of Cluster *P* samples is similar to that recorded for regional ice, except that the eastern limit, south of the fiord, for Cluster *P* is Tinutyarvik moraine in Adams Lake valley, whereas Valley of the Flies terminal moraine lies at the limit of regional ice according to the presence of $\geq$ 10% Class C pebbles. Cluster *P* samples are also characterized by a high percentage of Class C pebbles (65%).

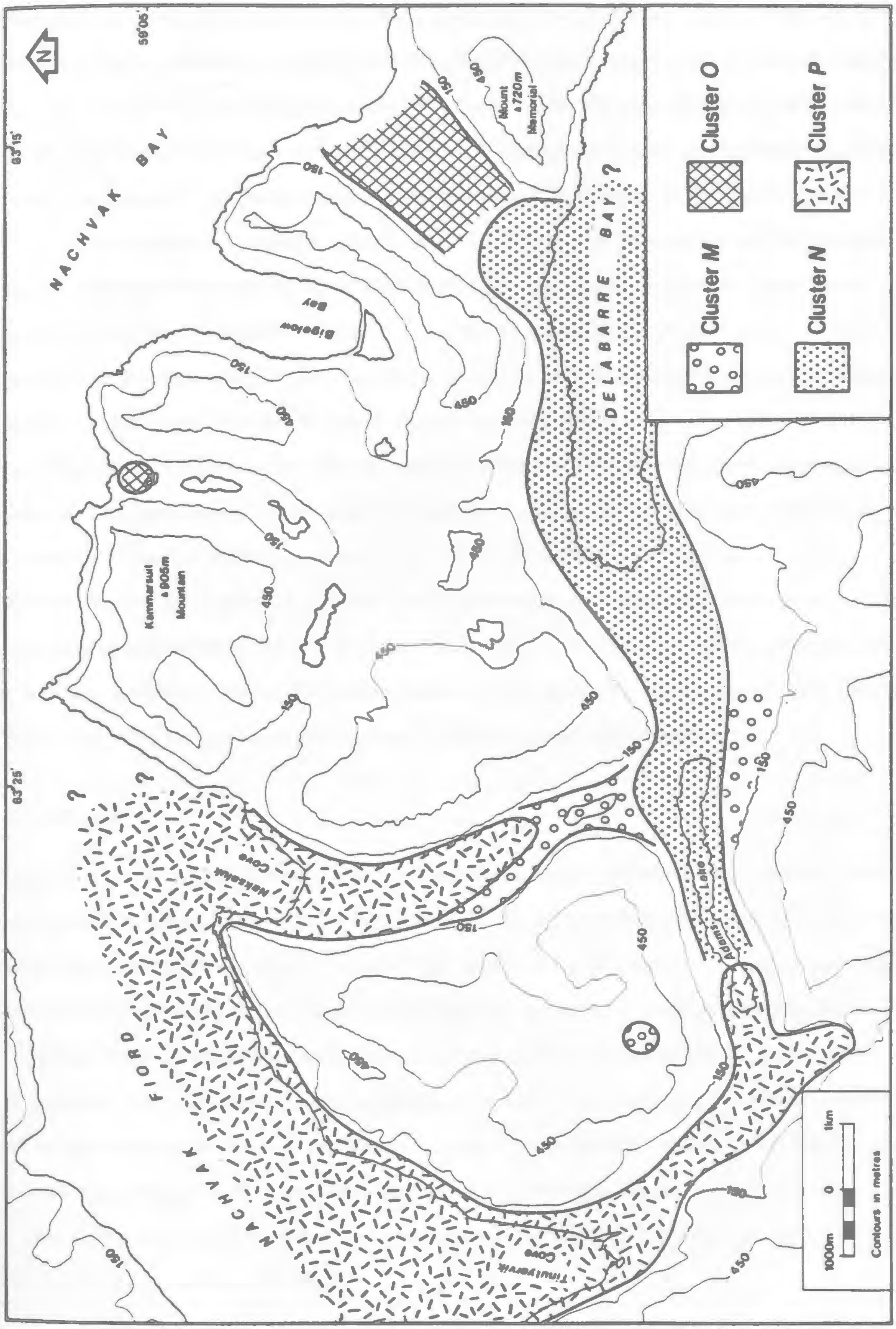

Figure 3-19: Geographical distribution of Clusters M, N, O and P.

Cluster N samples are bounded by Adams Lake moraine in Adams Lake valley and Valley of the Flies terminal moraine in Valley of the Flies [Adams Lake and Valley of the Flies drift]. Anomalies to Cluster N distribution are samples L2 and L3 in Tinutyarvik valley, and sample L22 on Naksaluk moraine in Naksaluk valley. Samples L2 and L3 have high percentages of Class B pebbles which is typical of Cluster N samples, but atypical of samples from Tinutyarvik valley. However these anomalous percentages can be explained by the location of L2 and L3 sample sites on the thrusted margin of Zone II (Fig. 3-15) where talus spreads of shattered Class B rock presently occur, and probably existed prior to the last advance of ice into the area. Sample L22 also contains high percentages of Class B pebbles and may reflect transport of Zone II lithologies by the western part of the ice lobe that backfilled Naksaluk valley. Samples L20 and L23 (>90% Class B pebbles) from Cluster M are also located on Naksaluk moraine and lend support to this hypothesis. In contrast, sample L18 which is located on a segment of Naksaluk moraine on the eastern slopes of Naksaluk valley has only 11.4% Class B pebbles.

The distribution of Cluster N samples generally occupies the area between the former limit of regional ice and the eastern limit of Cluster P samples in Adams Lake valley. Cluster N samples, in comparison to Cluster P samples, have a much lower percentage of Class C pebbles (12%), but a much higher percentage of Class B pebbles (54%; Table 3-2).

3.5.3. Discussion

Analysis of drift pebble lithology reveals definite ice flow patterns and ice limits in the study area. The former presence of regional ice is delimited by the distribution of glacial erratics (Class C pebbles) from the Churchill Inner Province (Fig. 3-16). This distribution suggests that regional ice advanced along the fiord, backfilling Naksaluk valley as far as Naksaluk moraine (80 m aht) and terminating west of where Kammarsuit valley east opens out to the fiord. Drift pebble lithology suggests that only local glacier activity occurred in the Kammarsuit valley system as indicated by the absence of Class C pebbles in samples from this drift sheet. The distribution of glacial erratics also suggests that ice advanced southeast through Tinutyarvik valley, east along Adams Lake valley and Delabarre Bay, and terminated in Valley of the Flies at the moraine (21.5 m aht) in the southern part of the valley.

Similarities in the lithological composition of drift pebble samples, as illustrated by the cluster analysis results, suggest that the Tinutyarvik moraine in Adams lake valley is the limit of Cluster P regional drift, south of the fiord. The apparent difference in the lithological composition of samples in southeastern Tinutyarvik valley (Cluster P) and Adams Lake valley (Cluster N) may be indicative of the distance over which regional ice travelled in lithological Zone II. Drift pebble lithology alone cannot differentiate between the drift sheets either side of Tinutyarvik moraine as there are no sample sites immediately distal to the moraine which might indicate an abrupt change in concentration of Class C glacial erratics. However, these two proposed regional ice limits will be compared with other data from the study area in order to resolve the most probable limit.

The lithological composition of Cluster M samples which characterize MII drift in Naksaluk and Adams Lake valleys suggests that these sites were last inundated by local ice (94.1% Class B pebbles). Therefore MII moraines do not represent a readvance phase of the more extensive MI regional ice advance, as tentatively proposed in the discussion of glacial landforms (see Section 3.2.5). Instead, it is more likely that the limit of MII drift marks the maximum extent of local glacier activity in Adams Lake valley (*Adams Lake glacial phase*).

If this assumption is valid, then it is reasonable to suggest that Adams Lake drift may also have been deposited by local ice. The lower concentrations of Class C pebbles in MII drift, as compared to Adams Lake drift, may be explained by the higher elevations of MII sample sites above the floor of Adams Lake valley. The most likely origin of Class C pebbles in drift sheets with predominantly local characteristics is from sediments deposited by former regional ice sheets. Reworking of this sediment during subsequent local glacier activity would be more extensive in the lower parts of the valley. Further speculation might suggest that a local ice advance of this type terminated at the moraine in southern Valley of the Flies, and that Adams Lake moraine represents a recessional stage from this maximum.

3.6. Drift geochemistry

In 1983 Evans initiated geochemical sampling for the Nachvak Fiord region (Evans, 1984). The main aim of this project was to gather information on the economic geology of the area through geochemical analysis of the local drift sheets. A second objective, but one which Evans was unable to complete due to restricted data, was to use the geochemical characteristics of the drift sheets to determine provenance, and hence former ice flow and dispersal.

The second stage of the sampling program was carried out in 1984 and 1985 along the outer region of Nachvak Fiord. Samples were collected from exposed drift sections and from the unweathered horizon of soil pits dug in drift. The Geological Survey of Canada prepared the samples and forwarded them to the laboratories of Bondar-Clegg & Co. Ltd., Ottawa, for analysis. Atomic absorption methods were used to detect the following metals - chromium, manganese, iron, cobalt, nickel, copper, zinc, molybdenum, silver and lead, whereas a fluorimetric technique was employed for uranium detection.

During analysis of the 1985 samples, the laboratory chose to process the silt and clay size fraction ($<63\mu$) [due to the coarse nature of the samples], in contrast to the clay size fraction ($<2\mu$) which was the basis for analysis in the two previous years. Consequently the three data sets cannot be directly compared, although relative numbers will be valid (R.A. Klassen, personal communication, 1986). As the 1985 data include 48 samples and provide a complete coverage of the study area, this set was used to provide a general picture of the distribution of metal content in the various drift sheets and to compare the results with the distribution of drift pebble lithologies (see Section 3.5). Evans' results and the 1984 samples are used for comparison of the geochemical characteristics of the two study areas, and to evaluate the data as potential indicators of ice flow direction and dispersal. The 1985 results are used specifically in an attempt to distinguish local and regional drift sheets on the outer coast.

The Geological Survey of Canada has been actively involved in till studies and their application to drift prospecting in many parts of the Canadian Shield, the Cordillera, and the largely unexplored Canadian Arctic. These studies usually take the form of large scale mapping projects (e.g. Shilts, 1973, 1974; Shilts and Klassen, 1976; Dyke, 1983, 1984), which are designed primarily to analyse the geochemical properties of regional drift sheets for mineral exploration; and secondly to determine former ice flow and dispersal (e.g. Klassen, 1983, 1984). The Nachvak

Fiord project differs from these studies in that it considers a fiord landscape at a very small scale. Unlike such areas as the central arctic which has experienced large-scale lowland glaciation, the Torngat Mountains have hosted both regional Laurentide ice and local highland ice, and therefore possess a complex suite of both local and regional drift. Results of geochemical analysis on these drift sheets may reflect this long history of ice interaction through poorly defined local and regional geochemical characteristics and patterns.

3.8.1. Metal concentrations, backgrounds and thresholds

The results of the geochemical analysis on samples from the study area are presented in Appendix B. Sample numbers correspond to the equivalent site numbers from the drift pebble sampling program (Fig. 3-16); unique geochemical sample sites are labelled with an alphabetic code. Sample GAA was collected from the base of a soil pit located between sample sites G32 and G33 on Adams Lake moraine. Samples GBB and GCC were taken from soil pits dug in MI drift in Valley of the Flies and sample GDD was collected from unit VC2 in southern Valley of the Flies cliff section.

Figure 3-20 displays the 1985 sample results as frequency histograms for each metal. Mean background and threshold values were estimated using a method outlined by Hawkes and Webb (1962). They suggested that, with small data sets asymmetrically distributed, the best approximation of 'background' was the median value, and that the 'threshold' should be that value which is exceeded by no more than 2.5% of the total number of observations, excluding markedly high erratic values. Table 3-3 summarizes these values for each metal, listing those samples which exceeded the threshold value. Threshold values used by Dyke (1983) for samples from Somerset Island, District of Franklin, are included for comparison.

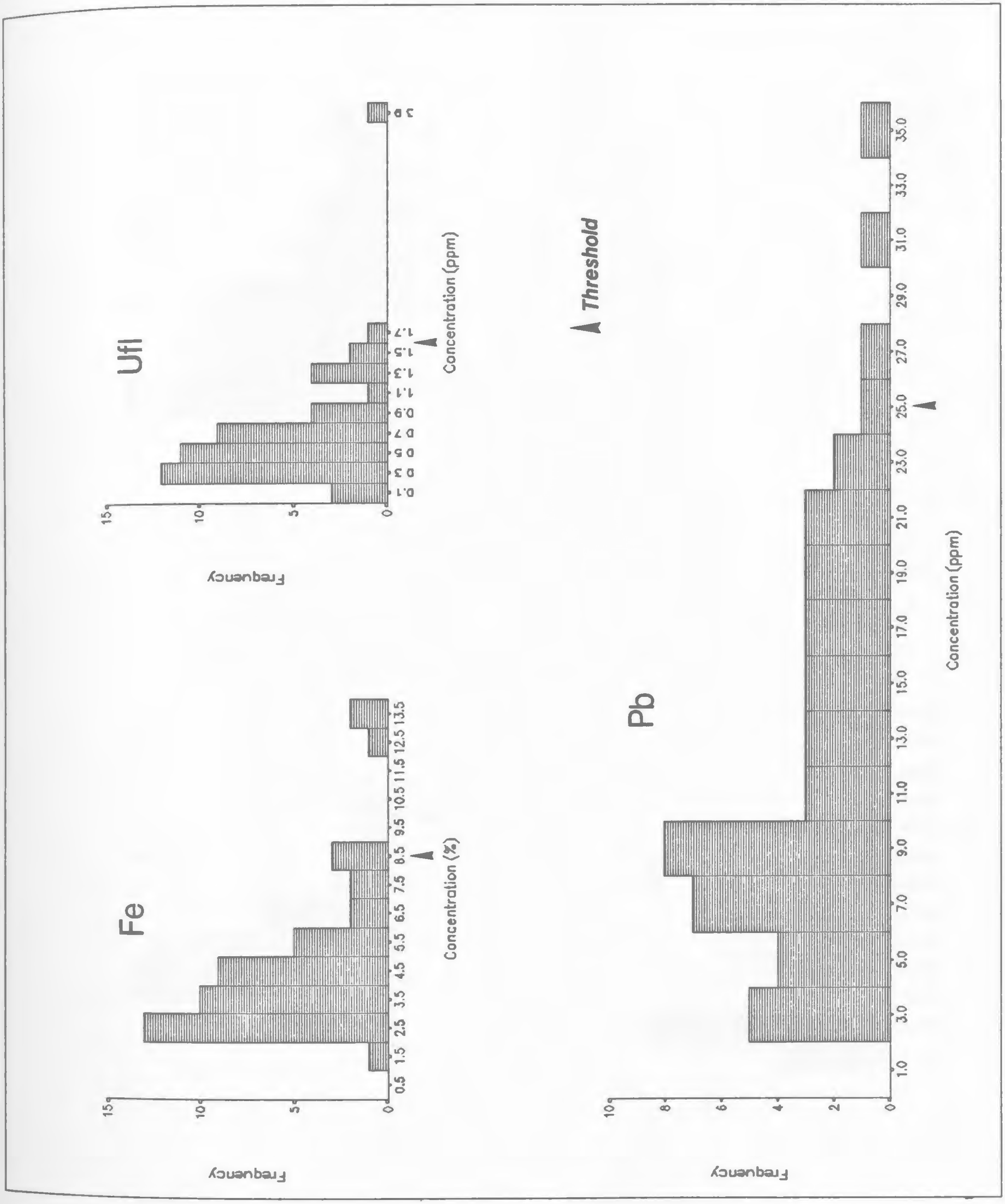

Figure 3-20: Frequency histograms of metal concentration for 1985 samples

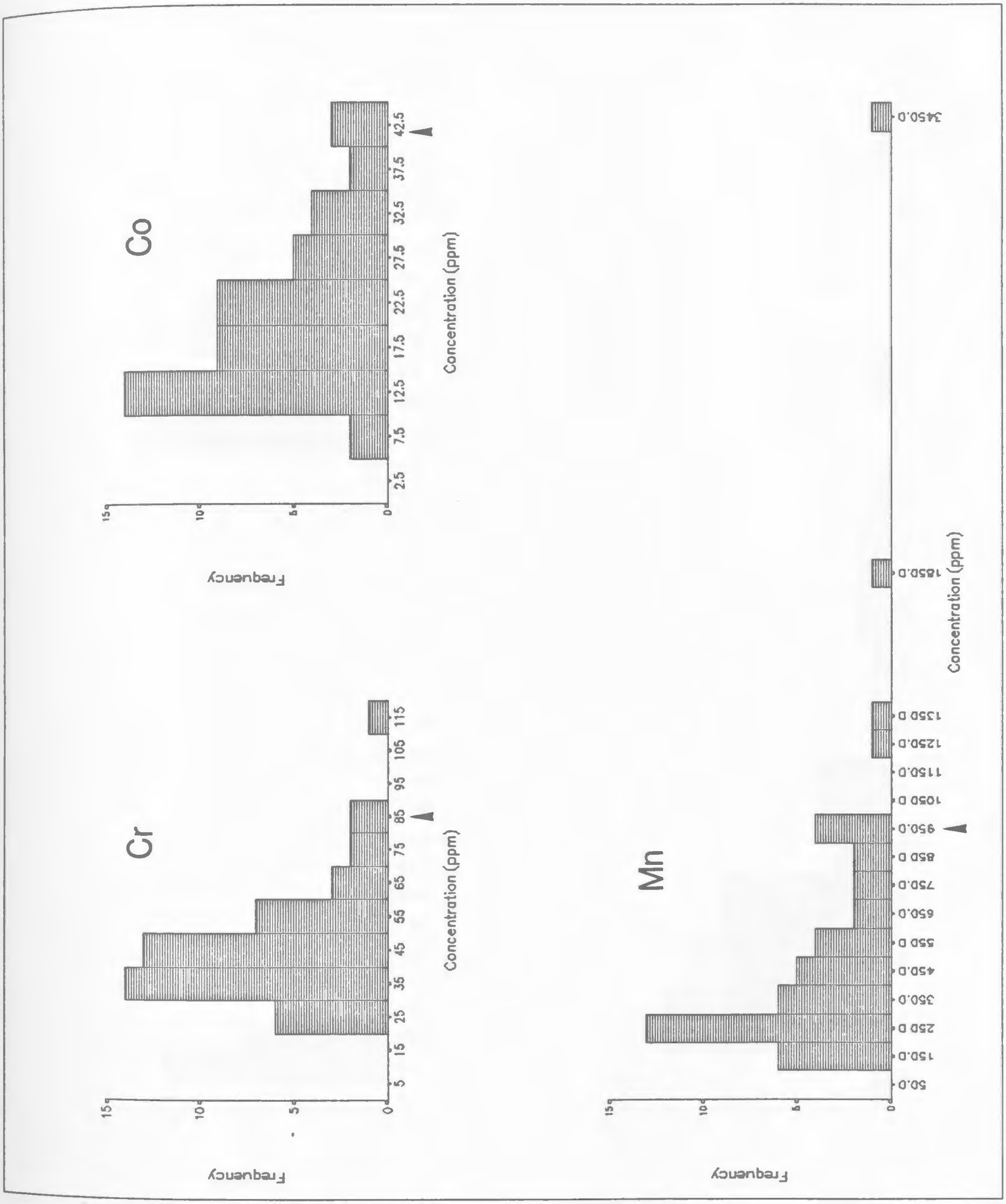

Figure 3-20 [Cont.]: Frequency histograms of metal concentration for 1985 samples

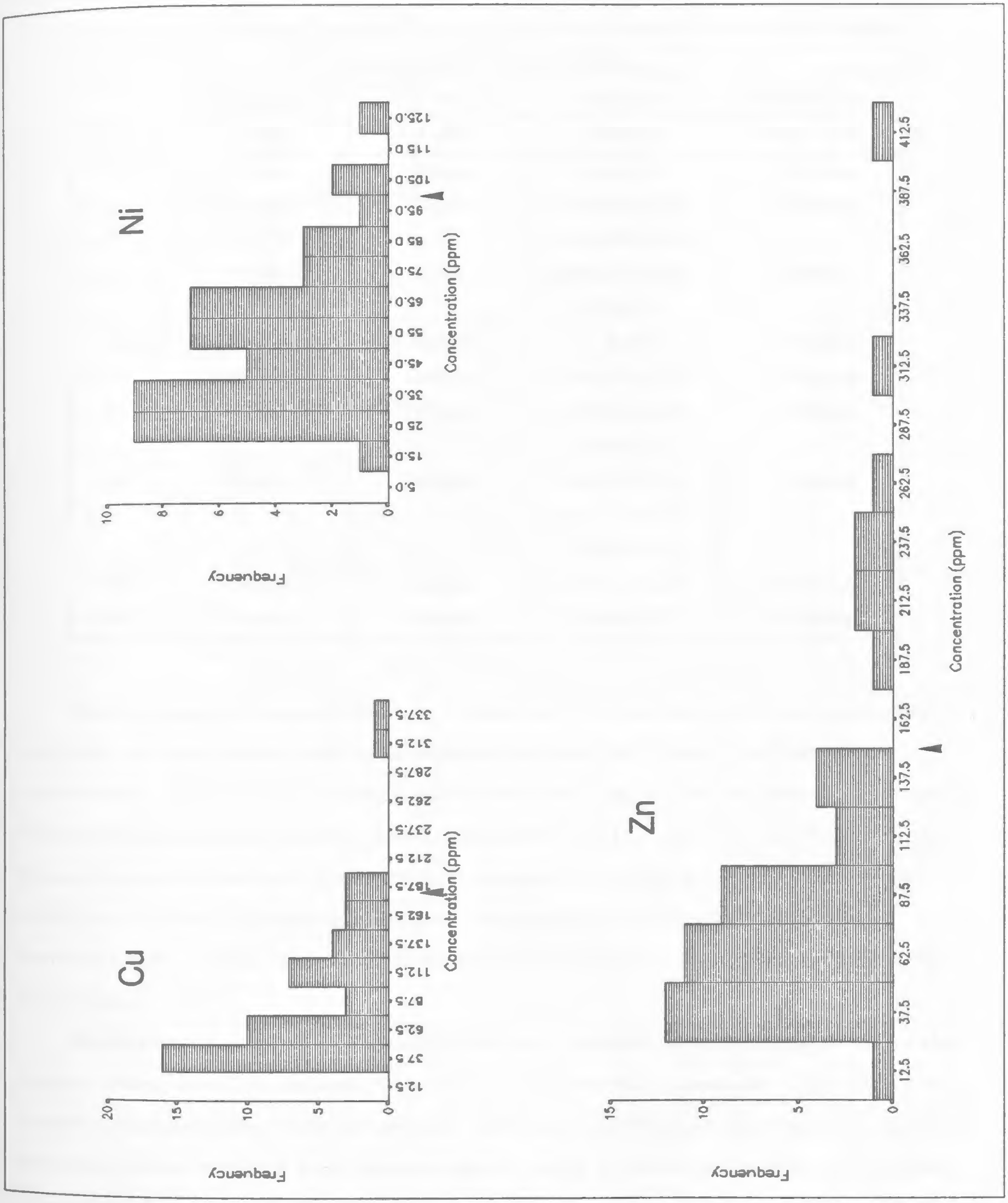

Figure 3-20 [Cont.]: Frequency histograms of metal concentration for 1985 samples

Table 3-3: Background and threshold values for metal concentration in 1985 samples.

Background and threshold values				
	Background	Threshold	Anomalous	Threshold value
Metal	Value	Value	Samples	Dyke (1983)
Cr	35ppm	85ppm	GAA,G43	>100ppm
Mn	225ppm	950ppm	G20,G25,G29	>1000ppm
			G31,G32,G34	
Fe	2.5%	8.5%	G26,GAA,G33	>10%
			G34,G37	
Co	12ppm	42ppm	G34	>100ppm
Ni	30ppm	100ppm	GAA,G34,G43	>100ppm
Cu	37ppm	187ppm	G26,GAA,G33	>100ppm
			G34,G35	
Zn	37ppm	150ppm	G26,G29,G31	>150ppm
			G32,GAA,G33	
			G34,G35	
Pb	7.5ppm	25ppm	G26,GAA,G37	>100ppm
Ufl	0.3ppm	1.6ppm	GAA,G37	0-1.6ppm

The calculation of threshold values for a particular set of samples is obviously site specific, and values will vary between study areas depending on local drift lithology and bedrock characteristics. Furthermore, the sample size fraction used in an analysis will directly affect the values for background and threshold concentration (Shilts, 1971). In spite of this, threshold values for metal content in this study [$<63\mu$ fraction] compare closely with those employed by Dyke (1983) [$<2\mu$ fraction]. Chromium, manganese, iron, nickel, zinc and uranium had similar threshold values, whereas cobalt was 50% lower, lead was 75% lower and copper was 85% higher in this study.

Samples that exceeded these threshold values have a marked distribution along Adams Lake moraine, sample sites G29, GAA and G31-G35. Isolated sites with anomalously high results were located in Naksaluk valley (G26), Adams Lake valley (G37) and Valley of the Flies (G43). All of these samples have relatively high concentrations of Class B pebbles from the sedimentary Ramah Group (Zone II on Fig. 3-15) (See Appendix A), but this alone does not explain the high results, as other samples with normal 'background' metal concentrations had similar or higher percentages of Class B pebbles. It is more probable that local bedrock properties are responsible for these anomalies. Local conditions are emphasized because regional drift samples show no evidence of these high metal contents. It is suggested that the above sample sites should be target areas for further mineral exploration.

The geochemical composition of drift is a function of its lithology, which in turn, depends greatly upon the composition of the rocks and sediments that were overridden by the glacier. As drift pebble lithology demonstrated in the previous section, regional fiord ice and local ice can be differentiated based on the lithological composition of their respective drift sheets. Therefore, it is assumed that the geochemical composition of the various drift sheets in the study area will also discern local and regional ice origins. This hypothesis was tested using both visual and statistical means.

3.6.2. Statistical analysis of metal concentrations

The first step in the analysis was to assign each geochemical sample to its respective drift pebble lithology cluster, *M*, *N*, *O* or *P*, according to its lithological composition (Table B-1, Appendix B). Through visual inspection of the data it was decided to divide Cluster *N* into two smaller clusters, *N1* and *N2*. Cluster *N1* consists of sample sites located in Zone I [Adams Lake drift] and Cluster *N2* consists of samples collected from Zone II [Valley of the Flies drift]. Comparison of the mean and standard deviation values of metal concentration in both clusters justifies this decision (Tables 3-4 & 3-5); Cluster *N2* has, in most cases, more than twice the metal concentration of Cluster *N1* samples. Drift lithology is presumed to be responsible for the difference between these two sub-populations. Cluster *N2* has a higher percentage of Class B pebbles (Zone II) and mean values of 1% Class A pebbles (Zone I) and 21% Class C pebbles (Zone III) compared to 32% and 5% respectively, in Cluster *N1*.

On the basis of the assumption that there is a close relationship between drift lithology and drift geochemistry, discriminant analysis was carried out on the geochemistry results in order to identify and describe the clustered geochemical samples and to discern those metals which were most effective in distinguishing the clusters. Furthermore the validity of the above assumption was tested through re-evaluation of cluster membership for each sample in a cluster, using parameters defined for cluster identification.

Table 3-4: Mean concentrations of metals for 1985 and 1984 geochemistry samples.

	Cr	Mn	Fe(%)	Co	Ni	Cu	Zn	Pb	Ufl
Mean concentrations <ppm> - 1985 samples									
ALL DATA	46	548	4.7	21	53	93	101	12	0.7
Cluster M	37	661	7.0	23	62	106	127	18	1.3
Cluster N1	44	378	3.7	19	41	63	77	11	0.3
Cluster N2	63	1136	7.1	32	79	168	199	19	1.0
Cluster O	40	352	3.6	17	36	87	64	10	0.5
Cluster P	41	267	2.7	14	43	57	47	5	0.3
Mean concentrations <ppm> - 1984 samples									
ALL DATA	98	792	5.9	40	144	201	175	22	1.2
Cluster P	102	477	5.2	34	155	199	139	12	1.1
Cluster Z	93	1106	6.5	46	133	204	210	32	1.2

Table 3-5: Standard deviation of metal concentration values for 1985 and 1984 geochemistry samples.

	Cr	Mn	Fe(%)	Co	Ni	Cu	Zn	Pb	Ufl
Standard deviations <ppm> - 1985 samples									
ALL DATA	19	548	2.8	9	25	68	81	8	0.6
Cluster N1	27	150	1.1	7	25	36	32	5	0.2
Cluster N2	19	906	2.9	9	27	96	109	8	0.4
Cluster O	7	101	0.8	6	19	55	20	3	0.1
Cluster P	7	221	0.8	6	13	27	25	3	0.1
Standard deviations <ppm> - 1984 samples									
ALL DATA	17	547	1.2	17	68	65	92	16	0.6
Cluster P	21	65	0.8	9	84	66	31	8	0.9
Cluster Z	11	650	1.3	23	55	72	122	15	0.2

Discriminant analysis attempts to statistically distinguish between two or more groups of samples. In this analysis the groups or clusters are defined by samples with similar lithological properties, as determined by cluster analysis (See Section 3-5). Discriminating variables that measure characteristics on which the groups are expected to differ were selected for the analysis.

These variables were chosen from the geochemical properties of the drift samples. Molybdenum and silver were not considered because of their low or undetectable values in many of the samples (Table B-2, Appendix B). Discriminant analysis ranks and combines the discriminating variables (discriminant functions) in a linear array so that the groups are as statistically distinct as possible or, in other words, the groups can be specifically described according to their geochemical properties. These discriminant functions are of the form

$$D_i = d_{i1}Z_1 + d_{i2}Z_2 + \dots + d_{ip}Z_p$$

where D_i is the score on the discriminant function i, the d's are weighting coefficients, and the Z's are the standardized values of the p discriminating variables used in the analysis. The maximum number of functions which can be derived is either one less than the number of groups or equal to the number of discriminating variables, if there are more groups than variables. The discriminant scores (D's) for samples within a particular cluster should be similar, whereas scores for samples in separate clusters should be different.

Discriminating variables comprising the functions were chosen by a stepwise selection procedure. This procedure begins by selecting the single best-discriminating variable that minimizes R, the residual variation[6]. A second discriminating variable is selected as the variable best able to improve the value of R in combination with the first variable. The third and subsequent variables are similarly selected according to their ability to contribute to further discrimination. Eventually, either all variables will have been selected or it will be found that the remaining variables are no longer able to contribute to further discrimination. When this point has been reached, the stepwise procedure halts and the discriminant scores are calculated using only the selected variables.

The SPSS^x subprogram *Discriminant* (*SPSS* Inc., 1983) provides several statistical tests for measuring the success with which the variables actually discriminate when combined into the discriminant functions. The eigenvalue is a measure of the relative importance of the discriminant function. When a single eigenvalue is expressed as a percentage of the total sum of eigenvalues, the resultant figure is an indication of the relative discriminating power of the

[6]The discriminating variable that minimizes the sum of unexplained variation between groups is selected. It is defined by the value $R = \sum \frac{1}{1 + (D_{ij}/4)}$ where D_{ij} is the Mahalonobis distance between groups i and j.

associated function. A second criterion for assessing discriminant functions is to test for the statistical significance of discriminating information not already accounted for by the functions. As each function is derived, starting with zero functions, Wilks' lambda is computed. Lambda is an inverse measure of the discrimination power in the original variables which has not yet been removed by the discriminating functions - the larger lambda is, the less information remains. Lambda can be transformed into a chi-square value for a test of statistical significance. The level of probability relating to a particular value of lambda is the percentage chance that the remaining information in the variables is due to sampling and measurement errors.

The graphical display of discriminant scores allows visual examination of group centroids and appraisal of group relationships. Even though group separation may be statistically significant, intermingling of group samples can occur. By plotting discriminant scores for each sample on a graph (if two or more discriminant functions are derived) or a histogram (if only one discriminant function is produced), deviant samples can be identified, suggesting that the discriminating variables are not totally successful.

Discriminant analysis can be used as a classification technique. If the values of the discriminating variables for a sample are known, then the likely group membership can be identified. Another use of classification is in testing the adequacy of the derived discriminant functions by classifying the original set of samples to see how many are correctly classified by the variables being used. A measure of success in the discriminative process can be observed in the proportion of correct classifications. The procedure for classification involves the use of a separate linear combination of the discriminating variables for each group. These classification equations are derived from the pooled within-group covariances and the centroids for the discriminating variables. A group equation takes the form

$$C_i = c_{i1}V_1 + c_{i2}V_2 + \ . \ . \ . \ + c_{ip}V_p + c_{i0}$$

where C_i is the classification score for group i, the c_{ip}'s are the classification coefficients with c_{i0} being the constant, and the V's are the raw scores on the discriminating variables. These equations produce a score for the respective groups, and the sample is assigned to the group with the highest score; that is, the group for which it has the highest probability of membership.

3.6.3. Results

Tables 3-4 and 3-5 provides information on group/cluster means and standard deviations for each variable considered in the analysis. This information reveals substantial differences between groups for many of the variables. The value for Wilks' lambda prior to the derivation of any discriminant functions is *0.07398*, suggesting a high degree of discriminating power in the variables (Table 3-6).

After the first step of the analysis, the following variables were selected as discriminators in sequence: Pb, Cu, Ni, Cr, Fe, Co and Zn. This set of variables is an optimal set, rather than a maximal set, because not every possible subset is tested for its discriminative potential. The sequence in which the variables are selected is not necessarily the same as their relative importance as discriminators, so therefore an alternative evaluation is required. For this analysis, the *F-to-remove* statistic was judged as the most critical evaluation of the discriminating power of a variable. *F-to-remove* is a multivariate partial *F* test which assesses the contribution of individual variables to the separation of groups. If, at any stage during the analysis, a selected variable loses its significant contribution to the equation and its *F-to-remove* value falls below a given level, then the variable is removed from the equation. Following selection of all significant discriminating variables, a list of their *F-to-remove* statistics are printed (Table 3-7). The relative importance of each variable as a discriminator is given by this statistic - the higher the value, the more important the discriminator. Therefore, in order of importance, the variables are: Cr, Ni, Cu, Fe, Co, Zn and Pb. These figures show that chromium is an exceptionally strong discriminator of the five clusters, with nickel and copper contributing significantly, but to a lesser extent.

These seven variables were weighted and linearly combined to form four discriminant functions (Table 3-6). The four functions together account for 100% of the variance between the groups. Although functions *1* and *2* account for 87% of the variance, the relatively significant lambda values associated with the two remaining functions suggest that they significantly contribute to the discrimination of the groups. Consequently, the remaining computations were based on all four functions.

Further information on group discrimination can be derived from the group centroids (Table 3-8) and a plot of all samples (Fig. 3-21). The group centroids are the mean discriminant scores

Table 3-6: Significance of functions derived from the discriminant analysis of 1985 data.

Discriminant Function Statistics							
	Eigen	%	After	Wilks'	Chi-		
Function	value	variance	function	Lambda	squared	df	Signif.
			0	0.07398	106.76	28	>99.9%
1	2.94713	63.94	1	0.29202	50.47	18	>99.9%
2	1.06101	23.02	2	0.60186	20.82	10	>97.7%
3	0.47352	10.27	3	0.88685	4.92	4	>70.5%
4	0.12759	2.77					

Table 3-7: The *F-to-remove* statistic for variables contributing to the discrimination.

F-to-remove Statistic	
Variable	F-to-remove
Cr	9.1232
Ni	6.4983
Cu	5.6736
Fe	3.2438
Co	1.9104
Zn	1.5731
Pb	1.5269

Table 3-8: Group centroid values for each cluster[7] derived from the discriminant functions.

Group Centroids				
Group	Function 1	Function 2	Function 3	Function 4
CLUSTER M	+3.14207**	-1.02295**	-0.15058	+0.11016
CLUSTER N1	-0.56065	+0.20560	-0.83339**	-0.42265**
CLUSTER N2	+0.78915	+1.63288**	+0.56610**	+0.01986
CLUSTER O	-1.52630**	+0.25972	-0.84914**	+0.82446**
CLUSTER P	-1.37353**	-0.83610**	+0.63976**	-0.03950

for each group on the respective functions. Therefore it is possible to discern those functions which distinguish between certain groups. Figure 3-21 shows that the groups are not clearly

[7] ** denotes those clusters which are clearly distinguished by the respective functions.

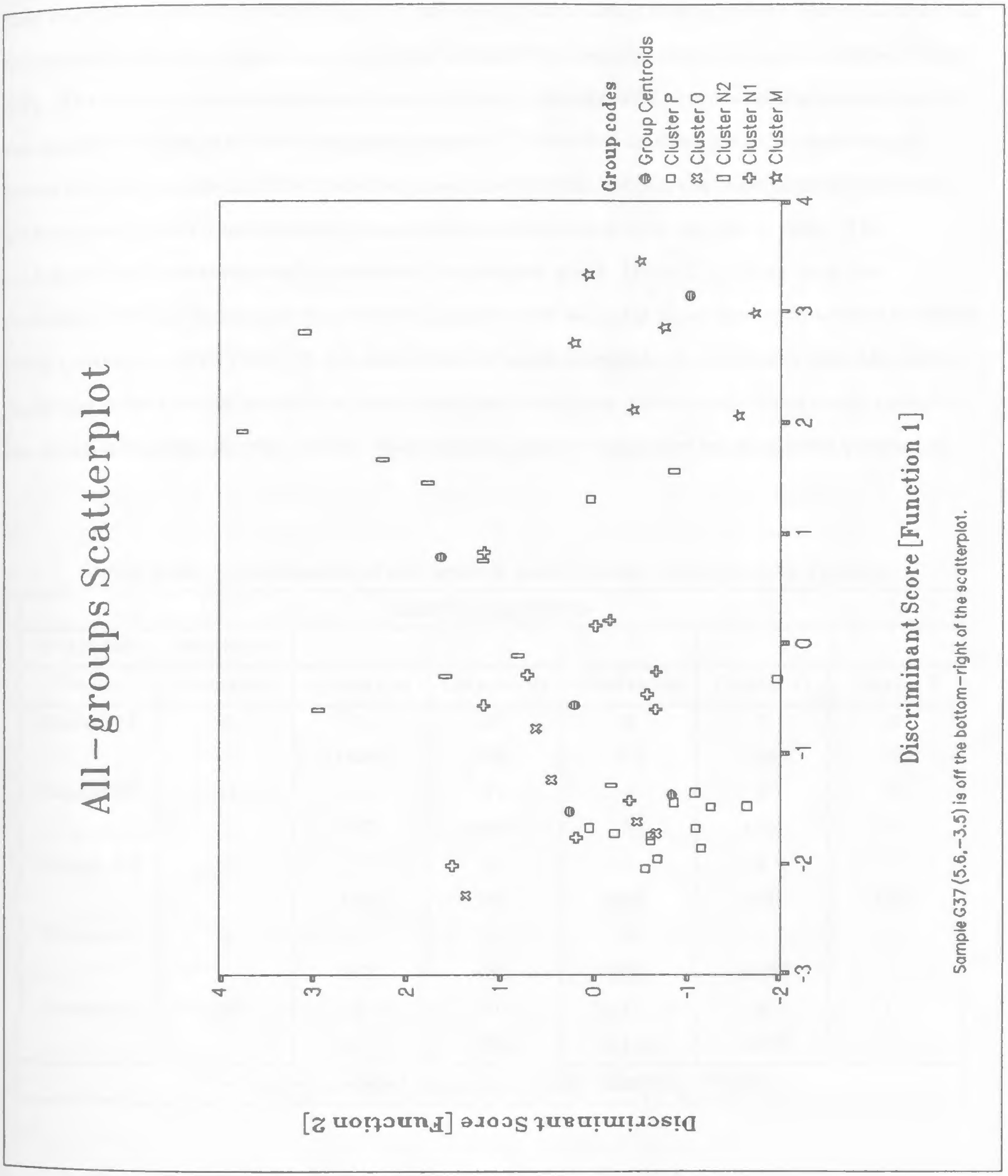

Figure 3-21: **Scatterplot of group samples and group centroids as determined by the first two discriminant functions.**

separated even though the discrimination is statistically significant (this graph is defined by the first two discriminant functions only). A measure of the overlap existing in the discrimination can be assessed when the original set of samples is classified using the discriminating variables (Table 3-9). The classification routine was able to correctly identify 90% of the samples as members of the groups to which they were originally assigned. Tables B-3 and B-4 give the discriminant scores for each sample and the predicted group membership. When the classification disagrees with the actual group membership, three asterisks are printed after the group code. The probability for membership in a predicted group is also given, $\{P(G/X)\}$, along with the probability that a member of the predicted group would be as far from the centroid as the sample being considered, $\{P(X/G)\}$. If the latter value is small, it signals the possibility that the sample might not belong to the population from which the groups are drawn, even though it is closest to the group indicated. In other words, these samples may be anomalies in the general population.

Table 3-9: Classification of geochemical samples using discriminating variables.

Classification results						
Predicted Cluster	Number of samples	Cluster M	Cluster N1	Cluster N2	Cluster O	Cluster P
Cluster M	8	8	0	0	0	0
		100%	0%	0%	0%	0%
Cluster N1	11	0	11	0	0	0
		0%	100%	0%	0%	0%
Cluster N2	10	1	0	8	0	1
		10%	0%	80%	0%	10%
Cluster O	5	0	0	0	5	0
		0%	0%	0%	100%	0%
Cluster P	14	0	0	1	2	11
		0%	0%	7.1%	14.3%	78.6%
Percent of predicted samples correctly classified: 89.58%						

In this analysis five samples were reclassified because of a predicted higher group membership in an alternative group. Three of these were from sites where lithologies were geographically anomalous (L3, L22, L28). Samples G01 and G18 were reassigned to Cluster *O* from Cluster *P*. Sample G03, initially classified in Cluster *N2* was assigned to Cluster *P*. Samples G22 and G28 were reassigned to Clusters *M* and *N2* respectively. Those samples which appear anomalous to the general population (that is, $\{P(X/G)\} < 0.25$) are GAA, G33, G34, G36, G37, G40 and G43. There is close correspondance between these predicted anomalies and those samples of anomalously high metal concentration derived by the method outlined by Hawkes and Webb (1962) (see Table 3-3).

3.6.4. Discussion

The discriminant analysis carried out on the sample geochemistry results was based on the assumption that drift lithology was directly related to drift geochemistry. The choice of drift pebble lithology clusters as a grouping mechanism for the geochemistry samples was dependent upon this assumption. The results of the discriminant analysis suggest that this relationship is strong (90% of samples correctly classified). Some of those samples misclassified (G01 and G18) were located in marginal zones of cluster territories and therefore possess similar probabilities of membership to either the revised cluster or the initial cluster. The reclassification of sample G03 from Cluster *N2* to Cluster *P* accords with a similar prediction derived from the drift pebble lithology results. Samples G22 and G28 have little probability of membership in their assigned cluster and are reclassified to new clusters with significantly higher probabilities of membership.

The all-groups scatterplot (Fig. 3-21) graphically displays the separation of the groups based on the first two discriminant functions. Cluster *M* is clearly the most homogeneous group with a clearly defined territory. Cluster *M* samples have characteristic Class B pebbles (Ramah Group lithologies) exclusively, and hence represent typical Zone II drift geochemistry. In contrast, the four remaining clusters, Clusters *N1*, *N2*, *O* and *P*, are more loosely distributed and contain overlap areas. Clusters *O* and *P* have similar distributions which may relate to the high percentages of Nachvak gneiss typical of their samples. Clusters *N1* and *N2* are subpopulations of Cluster *N* and therefore are expected to occupy adjacent territories with overlapping distributions.

The main purpose of discriminant analysis is to predict which samples have the highest probability of membership to mutually exclusive groups based on the evidence of various

Table 3-10: Comparison of geochemistry results of duplicate 1984[8] and 1985 samples.

Comparison of 1984 and 1985 geochemistry results <ppm>									
	Cr	Mn	Fe(%)	Co	Ni	Cu	Zn	Pb	Ufl
G11	40	220	2.3	14	39	43	36	4	0.2
G11+	96	440	4.6	30	136	204	114	5	0.3
G20	36	960	5.8	32	63	100	95	221	0.9
G20+	110	2200	6.2	80	180	235	144	55	1.4
G34	75	1800	8.9	43	100	187	300	24	1.3
G34+	98	1000	8.3	54	205	300	425	38	1.4

characteristics. Therefore it should be possible to classify geochemical samples from the adjacent Selamiut Range (Evans, 1984) on the basis of the geochemical properties of the various groups in outer Nachvak Fiord. Unfortunately, geochemical data collected prior to 1985 cannot be directly compared, due to different sample size-fraction analysis. Three sites, G11, G20 and G34, sampled in 1984 were revisited and resampled in 1985. A comparison of the geochemical analysis of both sets of samples (Table 3-10) reveals the range of concentrations detected due to the processing of two different size fractions. The 1984 results, based on the clay size fraction, show more than twice the metal concentrations for the same samples collected in 1985 which were processed using the silt and clay size fraction. Therefore the samples collected by Evans in 1983 cannot be used in this discriminant analysis. Instead, a second discriminant analysis was carried out using only the 1984 data and Evans' results. Samples were grouped into two clusters, Clusters *P* and *Z*. Cluster *Z* consists of all samples collected in 1984 which were not located in Cluster *P* territory, as defined by the drift pebble lithology analysis (Table B-1). Only ten samples were analysed in 1984 and therefore any further sub-division of the clusters is restricted by the number of samples. Evans' samples were classified as ungrouped in this analysis.

Results of the discriminant analysis are given in Tables 3-11 and 3-12. One discriminant function was derived [only two groups] based on the weighting and combination of the two geochemical variables, Pb and Co. These two metals were capable of distinguishing the two

[8] + denotes 1984 sample.

Table 3-11: Statistical significance of function derived from the discriminant analysis of the 1984 data.

Discriminant Function Statistics							
	Eigen	%	After	Wilks'	Chi-		
Function	value	variance	function	Lambda	squared	df	Signif.
			0	0.42669	5.9619	2	>94.9%
1	1.34364	100					

Table 3-12: The *F-to-remove* statistic for variables contributing to the discriminant analysis and the group centroids for both clusters derived from the analysis.

F-to-remove Statistic and group centroids			
Variable	F-to-remove	Group	Function 1
Pb	7.4451	Cluster P	-1.03678
Co	1.5471	Cluster Z	+1.03678

groups of samples with a high level of statistical significance. 90% of the samples were correctly classified (Table B-6). Sample G08 was reassigned to Cluster *Z* mainly due to its high lead content. No specific reason can be given for this high concentration and therefore it is presumed that this site has an anomously high lead content.

Based on the geochemical characteristics of the two sample clusters, the predicted membership for three representive[9] samples from the Selamiut Range was calculated. All three samples were assigned to Cluster *P* (Fig. 3-22), thereby suggesting a similarity in geochemical properties, and hence supporting an earlier conclusion derived from the drift pebble lithology results that Cluster *P* samples are evidence of regional ice movement from the west.

[9]These three samples represent the mean sample results for three groups of samples differentiated according to their relative ages, as defined by pedogenesis. The terms Ivitak, Nachvak and Superguksoak relate to progressively younger glacial phases (see Evans and Rogerson, 1986)

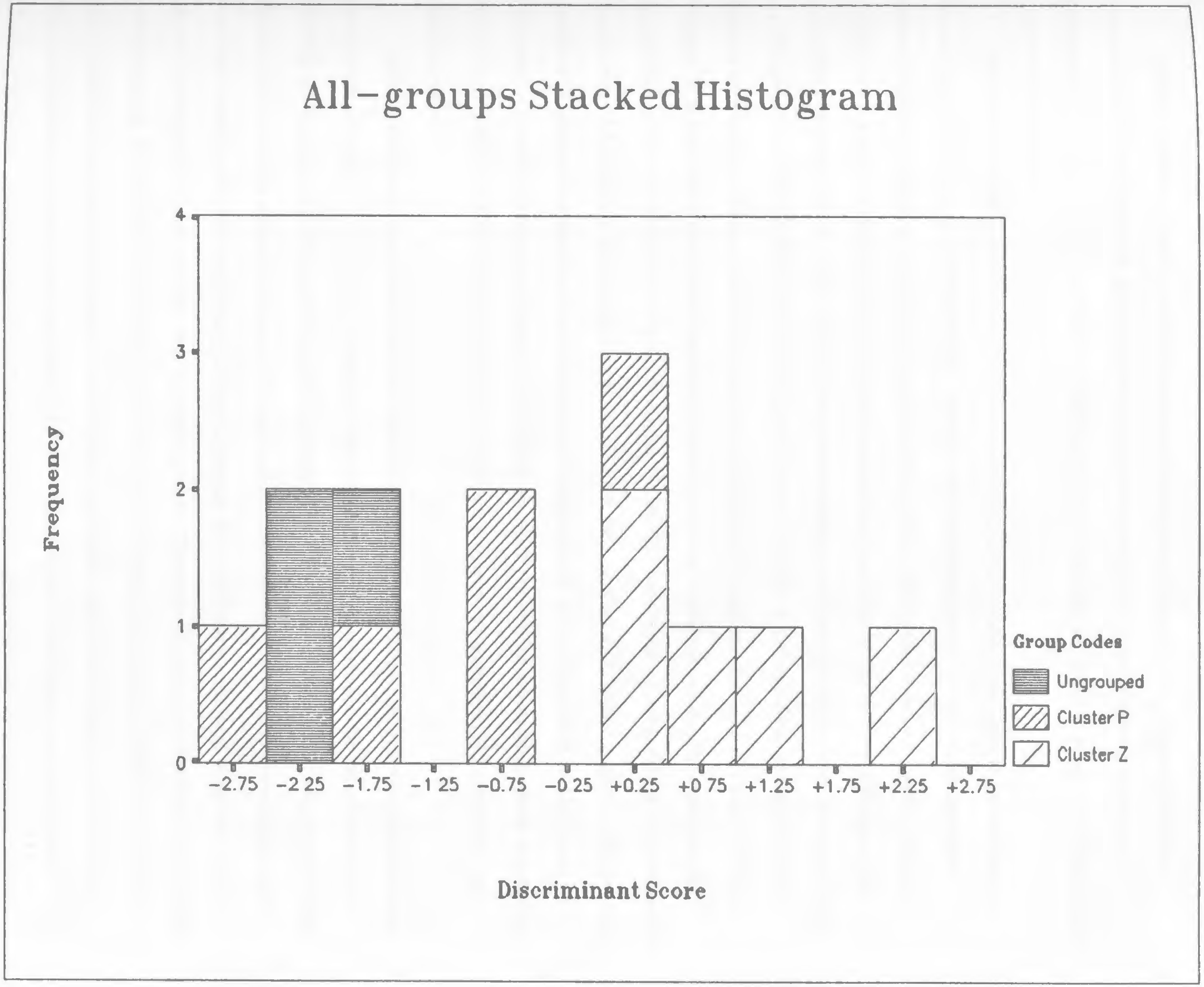

Figure 3-22: Stacked histogram of group samples as determined by the discriminant function.

3.6.5. Conclusions

The geochemistry data provided here constitute part of an on-going project which is concerned with the economic geology and drift geochemistry of Nachvak Fiord, northern Labrador. Analyses of data collected in 1984 and 1985 represent an interim report from the outer Nachvak Fiord area.

The geochemical properties of drift sheets located in the study area appear to be directly related to their lithological composition. Metal concentrations rarely exceed local threshold values, the latter being similar to those employed by Dyke (1983) on Somerset Island. Evans (1984) used Dyke's threshold values to isolate anomalous samples in the Selamiut Range. By using these criteria, Evans could only exclude three samples as non-anomalies from his total data set. Considering that Evans based his analysis on the clay size fraction of the samples, which appears to result in at least double the metal concentrations compared to the silt and clay size fraction, it is suggested that such threshold values are more applicable to geochemical results derived from the larger size fraction in the outer Nachvak Fiord area. Furthermore it is suggested that threshold values are site specific and should only be transferred to other regions with caution. Selection of anomalous samples was carried out independently by two methods in this analysis. Both sets of results correspond closely and suggest isolated pockets of anomalies which appear to reflect local bedrock characteristics. Regional drift sheets show no evidence of extremely high metal content.

Differentiation of drift sheets on the basis of their geochemical properties was an important goal of this analysis. In this respect discriminant analysis was quite successful and allowed the isolation of those metals which were good discriminators. The extrapolation of these geochemical characteristics to the adjacent Selamiut Range provided a test for the 'indicator variables' and provided further evidence supporting a regional origin for the Tinutyarvik and Naksaluk drift sheets. Differentiation of the drift sheets either side of Tinutyarvik moraine in Adams Lake valley was successful, but the discriminating variables may have been responding to the anomalously high values for samples from Adams Lake moraine and hence may have biased the end result. However it should be noted that sample G28 from Adams Lake valley was reassigned from Cluster P (Tinutyarvik drift sheet) to Cluster $N2$ (Adams Lake drift sheet) on the basis of the same geochemical criteria. Furthermore it seems apparent that if Adams Lake drift had a regional origin then similar regional drift sheets would show high metal concentrations. Instead it is only those sample sites related to local drift (G26, GAA, G33, G34, G36, G37, G40 and G43) which possess anomalously high concentrations.

In conclusion therefore, it is suggested that the geochemical composition of drift samples can be used as an indication of ice movement in the study area. Discriminating geochemical properties suggest that Tinutyarvik moraine in Adams Lake valley is the limit of regional drift (Tinutyarvik drift) south of the fiord, whereas Naksaluk Cove (Naksaluk drift) is the most easterly extent of regional drift along the fiord coast.

3.7. Summary

The previous sections have attempted to describe and synthesise the geomorphic and geologic evidence of glaciation in the study area. Although the different sources of information have been dealt with in discrete sections of the chapter, it is important to note that only those conclusions based on integrated observations and results can be regarded as unequivocal evidence for the style and extent of glaciation. The following is a summary of the conclusions that are supported by the preceding evidence.

The last regional ice advance into the study area terminated at the western end of Adams Lake and backfilled Naksaluk valley as far as Naksaluk moraine. This ice advance is discerned from previous glacial events on the basis of its drift properties. Both Tinutyarvik drift and Naksaluk drift display a predominance of glacial erratics from further west in the fiord, they possess a distinctive lithological and geochemical composition, and they have a similar till texture unrelated to tills further east in the study area.

The Adams Lake glacial phase is considered to be local in origin. The limit of this ice advance can be mapped using the maximum extent of lithological and geochemical characteristics associated with Adams Lake drift [Cluster *N* distribution]. Valley of the Flies terminal moraine marks this limit. MII moraine is considered to be the upper level of this advance in Naksaluk and Adams Lake valleys even though MII drift possesses a slightly different lithological and geochemical composition. Adams Lake moraine may therefore represent a recessional stage from this maximum position.

The Kammarsuit valley system preserves evidence of local cirque and valley glaciation. Drift lithology and geochemistry display characteristics consistent with a local ice origin, and moraine relationships on the valley slopes suggest that ice radiated out from the upland region. The presence of local drift at the mouth of Kammarsuit valley east and northern Valley of the Flies is a critical observation in the construction of maximum regional ice limits in the fiord.

It is suggested that the MI glacial phase represents an extensive regional ice advance which extended throughout the study area. This tentative conclusion is based primarily on the elevation and lateral extent of moraines and trimlines associated with this glacial event. MI drift characteristics are not typically 'regional' as defined by samples from Tinutyarvik and Naksaluk drift sheets; however, it is thought unlikely that the outer Nachvak Fiord area could host a glaciation of this magnitude in the absence of a regional fiord ice source.

Chapter 4

Acoustic stratigraphy of Nachvak Fiord and Adams Lake sediments

4.1. Introduction

This chapter describes the basin and the subbottom sediments of Nachvak Fiord and Adams Lake as interpreted from acoustic stratigraphic data. Analysis of these data has revealed approximately 180 m of sediment in the fiord and approximately 20 m of sediment in the lake. Such thicknesses of sediment were not evident in natural exposures in the study area and therefore these stratigraphic sections aid in the reconstruction of glacial and marine/lacustrine depositional environments during the late Quaternary period. The lateral tracing of sedimentary units in the fiord basin provides some continuity and control in the correlation of glacial events along the fiord hinterland. The steep fiord walls, that separate the tributary valleys of the fiord, preserve little evidence of past glacial events and therefore, in order to gain a regional perspective, terrestrial evidence must be extrapolated from cove to cove, since most of the coves are at the downstream end of tributary valleys which intercept the fiord coastline. Similarily, fiord acoustic data can be correlated with the extensive offshore acoustic stratigraphy, thereby providing an essential link between land-based and marine-based glacial chronologies.

4.2. Principles of acoustic profiling

Acoustic profiling is a seismic reflection technique carried out from a moving vessel. Acoustic pulses generated at a fixed time interval, travel down through the water column and are transmitted into, and reflected from, the sea/lake bottom and subbottom. Points of sound energy reflection denote changes in density of the material. The travel time of each reflected pulse is automatically measured and converted to a trace representing the depth on a recording chart.

The record of successive reflections produces a cross-section image of the sea/lake bottom and subbottom.

The vertical resolution of sucessive reflections in the acoustic record is determined primarily by the length of the acoustic pulse. Long acoustic pulses result in long reflections which will obscure subsequent reflections received shortly after the initial reflection. To avoid such information loss, a short acoustic pulse is required which can be generated using high frequency sound ($\geq$ 3 kHz). However, as sound attenuation in the ground increases exponentially with frequency, the use of high frequency sound limits the potential penetration. In some cases the depth of penetration can be increased by using greater output power and various amplification systems in combination with the high frequency acoustic source.

Although the records obtained from acoustic profiling techniques appear remarkably like drawn geologic sections, an understanding of some aspects of the geophysical system is required for an accurate interpretation (Tucker and Yorston, 1973; Payton, 1977; Van Overeem, 1978; Sheriff, 1980; McQuillin *et al.*, 1984). Some of the important principles of this technique are outlined below.

The area of the sea/lake bottom which receives the incident sound waves depends on the width of the sound beam, and on the water depth. The arc of the transmitted beam can be varied but a large beam results in less penetration. The area representing the reflected acoustic signals is smaller than the irradiated area as the points of reflection on the sea/lake bed act as a secondary source radiating sound waves in all directions. Only a small portion of the ascending wave front will be detected by the transceiver and converted into an electrical signal. This area that represents the received signal is known as the *Fresnel* zone. Continuous profiling records considerable overlap of Fresnel zones and therefore the recorded acoustic profile appears as a composite of repetitive signals, particularily in places with a highly variable reflectivity.

The ascending reflected signal may be reflected back downward from the water/sediment interface causing a *multiple* reflection which produces what appears to be a seperate event in the acoustic profile, since it is recorded some time after its original or primary reflection. The correct identification of a multiple reflection on the acoustic record, particularly the strong sea/lake bottom multiple, is important as these reflections may be misinterpreted as part of the stratigraphy in the geologic section.

The strength of the acoustic signal reflected from the sea/lake bottom or subbottom is generally regarded as indicative of the acoustic impedance of the reflection medium. However the loss of incident wave energy resulting from successive reflections down and up the geologic profile is usually compensated by the use of amplifiers. This results in the time-varied gain of reflected signals from deep layers of the profile. Therefore it is important to account for this compensation when the recorded strength of reflected signals is used in the interpretation of the acoustic profile.

Reflections may be detected by the transceiver from an object or landform located outside the Fresnel zone, if it is within the radiated area of incident wave energy, and if its surface is oriented favourably. This reflected signal will be received later than the reflected signal from the sea/lake bottom of the Fresnel zone. These signals are known as *side reflections* and may occur, for instance, when the recording vessel passes close to a rock knob on a basin slope, or on a sea/lake bottom. Side reflections are usually not repeated in the multiple reflections of the water/sediment interface (Van Overeem, 1978).

The presence of gas trapped in sediments is usually recognised by an abrupt decrease in penetration, and a characteristic lengthening of the reflected pulse. Normally the gas is dispersed through the sediment in small bubbles and extreme reflection or excessive attenuation of the acoustic energy occurs, preventing deeper penetration.

The difference between horizontal and vertical scales on the acoustic profile means that recorded slopes are greatly exaggerated. Small apparent slopes can be accurately corrected if several profiles are recorded at various angles to the slope. Actual slopes steeper than 20-30° occurring in a geologic section will rarely be observed, or predicted with much accuracy.

Verification of interpreted acoustic profiles can be carried out through examination of core logs from the survey area. However, such data are limited both in penetration depth and aerial extent. Sheriff (1977) suggests that stratigraphic interpretation of seismic data "is a constrained art, limited by fundamental considerations". It is important, therefore, to be aware of the limitations of acoustic profiling techniques and the geophysical system on which they rely. Once this information is appreciated, "successful stratigraphic interpretation....has to be a combination of three elements: principles, experience, and imagination" (Sheriff, 1977).

4.3. Nachvak Fiord

In 1984 a 3.5 kHz acoustic survey[10] was carried out in Nachvak Fiord. This survey provides preliminary information on the fiord basin morphology and subbottom sediments between the fiord threshold in Nachvak Bay and Townley Basin (Fig. 4-1). A total of 20 acoustic traverses was recorded along the fiord, providing for frequent cross-over points and correlation of acoustic profiles. The total length of the acoustic record exceeded 125 km with average penetration depths of 50 m. The main profile, line 19, is a transect along the centre of the fiord (Fig. 4-1), and represents the general longitudinal profile of the fiord basins and the subbottom stratigraphy of the uppermost 50 m of sediment. Results of this survey were published by Rogerson *et al.* (1986*a*). The following sections are a reinterpretation of this analysis, using recently acquired seismic data[11].

In 1985, seismic profiles[12] were recorded along the centre of the fiord, from Koktortoaluk Basin to the fiord threshold (Fig. 4-1), using both an air gun source capable of up to 500 m penetration, and the Huntec deep tow system which provided high resolution data with at least 50 m penetration. This additional information complements the 3.5 kHz acoustic data and provides a complete profile of the total sediment in the fiord basins.

Nachvak Fiord is a 45 km long glacial trough which cuts through the Selamiut Range of the Torngat Mountains. The fiord is 2 to 4 km wide, increasing gradually eastwards until it reaches Nachvak Bay which opens to the Labrador Sea (Fig. 4-1). The upper part of the fiord divides into Tallek Arm and Tasiuyak Arm which cut approximately due south for 12 km. At the head of Tasiuyak Arm, the glacial trough reverts to its westerly trend for a further 10 km along Nachvak Lake valley. The sidewalls of the fiord are generally steep, rising 1000 m vertically from sea level at Kutyaupok Mountain. The fiord cuts across the NNW to SSE structural grain which was imposed on the area by Precambrian tectonism (Wardle, 1983) (Fig. 2-4) and which controls the trend of mountain ranges and the drainage into the fiord.

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[10]This survey was carried out as part of a co-operative program between the Atlantic Geoscience Center and the Canadian Hydrographic Service.

[11]The acoustic unit notation used in the following sections does not correspond with that in Rogerson *et al.* (1986*a*).

[12]These records were obtained by B. MacLean, Atlantic Geoscience Center, Dartmouth, on the CSS Hudson cruise 85-027. Permission to refer to a copy of these records for the purpose of this thesis was kindly granted by H.W. Josenhans (personal communication, 1986).

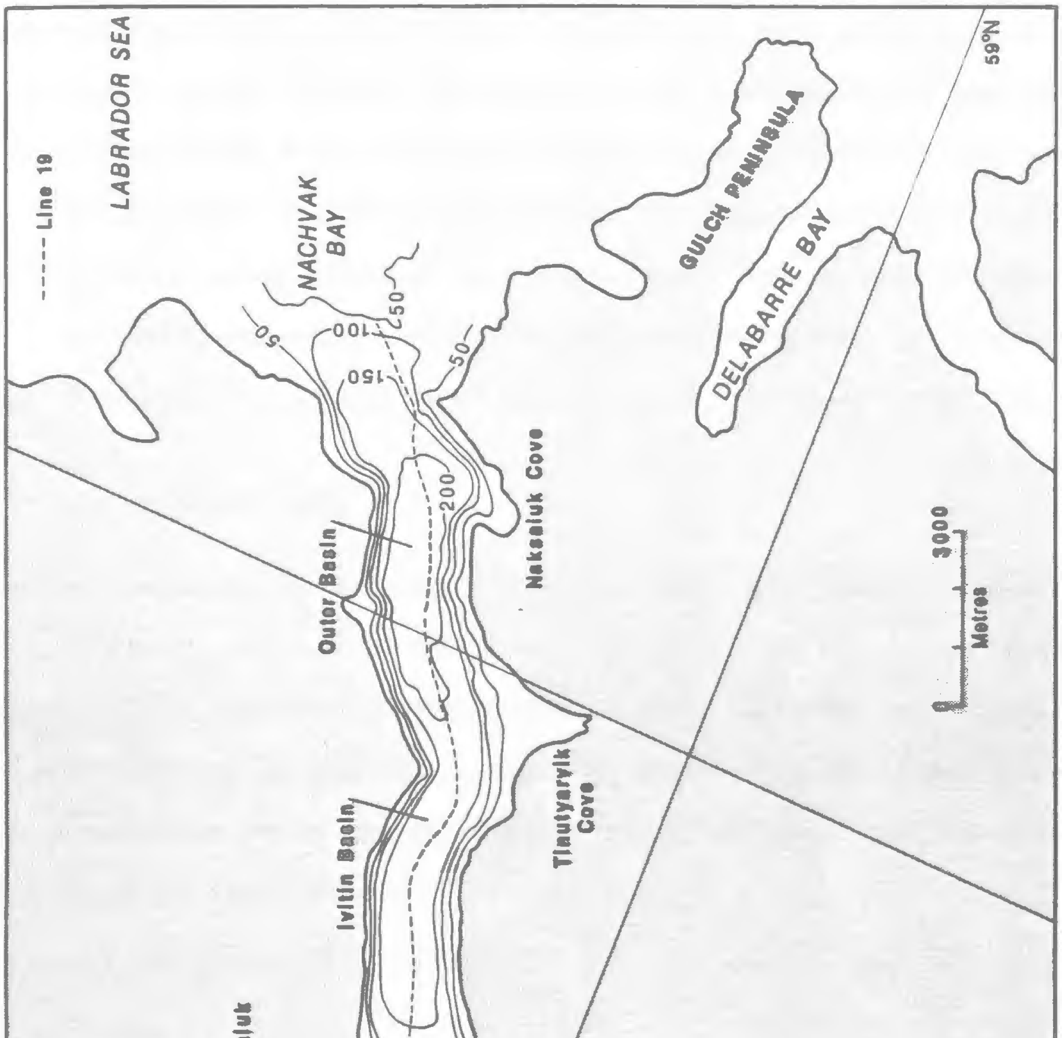

Figure 4-1: Nachvak Fiord bathymetry with location of basins and barriers. The dotted line represents the position of the 3.5 kHz profile shown in Figure 4-3.

Bathymetry in the fiord reveals a series of four basins between Townley Head and the fiord threshold. Tasiuyak Arm, to the west of Townley Head, was not investigated and may be either a single, separate basin or more than one. From west to east, the basins which were investigated are the Townley, Koktortoaluk, Ivitin and Outer Basins. Maximum water depths in the four basins are 90 m, 160 m, 170 m, and 210 m, respectively. The water depth is very shallow over the fiord threshold with frequent shoals and an average channel depth of 50 m extending east into Nachvak Bay. The four basins are separated by barriers which range from $\leq$ 50 m to 180 m below sea level. From west to east, these are Kogarsok, Ivitak, and Shoal Water Cove barriers.

4.3.1. Subbottom sediment units

The three sources of acoustic and seismic data are integrated to produce a stratigraphic section of fiord subbottom sediments. Cores taken in 1985 penetrated almost 6 m of the uppermost sediment (H.W. Josenhans, personal communication, 1986) and their analyses will provide a degree of control to the proposed stratigraphy. Figure 4-2 is the air gun profile of the fiord bottom and subbottom east of Kogarsok barrier. Figure 4-3 is the 3.5 kHz acoustic profile along line 19, and includes Townley Basin.

Figure 4-2 is available as three sections in the wallet at the back of this thesis.

Figure 4-2: Air gun profile of Nachvak Fiord Basin and subbottom sediments.

Six acoustic units, A to F, are identified from the above records. They are illustrated in Figure 4-2 and diagrammatically presented in Figure 4-4. Total sediment thickness is >50 m in Townley Head Basin, 140 m in Koktortoaluk Basin, 180 m in Ivitin Basin, and >160 m in Outer

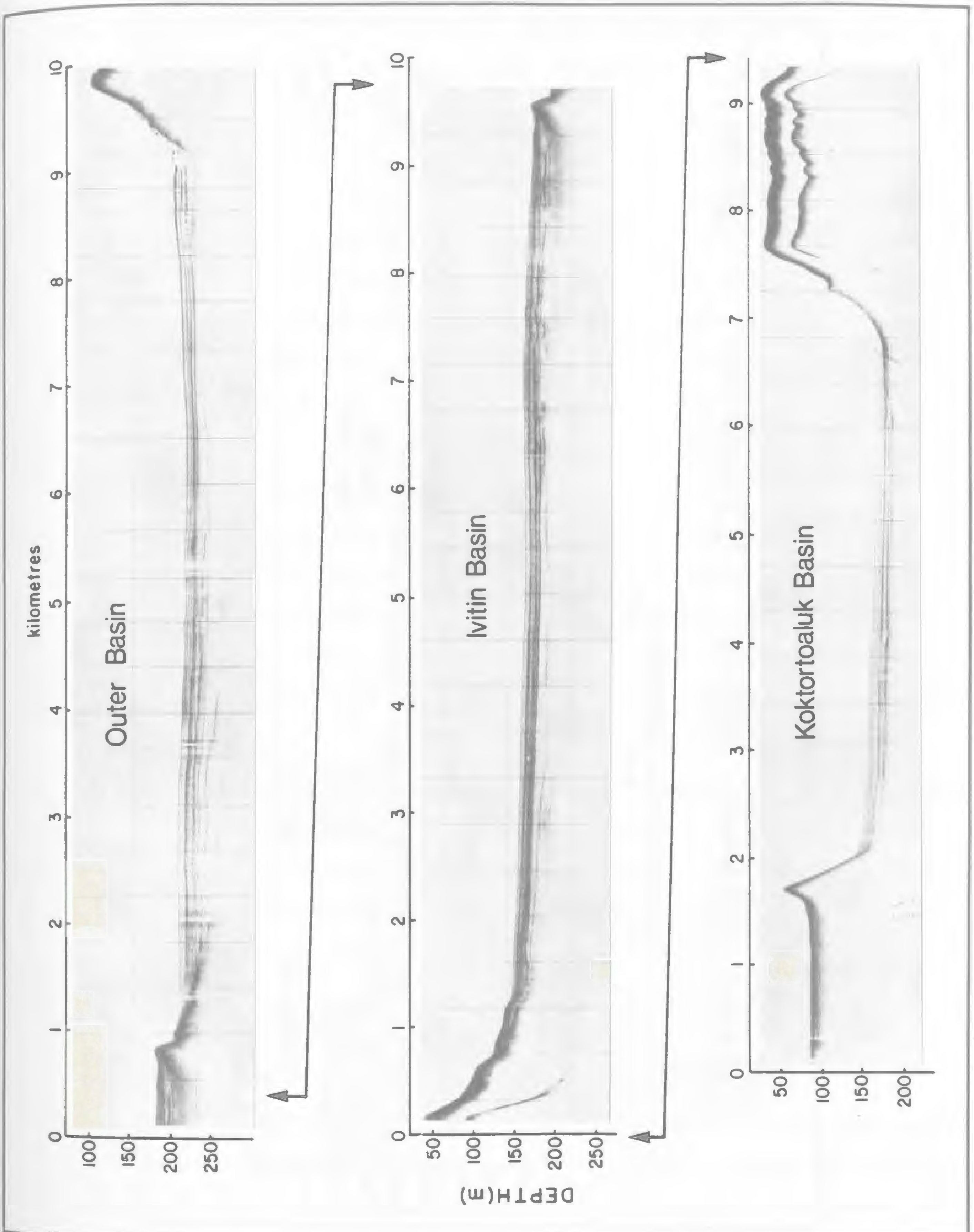

Figure 4-3: 3.5 kHz acoustic profile of Nachvak Fiord subbottom sediments.

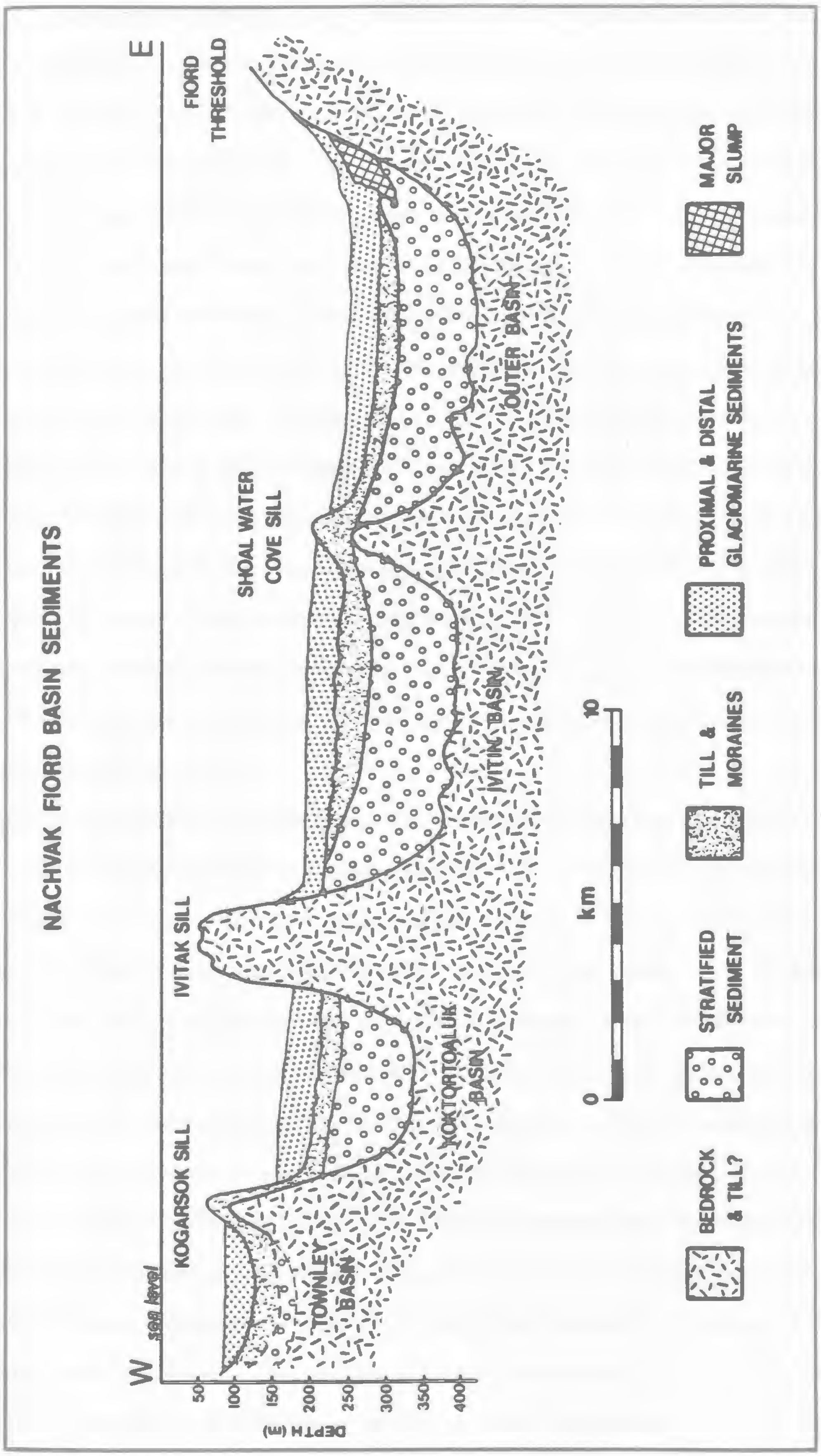

Figure 4-4: Diagrammatic-interpreted section along centre of fiord. Interpreted stratigraphy is based on analyses of acoustic profiling records.

Basin. The 3.5 kHz system does not have sufficient energy to penetrate below the uppermost 50 m of sediment and therefore only Units C to F are observed on Figure 4-3.

Unit A is the impenetrable base of the fiord, consisting of crystalline and metasedimentary rocks. The extrapolation of the bedrock geology units (Fig. 2-4) across the fiord suggests the approximate locations of sills which are verified by the presence of bathymetrically-high barriers in Figure 4-2. Kogarsok and Ivitak sills appear to be composed of anorthosite whereas the Shoal Water Cove sill is located near the Nachvak Fiord Thrust and is probably composed of Nachvak gneiss. Submarine exposure of the bedrock that composes the fiord basement is limited to the Ivitak sill and the fiord threshold; elsewhere, it is covered by acoustic reflectors.

The interference that occurs above Unit A in Ivitin and Outer Basins on Figure 4-2 is interpreted as side reflections, recorded from the basin slopes. The exact navigation of the recording vessel is not known and therefore this interpretation is speculative. However these reflections appear to mask the acoustic units overlying Unit A, as opposed to recording stratigraphical relationships within the basins. In some cases acoustic reflectors can be traced from Unit B into the interference area. There is a faint multiple reflection of the interference pattern on the acoustic record.

Unit B is an acoustically stratified unit with maximum thicknesses of 80 m, 130 m, and 80 m in Koktortoaluk, Ivitin, and Outer Basins, respectively. There is no record of Unit B in Townley Basin, but this is a result of the limited penetration of the 3.5 kHz system and therefore it may exist below the 50 m range of the acoustic profile in this basin. Unit B reflectors are predominantly parallel to sub-parallel and dip sub-conformably with the slope of Unit A at the western margins of the basins. Along the eastern margins they show an onlap relationship with the fiord basement. In the centres of the basins, they appear to fill the basin-form of the upper surface of Unit A and are recorded as flat, parallel reflectors, 1-2 m apart.

The surface of Unit B that underlies Unit D in the western part of Ivitin Basin displays either an oblique progradational reflection configuration or an erosionally truncated surface Below this surface area reflectors are chaotic to wavy with internal convergence (offlap) occurring towards the centre of the basin. The surface of Unit B underlying Unit C in the western part of this basin also displays a top-discordant relationship which may be described as truncation.

The surface of Unit B at the eastern margin of Ivitin Basin is coincident with the upper surface of Unit A (Shoal Water Cove sill). A similar situation occurs at the western margin of Outer Basin where the upper limit of the easterly dipping reflectors occurs at approximately 230 m below sea level, the upper surface of the Shoal Water Cove sill. Unit B possesses the same reflection characteristics in Outer Basin as in Ivitin Basin. A chaotic to wavy reflection pattern gradually changes to a flat, or slightly dipping, parallel configuration towards the centre of the basin. (Side reflections obscure most of this unit in the centre of the basin.) The upper surface of this unit underlies Unit C as an unconformable boundary, the truncation of reflectors being evident along the contact plane.

Unit B in Koktortoaluk Basin is subdivided into two subunits, B_1 and B_2. Both subunits have similar parallel to sub-parallel reflection patterns and have undergone post-depositional deformation as a single unit along the same fault and fold planes. The two subunits are differentiated on their contrasting relationships with the fiord basement. Subunit B_1 displays basin-fill characteristics with onlap occurring on both margins of the basin. Subunit B_2 possesses a steeply-dipping, diverging reflection pattern, conformable with the basin margins. This configuration suggests a sediment source to the west, with possible lateral variations in the rate of deposition (truncation of reflectors upslope). Alternatively, the apparent divergent reflection pattern may be due to progressive thinning of strata to below the resolution of the air gun system.

The two subunits are discerned along a discordant boundary which primarily signifies the lower limit of easterly dipping reflectors at the western margin of the basin. Coincident with this observation is the apparent baselap relationship of the lowest reflectors of subunit B_2 with the upper surface of subunit B_1. This reflection configuration may be described as either downlap or onlap, denoting the lateral commencement of sediment deposition. However this interpretation ignores the possible post-depositional deformation of initially concordant strata which could result in a contorted or hummocky reflection configuration.

Unit B is bounded by unconformities at its lower surface with Unit A, and at its upper surface with Units C and D.

Unit C occurs in the four basins of the fiord. Total penetration of this unit was achieved with the air gun system in Koktortoaluk, Ivitin, and Outer Basins, but only partial penetration

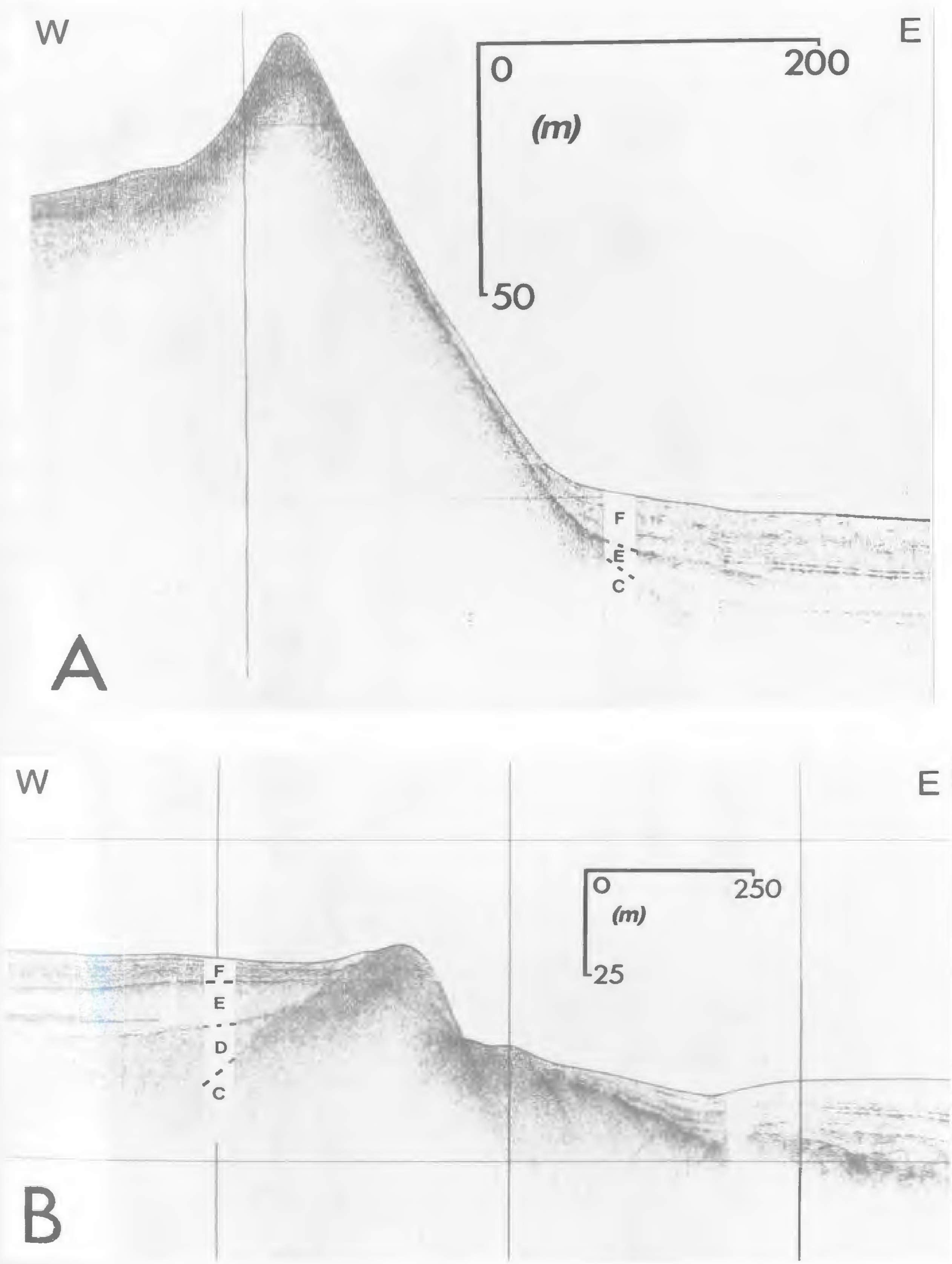

Figure 4-5: 3.5 kHz profile of moraine ridges at [A] Kogarsok Barrier and at [B] Shoal Water Cove Barrier.

was recorded with the 3.5 kHz acoustic system (Figs. 4-2 & 4-3). Maximum sediment thicknesses for this unit are $\geq$ 20 m, 30 m, 50 m, and 35 m in the four basins from west to east in the fiord. The external form of this unit varies from wedge-shape in Koktortoaluk Basin to lens-shape in Ivitin and Outer Basins. The areas of the fiord where Unit C is not covered by sediments contain some well-defined morphologies. The barrier overlying Kogarsok sill, between Townley Basin and Koktortoaluk Basin is particularly distinctive, consisting of a steeply sloping, narrow ridge oriented across the fiord (Fig. 4-5). Similarly, the barrier overlying Shoal Water Cove sill, between the Ivitin and Outer Basins is a narrow ridge sloping down both upfiord as well as downfiord (Fig. 4-5). Both features have the geomorphic characteristics of moraine ridges.

Internal reflectors in Unit C have a hummocky to lenticular configuration. In Koktortoaluk Basin there appear to be two distinct subunits within Unit C. The lower subunit C_1 forms the basis of the wedge-shape unit, but overlying this, is a lens-shape subunit C_2. The lower boundaries of Unit C in the basins are similar to the general trend of stratification in Unit B, although in some areas the contacts are unconformity surfaces. In contrast, the upper boundary of Unit C has a hummocky, irregular surface which mimics the configuration of the internal reflectors (Figs. 4-6 & 4-7).

Figure 4-6 is available in the wallet at the back of this thesis. Note orientation of section.

Figure 4-6: Huntec profile of western portion of Koktortoaluk Basin sediments.

Unit C is interpreted as representing glacial till and moraine complexes. This interpretation is confirmed by traverses in 1985 by the submersible *Pisces IV* which revealed that the Shoal Water Cove feature is a steep bouldery ridge with no evidence of exposed bedrock (Rogerson *et* *al.*, 1986*a*). The surface hummocks of Unit C may also be regarded as moraine-like features; however, justification for this interpretation will be presented in the next section.

Figure 4-7 is available in the wallet at the back of this thesis. Note orientation of section.

Figure 4-7: Huntec profile of western portion of Ivitin Basin sediments.

Unit D occurs in the four basins of the fiord and can be subdivided into two subunits, D_1 and D_2. Subunit D_1 consists of the ponded sediments which are located in the depressions associated with the surface of Unit C (Fig. 4-7). These sediments attain maximum thicknesses of 6 m in places. Subunit D_2 is finely stratified with parallel reflectors. One strong internal reflector occurs half-way down through the unit and reflects sufficient energy to almost obscure the fine stratification of the lower sediment. Maximum thicknesses of subunit D_2 are $\leq$ 5 m, 15 m, 15 m, and 20 m in the four basins, from west to east. Characteristic of this unit are the major slumps that occur along its upper surface near the basin margins (Figs. 4-7 & 4-8). In some cases these slumps deform the upper 5-7 m of sediment. Unit D is interpreted as an ice-proximal deposit displaying cyclic sedimentation of fine grained sediment.

Unit E is 10 m thick in Koktortoaluk and Outer Basins, 7 m thick in Ivitin Basin, and <5 m thick in Townley Basin. It is an acoustically stratified unit with reflectors varying in strength and amplitude. The reflection configuration is predominantly parallel, mimicking the upper surface of Unit D. Where no slumps occur, an onlap relationship is apparent with the basin margin. This unit contains minor gas pockets and displays lens-shaped ponded features at its lower boundary. Unit E is interpreted as ice-proximal to ice-distal glaciomarine sediment which has settled out along the bottoms of the basins. Acoustic variability within the unit may be due to changes in sediment type and influx associated with a receding or fluctuating ice margin.

Figure 4-8 is available in the wallet at the back of this thesis. Note orientation of section.

Figure 4-8: Huntec profile of eastern Outer basin sediments

Unit F occurs as a sheet drape over the underlying sediments and barriers. It is thickest in the centre of the basins but thins along the lower slopes of the basin margins and may be absent on the steeper slopes. It attains maximum thicknesses of 3 m, 6 m, 8 m, and 10 m in the four basins from west to east along the fiord. The lower strata have an even, parallel reflector configuration while the upper strata are acoustically transparent and provide a conformable cover on the fiord bottom. Minor slumps disrupt this reflection pattern and gas pockets result in contorted reflectors. Frequent point source reflections occur in the lower strata and are interpreted as dropstones. Unit F is continuous with modern sediments which are presently being deposited on the fiord bottom in basin locations.

The lower part of Unit F is interpreted as ice-distal sediments; these are generally restricted to the basin bottoms. The lateral continuity of reflectors with their even, parallel configuration suggests that they were deposited through uniform rain out of fine sediments. The presence of dropstones in these strata is evidence of deposition from icebergs. The upper part of Unit F reflects depositional processes similar to those operating in the fiord environment at the present time.

In Koktortoaluk Basin, Unit F contains a wedge of sediment (subunit F_1) between 1 m and 10 m in thickness (Figs. 4-2, 4-6 & 4-9). It is a stratified deposit that pinches out to a single reflector towards the eastern margin of the basin. Significant slumps occur along the surface of this subunit.

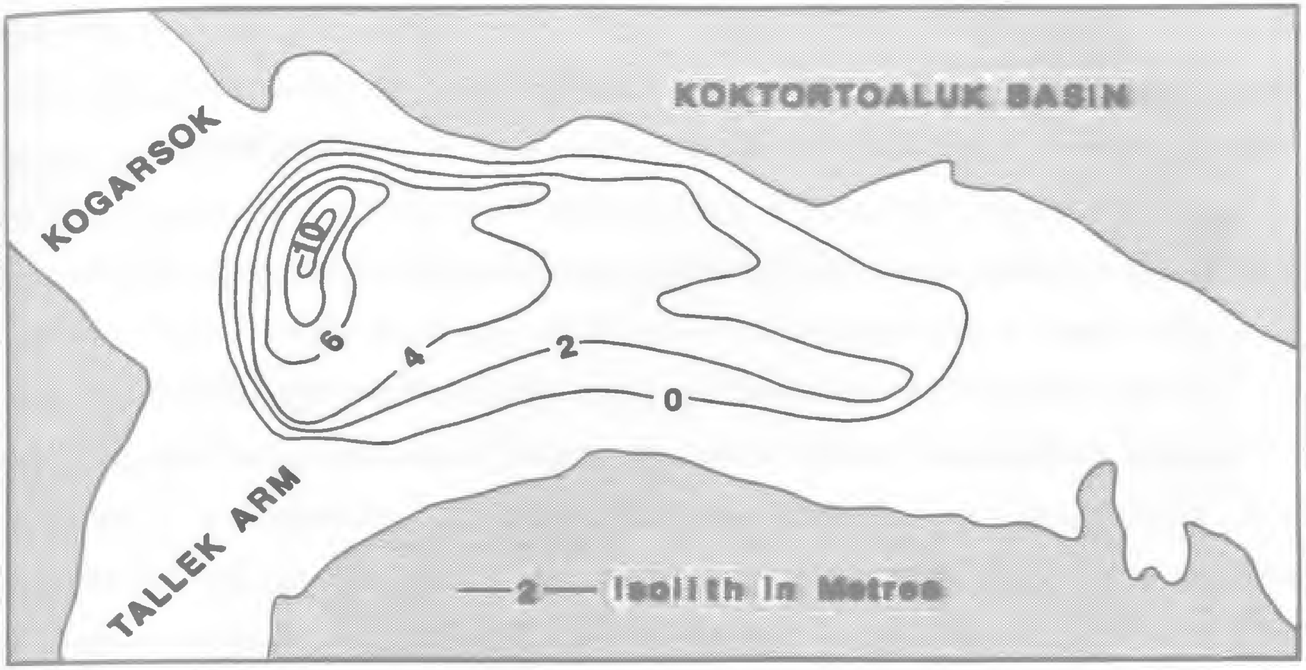

Figure 4-9: Isopach map of the stratified subunit within Unit F in Koktortoaluk Basin.

4.3.2. Discussion

Any depositional model proposed for Nachvak Fiord sediments, based on the data outlined in the previous section, must involve a large degree of speculation, as the interpretations of acoustic units are tentative and the profiles of lower units (A-C) are recorded from a single traverse along the centre of the fiord. While acknowledging the restrictive nature of the data base, an attempt is made to synthesize it into a working hypothesis which can be tested against terrestrial evidence from the study area.

Unit C is interpreted to be a glacial till based on its reflector-free acoustic properties, and on the confirmation of its classic moraine morphology where exposed on the fiord bottom. Therefore any glacial depositional model constructed for Units C to F must account for the stratigraphical relationship between this till unit and the underlying acoustically stratified Unit B. (The origin of Unit B will be discussed later in this section.) The lower surface of Unit C is generally conformable with the trend of stratification in Unit B, although the boundary between these units is an unconformity in Ivitin and Outer Basins. Where Unit C pinches out at the margins of these two basins (240 m below sea level in Outer Basin and 190 m below sea level in Ivitin Basin), a distinct erosion surface can be identified coincident with marked deformation and loss of stratification in Unit B. Either side of Shoal Water Cove sill similar deformation in Unit B can be observed. It is proposed that the distinct erosion surfaces at the margins of these basins and the top surface of Shoal Water Cove sill (approximately 230 m below sea level) represent grounding points of an ice shelf which extended over the fiord basin. If these three points are connected to form a grounding plane, then the resulting easterly dipping plane possesses an approximate slope of 1:500. This slope is similar to the calculated slope of 1:700 for the highest raised marine shoreline reconstructed from evidence in the study area[13] (See Section 5.3.3, Table 5-6). Therefore this plane was approximately horizontal in the isostatically depressed ice-marginal environment. A third grounding point may be represented by the top of the Kogarsok sill (140 m below sea level) which intersects this plane to the west. The marked deformation in Unit B below, and adjacent to the grounding points may reflect the loading pressures exerted on these deposits.

Subglacial deposition is postulated for Unit C, occurring as the ice shelf thinned and melted along its base. This unit is probably time-transgressive throughout the fiord basins, deposition occurring first in the deeper Outer basin. Shoal Water Cove moraine is interpreted to be a lift-off moraine, formed as the ice shelf became increasingly buoyant in response to rising sea levels and ice thinning. The moraine is a narrow ridge sloping down both upfiord and downfiord, a description consistent with a lift-off origin (King and Fader, 1986). Kogarsok moraine has a distinct asymmetrical end moraine shape and may reflect deposition against a grounded ice margin, during ice recession.

The upper surface of Unit C in Koktortoaluk and Ivitin Basins is characterized by hummocks with variable relief (in some cases >8 m). These features are interpreted as lift-off moraine ridges[14], and may have been formed in response to changes in ice shelf buoyancy and the position of grounding lines. These moraines may represent surges or readvances of the ice shelf into these basins as there appears to be a multiple till section within Unit C in Koktortoaluk

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[13] It is suggested that the slope of this grounding plane/erosion plane should be greater than that of the highest marine shoreline in the study area as isostatic recovery would have occurred as ice receded, prior to the development of shoreline processes. The 'approximate' nature of this slope comparison is stressed considering the extremely rough calculation of the grounding plane slope (measured along one traverse only).

[14] These features are recognisable on the extensive 3.5 kHz survey and therefore are not thought to be isolated hummocks of Unit C drift.

Basin, and the thickness of Unit C in Ivitin Basin is much greater than that in Outer Basin. These lift-off moraines generally occur in association with subunit D_1 which appears to be a ponded sediment, defined by abrupt termination of its reflectors against the steep slopes of the moraine ridges (Figs. 4-6 & 4-7). Contemporaneous deposition of the lift-off moraines and the glaciomarine sediment is suggested because the later deposition of subunit D_2 clearly mimics the ridge morphology.

Development of open water under the ice shelf is proposed for the initiation of subunit D_2 deposition. This sediment is characterized by finely stratified acoustic reflectors which may represent periodic deposition from suspension. The conformable, uniform character of the regularly spaced reflectors suggest sediment influx from a blanket source. There is no indication that this subunit was deposited by underflows generated at the buoyancy line as postulated by Mackiewecz *et al.*, (1984) for apparently similar ice-proximal glaciomarine sediments in Muir Inlet, Alaska. King and Fader (1986) identified similar sedimentary characteristics in facies A of the Emerald Silt on the continental shelf off Nova Scotia and Newfoundland and supported their observations with data from core analyses. Consequently, they were able to define mechanisms for the release of material from the base of the ice shelf in order to explain the rhythmically banded silt and clay facies. Drewry and Cooper (1981) suggested that cyclic variations in the release of sediment from an ice shelf may be due to a freeze-thaw cycle at the subglacial surface around grounding-line zones, an interpretation supported by King and Fader (1986). Kontopoulus and Piper (1982) identified similar cyclicity in their facies E, described from cores taken in Kaipokok Bay, Labrador. Using a floating ice model, they suggested that variable discharge resulted in the deposition of cyclic layers of mud and muddy sand, with sub-ice tidal currents playing a major role in sediment dispersal and deposition.

The greater thickness of subunit D_2 in the Outer Basin may suggest that open water conditions developed earlier in this deeper basin, or that approximately 5 m of this subunit was eroded from Koktortoaluk and Ivitin Basins during a readvance, or during repeated oscillation of the grounding-line in these basins. The surface and upper strata of Unit D are associated with major slumps at the basin margins. These slumps may have resulted from the loss of ice support and subsequent sediment dilatancy as the ice shelf thinned and became buoyant

The variability in the acoustic stratification of Unit E is suggested as evidence for a fluctuating ice margin in a proximal to distal location. It is proposed that ice shelf disintegration had occurred and that ice cover was restricted to drifting ice during this period. Deposition of Unit E may have been bimodal with stream underflow and sediment gravity flow as major elements, and iceberg sedimentation constituting a lesser input. The resultant facies is similar to that described by Powell (1981) for glaciomarine sediments associated with a rapidly retreating tidewater glacier with an ice front actively calving in deep water. Such ice front conditions may have existed at Kogarsok moraine and in locations further west from Townley Basin. Minor slumps are evident in, and on the surface of this unit. These may originate from high or variable sedimentation rates resulting in over-steepening of slopes, or iceberg activity close to basin margins.

Unit F is interpreted as comprising ice-distal sediments continuous with modern sediments. The well-stratified and laterally continuous reflectors in the lower strata suggest that this unit was formed by uniform rain out of suspended material, complemented by iceberg sedimentation processes. The upward decline in the number of dropstones, and the acoustic transparency in the upper part of this unit suggest the waning influence of icebergs and probably, the absence of an active calving ice front.

The wedge of acoustically stratified sediment, subunit F_1 in Koktortoaluk Basin may be associated with a significant influx of sediment from Tallek Arm, to the south of the basin. This subunit does not reflect the characteristics of a slump (Rogerson *et al.*, 1986*a*), although it is overlain by slumped material, possibly associated with the undercutting of sediments at the base of Kogarsok moraine (Fig. 4-6). A single 3.5 kHz acoustic traverse of Tallek Arm revealed modern deltaic sediments underlain by acoustically stratified units. These latter sediments may be related to an ice-dammed condition in the arm while ice occupied Koktortoaluk Basin. Alternatively, they may originate from the catastrophic emptying of glacier-dammed lakes west of the drainage divide. Air photo interpretation indicates that at least one such lake did exist in the upper Korok valley. It overflowed to the north, down the Palmer River into Tallek Arm (R. Klassen, personal communication, 1985).

It is suggested that Units C-F represent a glacial-deglacial cycle, progressing from a partially grounded ice mass phase, through a floating ice shelf phase, to ice front recession and eventual total deglaciation of the fiord basins. Readvances and ice front oscillations are tentatively proposed as integral parts of this cycle. Unit B therefore, was deposited prior to this glacial event.

Unit B is interpreted as a stratified, basin-fill deposit with steeply dipping strata at the western margins of Koktortoaluk, Ivitin, and Outer Basins, indicating a westerly source for this sediment. Subunit B_1, in Koktortoaluk Basin, has a similar reflection pattern, although it is suggested that the likely origin of this deposit is not from the west, but from Tallek Arm to the south. This is concluded on the basis of the most probable source of sediment influx other than Townley Basin, as the reflection configuration of this subunit does not indicate a sediment source from the west.

A relationship between the deposition of Unit B in Ivitin and Outer Basins is tentatively suggested. If these basins are regarded as settling ponds for sediment influx then it is probable that deposition of Unit B occurred across the Shoal Water Cove sill and into Outer Basin from a sediment source at Ivitak sill. The strata overlying Shoal Water Cove sill were subsequently eroded, although the steeply-dipping reflectors adjacent to the sill in Outer Basin may be the truncated remains of these strata.

Two hypotheses are proposed regarding the origin of Unit B, neither of which can be supported with any definitive evidence. First it is suggested that these sediments represent marine/lacustrine deposition of terrigenous sediment influx. The steeply dipping reflectors at Ivitak and Kogarsok sills suggest sediment sources at these locations which are coincident with major drainage input to the fiord. The McCornick River valley drains most of the high mountains in the Selamiut Range and flows into the fiord at Ivitak Cove, while the Kogarsok and Palmer Rivers drain most of the fiord hinterland west of Ivitak Cove and flow into the fiord at Townley Basin and Tallek Arm, respectively. Deposition may have been into a marine or lacustrine environment as there is the possibility that this event was coincident with a low sea level stand, and therefore the lowest point of the fiord threshold (50 m below sea level) would determine the character of the depositional environment. If sea level fell by more than 50 m then the fiord would exist as two lakes, (acting as settling ponds) with Ivitak sill sub-aerially exposed between them.

Alternatively, Unit B may be a proglacial deltaic unit deposited during the penultimate deglaciation of the fiord. This interpretation suggests that Kogarsok and Ivitak sills were sites of major still-stands during ice recession. However, no glacial drift associated with these sediments was observed on the acoustic profiles. It is possible that Unit A consists not only of the fiord bedrock basement, but also acoustically impenetrable tills which reflect the total incident acoustic energy. The side reflections recorded above Unit A in the basins may be interpreted as till-like units because of their high reflectance and acoustically massive structure. However, this latter interpretation is suggested to be incorrect for the reasons outlined in the previous section.

4.3.3. Regional Context and Interpretation

This section on the regional context of Nachvak Fiord sediments is confined to the marine geology of the Labrador Shelf and adjacent shelves off Baffin Island and off southeast Canada. Chapter 7 attempts to integrate terrestrial evidence from the study area with the tentative depositional model proposed for Nachvak Fiord to produce a regional stratigraphic and chronologic framework.

Quaternary geologic studies on the continental shelves listed above have suggested that pre-Late Wisconsinan ice advances were much more extensive than the Late Wisconsinan ice advance. King and Fader (1986) have proposed a model for glaciation on the southeast Canadian continental shelf. They suggested that in Early Wisconsin time the entire shelf was occupied by an ice sheet which wasted to an ice shelf during the early part of the Middle Wisconsin, and eventually receded to the coastline of Maine and New Brunswick by the Late Wisconsin. Recognition of a specific Late Wisconsinan ice readvance is restricted to local areas of moraine building and drift deposition.

Praeg *et al.* (1986) described the Quaternary geology of the southeast Baffin Island continental shelf. They identified an extensive drift unit over the shelf which is suggested to be of pre-Late Wisconsinan age (MacLean, 1985). Overlying glaciomarine sediments record deposition from Mid-Wisconsinan time to the Holocene, and contain evidence of ice-proximal to ice-distal glaciomarine environments. The margin of Late Wisconsinan ice is thought to be limited to the mouths of the fiords on southeastern Baffin Island.

Fillon and Harmes (1982) developed a glacial depositional model for Labrador Shelf sediments based on acoustic stratigraphy and lithological evidence. They envisaged ice grounding on the shelf during the Late Wisconsinan maximum, followed by basal melting and ice lift-off in the shelf basins by 9770 years BP. During the period 9770 - 8380 years BP glaciomarine muds, originating as runoff from ablating valley glaciers in Labrador, were laid down in the basins. This sequence was then overridden by a possible glacial surge from either Hudson Bay, Ungava Bay or Foxe Basin, 8380 - 7500 years BP. Gradual deglaciation of tidewater glaciers was followed by deposition of ice-rafted debris.

Josenhans *et al.* (1986) described a deglacial sequence of till and overlying glaciomarine sediments on the Labrador Shelf which was tentatively dated at $\leq$ 20 ka BP, and represents an extensive Late Wisconsinan ice advance. Underlying this stratigraphic sequence is an older till sheet which extends over the shelf margin. They also recognised a significant volume of sediment associated with a late readvance of ice from Hudson Bay. This carbonate-rich silt unit (Qeovik silt) directly overlies the upper till; evidence, they suggested, for a rapidly retreating, non-depositional ice shelf retreat phase, directly followed by ice-rafted sediment deposition from Hudson Bay.

The contribution of the tentative depositional model proposed for Nachvak Fiord sediments to the general debate on the maximum extent of ice during the Late Wisconsin is limited, due to the following restrictions: (1) there is no acoustic stratigraphic continuity from shelf to fiord bottom sediments, (2) at present, there is no dating control on any of the inferred sedimentary facies from the fiord, and (3) no lithological data are available for fiord subbottom sediments which may act as a correlative tool for shelf and fiord sediments.

While acknowledging these restrictions, several general comments can be made concerning the incompatibility of observed fiord stratigraphy and the style of glaciation proposed for the Labrador Shelf by Fillon and Harmes (1982) and Josenhans *et al.* (1986). Both these studies suggested that Late Wisconsinan ice extended onto the shelf as a grounded ice mass. Clark and Josenhans (1986) and Josenhans *et al.* (1986) proposed a low-sloping, thin piedmont glacier style for ice advance onto the shelf and suggest that the ice was grounded, yet close to hydrostatic equilibrium, so that the upper till unit could have been deposited in an ice-contact environment. They also suggested that this ice advance was significantly erosive, resulting in the "development of smooth unconformities between tills and a lack of stratified intertill sediments" (p. 1211). They explained the absence of any stratified glaciomarine sediment, overlying the till unit, associated with the deglaciation of this ice margin by proposing that there were low quantities of englacial debris in the ice, due to reduced erosion over a highly resistant glacier bed. The inherent inconsistencies in this model are compounded by the fact that the upper till unit is found at depths of 750 m below present sea level. To retain a thin ice mass grounded at this depth would require that relative sea level fell to an extremely low level during the Late Wisconsinan period. Abundant evidence exists, in the field and in the literature that relative sea levels were higher than present around the margins of the Laurentide ice sheet during glacial maxima. The identification of thick (80-130 m) stratified deposits (Unit B) below the till unit (Unit C) in Nachvak Fiord suggests that the last glacial advance in this region did not erode sediment deeper than 230 m below present sea level. The 40-50 m thick stratified sediments that overly the till in the fiord suggest that there were significant amounts of englacial and supraglacial debris in the glacial ice that flowed through the fiord basins (Fig. 4-10).

There is no definitive evidence to assume a correlation between the till unit in Nachvak Fiord and the upper till unit on the Labrador Shelf. Both Fillon and Harmes (1982) and Josenhans *et al.* (1986) use radiocarbon age determinations on shells and total organic matter to support a Late Wisconsinan age for the upper till unit on the shelf, although virtually none of the samples were taken from the till unit itself, rather from overlying sediments. Those taken from the till were from the top 50 cm of sediment. Furthermore, total organic dates are regarded as locally unreliable due to contamination by reworked terrestrial carbon (Fillon *et al.*, 1981*a*). Vilks and Mudie (1978) proposed early deglaciation of the southeastern Labrador Shelf based on foraminiferal and pollen records from a cored mud sequence. However, their proposed date of deglaciation at $\geq$ 22 ka BP was also based on a radiocarbon date from total organic matter in the sediment. Josenhans *et al.* (1986) report 16 dates within the range 20-26 ka BP for sediments deposited on, or after deglaciation.

It is evident that a large volume of data has been compiled on Labrador Shelf sediments. The interpretation of these data as evidence for an extensive last glaciation of the shelf is undisputed and closely correlates with interpreted glacial histories of the continental shelf to the north and to the south. However the contention that this extensive glacial event occurred during

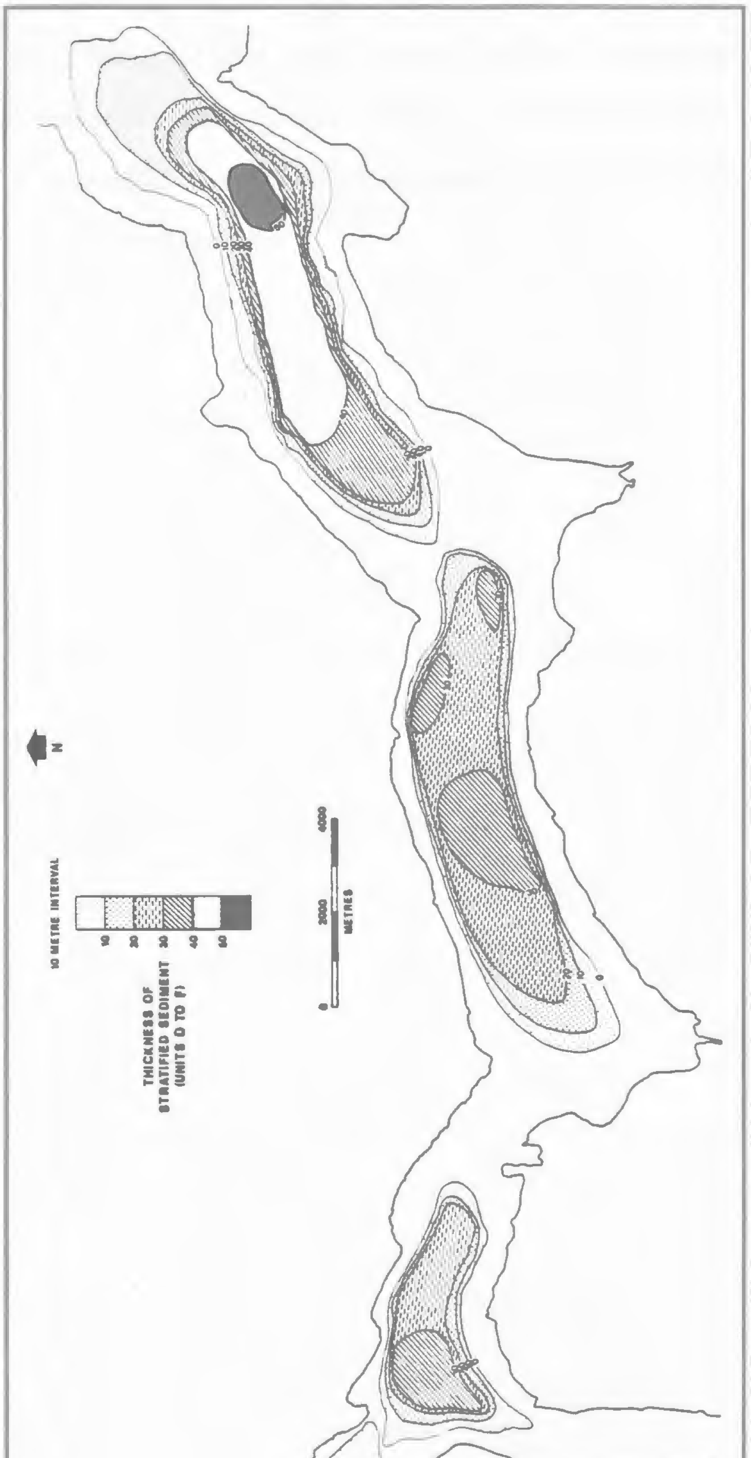

Figure 4-10: Isopach map of stratified sediments [Units D to F] in Nachvak Fiord.

the Late Wisconsin is debatable and conflicts with the evidence from adjacent shelf sediments. A possible solution to this problem may be achieved by accurate tracing of the shelf sediment units into the fiords and bays of the Labrador coast, where terrestrial evidence of glaciation may be integrated with the acoustic stratigraphy to produce a relative or absolute framework of glacial events[15]. However the success of this approach depends on unequivocal land-based chronologies and good control of land/sea correlations.

[15] Clark and Josenhans (1986) propose that acoustic units have been traced from the Labrador Shelf into the fiord basins, but no evidence to support this claim is provided. As Josenhans *et al.* (1986) suggest, the shallow, inner shelf area with its thin mantle of sediments hampers such lateral tracing of sediment units.

4.4. Adams Lake

In the summer of 1985 a 3.5 kHz acoustic survey and coring program[16] was carried out on selected lakes in the Torngat Mountains of northern Labrador. Adams Lake, situated 6 km to the south of Nachvak Fiord, was chosen because of its coastal location and position adjacent to the fiord, and because recent research on the glacial geology of the area provides a perspective to the interpretation of lake sediment stratigraphy. Results of the Adams Lake survey were published by Bell *et al.* (1987) and are reproduced in this section.

Adams Lake is situated approximately 6 km south of Nachvak Fiord. It is a narrow linear body of water that is 36 m above sea level, 3.9 km in length, and is oriented in an east-west direction. It varies in width from 100 m to 400 m and is bounded by steep valley walls (400-500 m high) along most of its central portion. The lake drainage basin comprises the sparsely vegetated slopes of the valley sides and a poorly defined part of the southeastern margin of Tinutyarvik valley. Active erosion of the upper slopes by surface runoff combined with downslope mass movement of ice marginal deposits contributes material to the lower slopes and lake margins. Alluvial fans occur along the lower slopes, grading down from marine limit into the lake (Fig. 4-11). The fans are virtually inactive at present, and overall, rates of sediment input to the basin appear to be low. However the fans may be a relatively important sediment source through erosion during high discharge events.

No significant inflowing streams were observed during the summer and there is only one stream outlet from the lake, located in the south-east corner. Groundwater may play a large role in lake recharge and discharge with recharge augmented by snowmelt runoff in the spring.

Adams Lake basin is situated within a sedimentary series of the Lower Proterozoic Ramah Group (Figs. 2-4 & 3-15). The lake threshold (eastern margin) is located on a fault which marks the unconformity between the lowest Quartzite Member of the Rowsell Harbour Formation (Ramah Group) and the local Archaean gneisses. Six hundred metres to the west of Adams Lake is the Nachvak Fiord Thrust which delimits the western extent of the Ramah Group. It is not known if the lake basin is directly underlain by Proterozoic sediments and/or Archaean gneisses.

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[16] This program was carried out in co-operation with Dr. R.A. Klassen, Terrain Sciences Division, Geological Survey of Canada, Ottawa, as a contribution to Canada-Newfoundland Mineral Development Agreement 1984-1989.

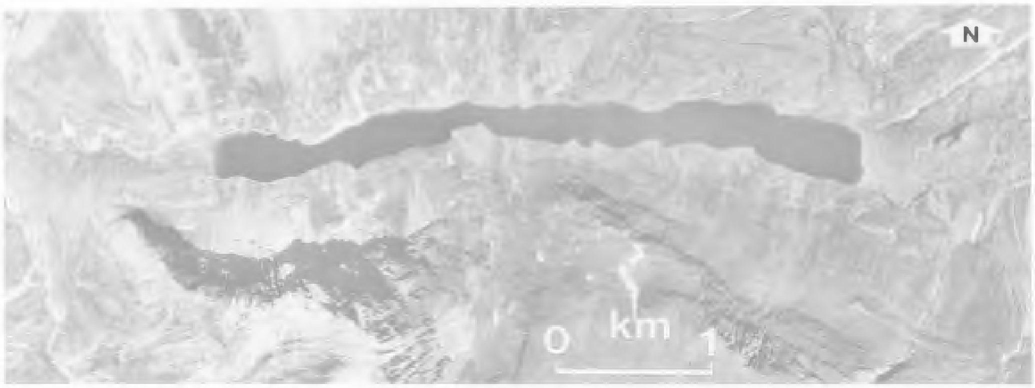

Figure 4-11: Air photograph of Adams Lake and the southeastern margin of Tinutyarvik valley.

Within the lower formations of the Ramah Group are diabase sills which outcrop normal to the east-west orientation of the lake basin.

Adams Lake is characterized by a series of basins separated by relatively shallow barriers (Figs. 4-12 & 4-13). The basins, denoted X, Y and Z, are progressively deeper towards the east, attaining maximum depths of 12 m, 19 m and 21 m, respectively. Basin *Z* contains a smaller basin along its eastern margin, which will not be treated as a separate basin in this analysis. The Barriers *U*, *V* and *W*, separating the basins, are distinct features between 3 m and 6 m below lake level. Barriers *V* and *W* are separated by another small basin, but in the context of this study, the Barrier Complex *VW* will be regarded as a single barrier except in discussion of its origin and significance. The lake basin has an asymmetric 'U'-shaped cross section with steeper slopes on its southern margin and gentle or stepped slopes along its northern edge (Fig. 4-14).

4.4.1. Subbottom sediment units

The Adams Lake acoustic survey was carried out from a rubber Zodiac using a Raytheon RTT 1000 system, with 3.5 kHz and 200 kHz acoustic signals. A 12 volt D.C. battery was used as an energy source for the equipment. The Zodiac was maintained at a constant speed and direction within the constraints of field conditions. Orientation and location of acoustic firing lines were recorded using a magnetic compass and through identification of specific shoreline features. Accuracy of the portrayed lines in Figure 4-12 is suggested to be ±25 m. Following

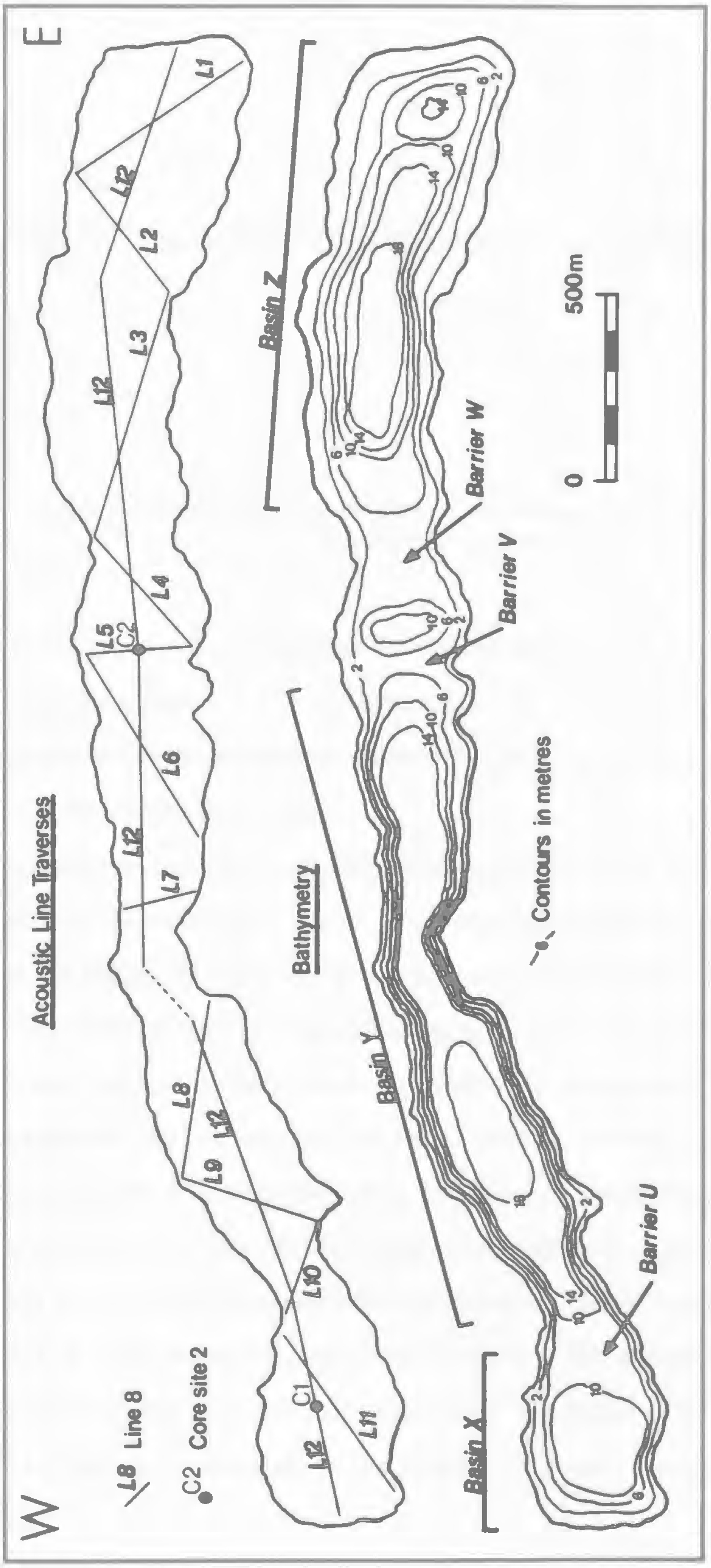

Figure 4-12: Adams Lake bathymetry with location of acoustic line traverses and core sites.

Figure 4-13 is available as two sections in the wallet at the back of this thesis.

Figure 4-13: 3.5 kHz acoustic profile of Line 12 in Adams Lake, with interpreted line drawing of acoustic units.

examination of the acoustic record, core sites were selected and marked using floats. Three major units are recognized based on their 3.5 kHz acoustic signature. These units, which may represent facies, are further subdivided based on their acoustically-defined properties and stratigraphical relationships within the lake basin (Fig. 4-13).

Unit A[17], the lowermost unit, is acoustically massive, without internal structure. Penetration of this unit in the profiles is limited. The surface morphology is variable and irregular, and defines the basins in which the overlying units are recorded. Unit A is interpreted as either bedrock or till. Both media are considered likely to have an upper surface that could act as a strong reflector, and because of the limited penetration by the acoustic signals, no attempt is made here to distinguish between bedrock and till on the record. During seismic reflection studies of sediments under Lake Superior, Landmesser *et al.* (1982) reported problems in distinguishing basal till from bedrock, especially when the till was present only as a thin veneer. Diabase sills in the vicinity of Barriers *V* and *W* which are mainly composed of Unit A, would suggest that this unit represents bedrock in those locations (Fig. 3-15). Barrier *U* has a pronounced asymmetric shape and steeper distal (eastern) slope, characteristic of an end moraine feature. The position of this feature on the lake bottom corresponds to the location of an end moraine (Tinutyarvik

[17] The acoustic units described here are assigned an alphabetical notation. Coincident notation for fiord and lake acoustic units does not imply any similarity in acoustic or geologic properties.

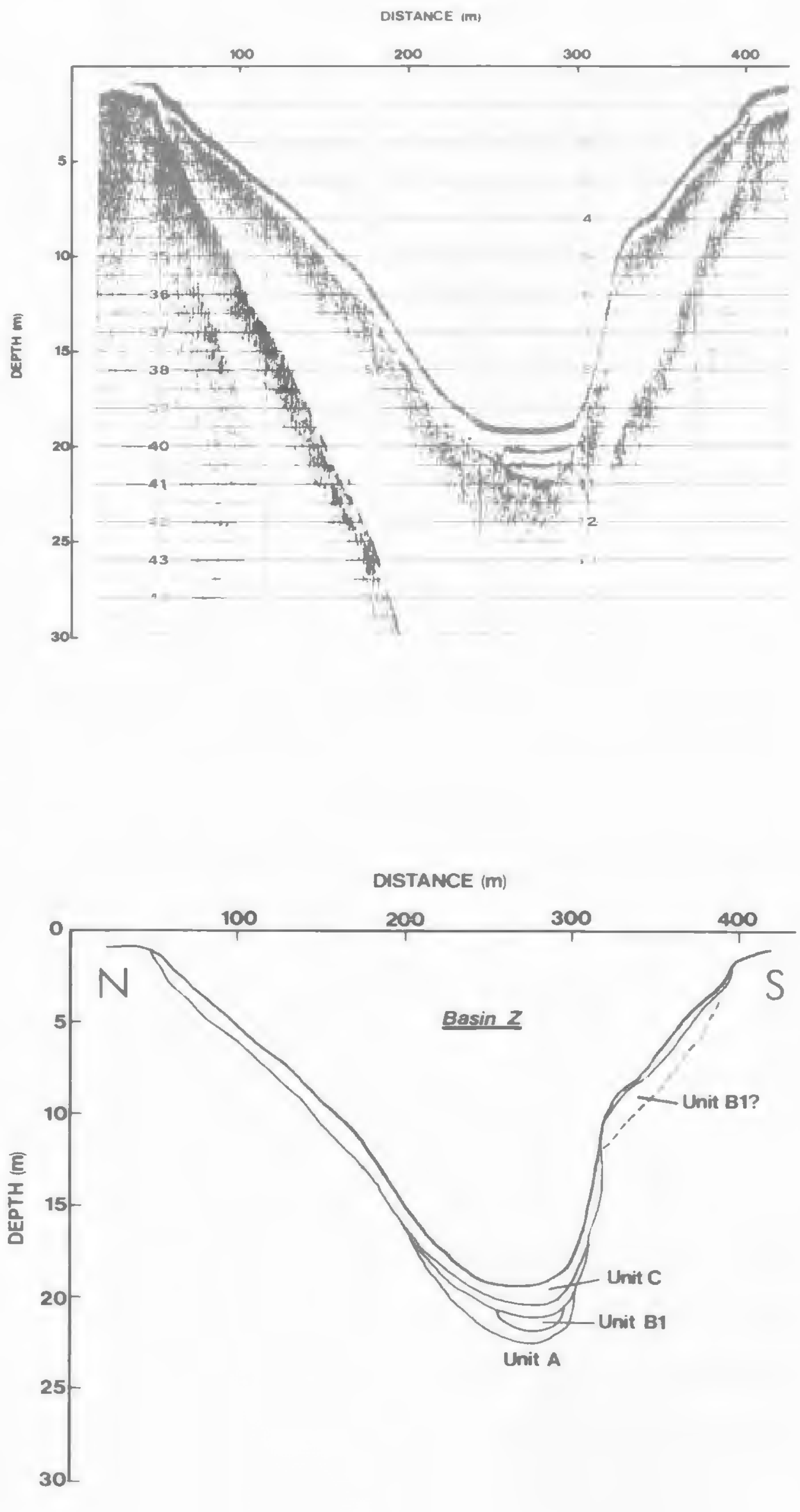

Figure 4-14: 3.5 kHz acoustic profile of Line 2 in Adams Lake, with interpreted line drawing of acoustic units.

moraine) observed in Adams Lake valley (Fig. 4-15). It is suggested that these two features are geomorphically associated and that Unit A comprises morainic drift/till in this locality.

Figure 4-15: Photograph looking west towards Tinutyarvik moraine [A]. The chart recorder shows acoustic profiles for Line 8 and beginning of Line 9. The profile displays extensive Unit B deposits [>10 m], immediately east of Barrier *U*.

Unit B is an acoustically stratified unit which is divided into subunits based on acoustic stratigraphical relationships within the lake basin. Subunit B1 occurs in Basins *X*, *Y* and *Z*. It may be classified as a basin-fill deposit, showing an onlap relationship with underlying Unit A in Basins *Y* and *Z*, but having a poorly defined lower acoustic boundary in Basin *X*. Stratification in subunit B1 varies from well-defined in Basins *X* and *Z* and eastern Basin *Y* to discontinuous with internal deformation in western Basin *Y*. This subunit has infilled a large basinal depression in the centre of Basin *Y* and is at least 10 m thick in this locality; elsewhere it is approximately 2-3 m thick. The upper boundary of subunit B1 is marked by a strong reflector, subunit B2, in Basins *X* and *Y* and is the uppermost internal reflector in Basin *Z*.

Subunit B2 is approximately 0.5-1.0 m thick (Fig. 4-16) and occurs in the bottom of Basins *X* and *Y*, pinching out along the lower basin slopes. It is a strong well-defined reflector which suggests an abrupt change between the composition of this and surrounding units. The strength of the reflection signal varies but the subunit generally appears conformable with the stratification and upper surface of the underlying subunit B1.

Subunit B3 also occurs in Basins *X* and *Y* and conformably overlies subunit B2 (Fig. 4-16). Its upper boundary is marked by the uppermost reflector in both basins. It is predominantly an acoustically stratified unit although towards the upper boundary the unit becomes more

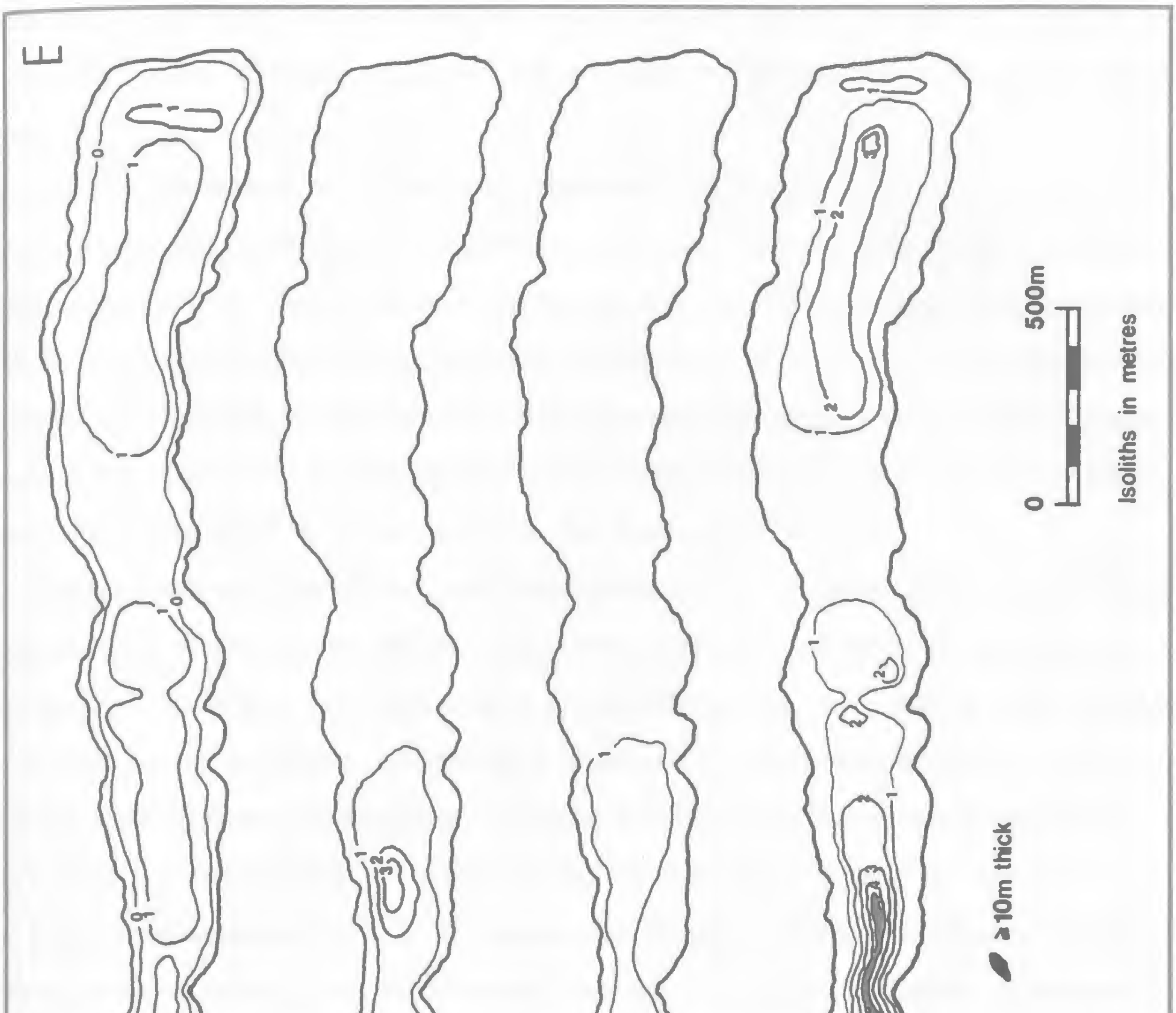

Figure 4-16: Isopach map of subunits B1, B2 and B3, and Unit C in Adams lake.

acoustically transparent. For the most part it is a basin-fill deposit pinching out along the lower slopes of the basins. The unit varies from 0.5 m to 3 m in thickness and is thickest in the centre of Basin *Y*.

Unit B is interpreted as a proglacial sedimentary sequence, laid down in the semi-isolated basins of Adams Lake. Thickest accumulations of Unit B sediments occur in the western basins suggesting that density currents, derived from local glacial drainage systems in the west, settled out in Basins *X* and *Y*, Barrier complex *VW* preventing large scale sediment transport to Basin *Z*. The semi-isolated nature of the basins as 'settling ponds' obscures the probable diachronous deposition within this unit, particularly subunit B1 in Basin *X* which may only have formed following the recession of the ice margin from the Tinutyarvik moraine.

The existence of a former proglacial lake dammed by Quartzite moraine is suggested by lake shorelines south of the moraine, and by a large alluvial fan which appears to originate from a deep gulley in the moraine (Fig. 3-3). Drainage of this glacial lake and downcutting of the moraine may be recorded by the strong acoustic signal of subunit B2 which may be the record of a significant input of coarse clastic debris. Faulting of Unit B stratified sequence near to the eastern margin of Basin *Y* (Fig. 4-13) may have resulted from the meltout of buried ice.

Unit C, the uppermost unit in the acoustic record, appears as a sheet-drape over the sediment package beneath. It is a transparent to semi-transparent unit which can contain single well-defined internal reflectors. It is recorded in all basins and on the tops of barriers and has a uniform thickness of approximately 1.5 m (Fig. 4-16). Unit C is interpreted as a relatively homogeneous fine-grained sedimentary unit deposited postglacially, under low sediment influx rates, in a lacustrine environment.

4.4.2. Summary

The interpretation of acoustic units in terms of facies and depositional environments is tentative and will be discussed in Chapter 7 in conjunction with terrestrial evidence of glacial and deglacial events in the area. However, the above acoustic stratigraphy can be summarized by stating that the three major acoustic units identified from the 3.5 kHz record may be interpreted as representing sedimentary facies related to a glacial/deglacial cycle. Unit A, the lowermost recognisable unit, is characterized by its predominantly massive structure and reflector-free image.

This unit composes the most westerly barrier between Basins *X* and *Y* where it is suggested to be a till-like unit; elsewhere it defines the shape of basins and may be drift or bedrock. Unit B is predominantly an acoustically stratified unit and is interpreted to be a proglacial deposit ranging in depositional environment, from ice-proximal to ice-distal, towards the east. Subunit B2, interpreted to be a coarse clastic deposit, is tentatively related to the draining of a nearby moraine-dammed lake. Unit C, an acoustically transparent unit with a uniform thickness of *ca.* 1.5 m, is interpreted as a postglacial lacustrine sediment.

Barrier *U*, the most westerly barrier, has a characteristic end-moraine morphology and is thought to be the subaqueous extension of the Tinutyarvik moraine which cuts across the lake at the same location. Barriers *V* and *W* may be either bedrock, or moraines formed during an earlier phase when ice was more extensive. The Adams Lake moraine at the eastern end of Adams Lake is evidence of a further eastward extent of ice that predates the Tinutyarvik moraine.

4.5. Conclusions

In this chapter an attempt has been made to supplement the terrestrial glacial geology (Chapter 3) with a tentative description of subaqueous sediments. The acoustic stratigraphies recorded in Nachvak Fiord and Adams Lake have been interpreted according to acoustic facies, and tentative models of deposition are proposed. The preliminary nature of these analyses is emphasized, and the detailed examination of cores from surface sediments is necessary, if the models should have a solid framework with some degree of control and verification. An attempt has been made to assess possible mechanisms for deposition of the proposed facies, and hence some units may appear to have an ambiguous origin. Further analysis may help to eliminate some of the least probable depositional processes.

Discussion of the proposed stratigraphic sequences in Nachvak Fiord and Adams Lake reveals the inherent need for supportive evidence from terrestrial sources. The recognition of a 'moraine-like' feature or a 'till-like' unit in the acoustic record must be compatible with terrestrial observations, and likewise the converse of this statement must hold true. This informal 'rule-of-thumb' restricts the imagination of the observer, but provides important supporting evidence for local glacial chronologies. This will be the basis of the integration of terrestrial and subaqueous data from the study area. Unfortunately, with distance offshore, the importance of land-based observations or chronologies diminishes, resulting in increased speculation and more tentative interpretations. An obvious solution to this problem is to trace offshore units inshore and, where possible, onto land.

Chapter 5

Raised marine shorelines and relative sea level change

5.1. Introduction

This chapter deals with the late-glacial and postglacial sea level history of outer Nachvak Fiord. Geomorphological evidence of higher sea levels is common along the Labrador coast and many examples have been reported during the last century. Even the earliest observers attributed the "sinking and rising of the land" to the Pleistocene ice sheets (Delabarre, 1902). This relationship between the elevation and pattern of emerged shorelines and the decay of the last ice sheets is a particular concern of this research.

Although many geologists have commented on the raised marine features of the northern Labrador coast, there has been a marked absence of detailed research on relative sea level change in the region. Löken (1962*b*) made an important contribution with his paper entitled *"The late-glacial and postglacial emergence and the deglaciation of northernmost Labrador"*, but apart from this, researchers working in northern Labrador have had to rely for published references upon extrapolated interpretations from Baffin Island or large scale models designed for Eastern Canada.

This study is confined to a small area where detailed recording of the raised marine features has provided a comprehensive data set for shoreline correlation and presentation. The extent and geometry of these raised shorelines are analysed and a glacio-isostatic model is proposed for the study area. The purpose of this model is to serve as a test of old and new theories of deglaciation and relative sea level change in northern Labrador. Although the model is consistent within the study area, it requires accurate regional control from adjacent areas which may be the subject of future research.

5.2. Shoreline Altitude Measurements: Assumptions, Procedures and Errors

An integral part of Quaternary palaeoenvironmental reconstruction concerns sea level change resulting from the effects of glaciation. The basic data from which sea level movements are described consist of marine shoreline altitudes. Despite the importance of this measured component, many papers concerned with sea level change make little reference to the techniques and equipment used in data collection, nor to the error margins associated with the results obtained. Andrews (1970) and Gray (1983) outlined the substantial differences in techniques employed for various study areas and emphasized the importance of standard procedure in measurement of shoreline altitudes and calculation of sea level movements.

Research on relative sea level change in northern Labrador has been carried out since the end of the last century (Bell, 1885; Low, 1896; Daly, 1902); yet observations remain patchy, grouped in localized study areas with little agreement between results (Wenner, 1947 *in* Johnson, 1985). Whether discrepancies are the result of various methodologies is impossible to prove, but the requirement for systematic and detailed observation is clear. This study attempts to describe and quantify measurement error, specific to research carried out in Nachvak Fiord, based upon the example of similar research undertaken in Scotland (for example, Gray, 1983).

5.2.1. Datum Lines

Previous studies in the Canadian Arctic have used a variety of methods to compensate for the absence of bench marks tied into the national datum line. These alternatives, termed 'water level' or 'surrogate water level' methods by Gray (1983) are an attempt to provide consistency in altitude measurements within a field area and to allow reasonable comparison of data between field areas.

'Water level' methods assume a horizontal water surface with negligible fluctuation of the water plane over the measurement period. This method is most reliable when applied to lake levels, (for example, Ives (1960, p.54) Indian House Lake, Labrador-Ungava peninsula), although the sea surface has also been used in Scotland (Gray, 1983). The occurrence of a small tidal range in order to reduce errors due to sea level fluctuations is desirable, but not essential if sea level is corrected to national datum at the time of observation using tide tables. However, as Johnson

(1985) has pointed out, with particular reference to northern Labrador, the lack of tidal observations along the coastline, and variations in the timing and amplitude of local tides in large irregularly-shaped fiords introduce sources of error in this particular method.

Archer (1968) argues that the use of mean sea level as measured by the mid-point between the local high and low tides based on a number of observations, is a reasonably accurate datum throughout a study area. Johnson (1985) suggests that the determination of mean sea level datum would require a minimum of at least one lunar month's observations to be applicable even on a local basis.

The use of biological surrogates as an indicator of sea level has been tested by Sollid *et al.* (1973) and Sutherland (1981) and according to Gray (1983) this method has considerable potential in remote areas. The tests were carried out using the upper growth limit of *Fucus vesiculosus* and barnacles, respectively. This method has been employed not only on living species but also on the upper limit of sea drifted materials, especially seaweed. Synge and Stephens (1966) used a datum based on freshly drifted sea salad (Enteromorpha sp.), whereas Mathews (1967) and Dyke (1979*a*) measured the upper limit of all drifted material, or flotsam, as indicative of the upper limit of wave action.

This study employed a combination of both the 'water level' and 'surrogate water level' methods in the measurement of strandline elevations. Considering the cautionary remarks of Johnson (1985) with respect to the paucity of tidal observations along the Labrador coast, and acknowledging the irregular fiord coastline with its deeply indented bays and bights, it was decided to employ the upper limit of drifted material as an indication of the high water mark in the individual coves. In some cases it was possible to compare this datum level with the flat under surface of shore-fast ice. The latter was a non-biological surrogate used by Blake (1970, 1975) as a reliable indicator of the level of the highest tides. Both indicators concurred, although Blake's method was never used as the only approximate of the highest tide level.

Once the high water plane was identified, a bench mark was set up nearby and subsequently used as the opening and closing point of traverses. This approach eliminated the possibility of identifing a higher tide level constructed following the original observation. These bench marks are labelled on Figure 5-1 and their heights above the high tide datum plane are given in Table 5-1.

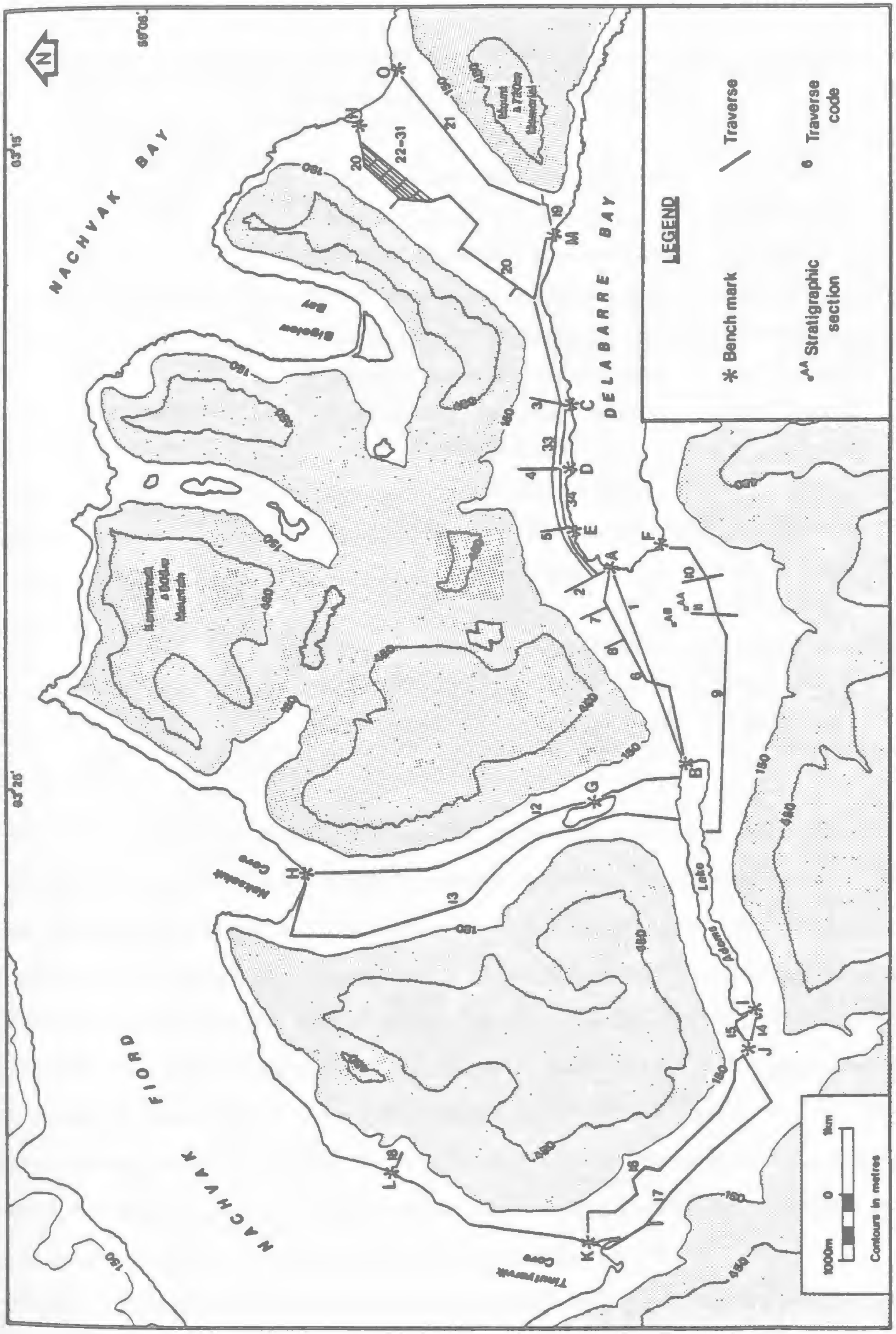

Figure 5-1: Location of bench marks and levelling traverses in the study area.

Table 5-1: Location and elevation of bench marks constructed in the study area.

Location and Elevation of Bench marks		
Bench mark	Location	Elevation
A	Northwest shore of Delabarre Bay	00.250 m
B	Eastern shore of Adams Lake	36.613 m
C	Northern shore of Delabarre Bay	02.043 m
D	Northern shore of Delabarre Bay	00.622 m
E	Northern shore of Delabarre Bay	00.670 m
F	Southwest shore of Delabarre Bay	00.200 m
G	Eastern shore of Gurnot Lake	79.774 m
H	Naksaluk Cove	02.721 m
I	Western shore of Adams Lake	36.145 m
J	Rock knob in western Adams Lake valley	52.685 m
K	Tinutyarvik Cove	01.249 m
L	Shrule Cove	01.889 m
M	Valley of the Flies southern shore	01.651 m
N	Valley of the Flies northern shore	02.316 m
O	Valley of the Flies northeast shore	00.000 m

The direct 'water level' method was used to transfer the altitude of *Bench mark B* along the shore of Adams Lake to *Bench mark I* at its western extremity, and to open and close traverses at points along the lake shoreline. The time taken to transfer altitudes was never longer than thirty minutes and therefore any error generated through the use of this method was negligible.

The high tide levels which actually occur along the coast are the result of predictable tidal oscillations together with the effects of meteorological conditions. Strong or prolonged winds, abrupt changes in barometric pressure, or prolonged periods of very high or very low pressures introduce fluctuations in the water level which invalidate the assumption that the high tide level is a horizontal plane. Unless all observations are made during a stable meteorological period a source of error is introduced into the calculation of shoreline altitudes.

Adams Lake is a long narrow finger lake oriented in an east-west direction along the connecting valley between Tinutyarvik valley and Delabarre Bay (Figs. 4-11 & 5-1). The use of lake level as a temporary datum plane facilitated the checking of surrogate high tide levels in

Tinutyarvik Cove and Naksaluk Cove, *Bench marks K* and *H* respectively, against the initial high tide observation in Delabarre Bay (*Bench mark A*). The time elapsed between the construction of *Bench marks K* and *H* and *Bench mark A* was 21 days and 16 days respectively. It was not logistically possible to construct all bench marks within a reasonable time period coinciding with stable meteorological conditions, and therefore errors, due to either a higher tidal regime or a changing meteorological factor influencing tidal magnitude, may exist in the data. Calculation of the error[18] due to either of the above circumstances gave values of 0.094 m and 0.093 m for *Bench marks K* and *H*, respectively[19]. The consistency in error may be coincidental or it may suggest the influence of similar factors (as above) affecting tidal conditions at *Bench marks K* and *H* subsequent to the initial observation of high tide level at *Bench mark A*.

Levelling data provide an opportunity to test the consistency of the latter results. Fourteen days following the measurement of *Bench mark A*, a second bench mark was constructed on Delabarre Bay, *Bench mark F*, which was levelled from the high tide line; that is, the upper limit of drifted material. On this occasion the traverse was closed at *Bench mark B*. The datum plane error between *Bench marks A* and *F* can be calculated by subtracting the combined closing errors of the traverses from *Bench marks F* to *B* to *F*[20] and *Bench marks B* to *A* to *B* [0.173 m] from the closing error of the simulated traverse from the datum planes at *Bench marks F* to *A* to *F* [0.256 m]. This datum plane error was calculated to be 0.083 m, and is significantly close to the datum plane errors at *Bench marks K* and *H* to support the consistency in these results and to suggest that there was a real change in the elevation of the datum plane by approximately 0.1 m since the construction of *Bench mark A*.

It is not possible to determine which factor, either tidal oscillation or the influence of meteorological conditions, is responsible for the elevated high tide datum plane, as there is no

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[18] This error value was calculated by subtracting the sum of the closing errors of each segment of the traverses, via *Bench mark B*, from the cumulative closing error produced if a direct traverse was conducted from the datum plane at *Bench mark A* to either datum planes at *Bench mark K* or *Bench mark H*. For example, the datum plane error at *Bench mark K* = the closing error of the traverse from the datum planes at *Bench marks A* to *K* to *A* - [the closing error of the traverse from *Bench marks A* to *B* to *A* + the closing error of the traverse from *Bench marks B* to *K* to *B*].

[19] Although these values are relatively small considering the mean length and mean elevation of the traverses concerned, 4255 m and 49.5 m respectively, they are significant in comparison to the mean closing error of 0.14 m for these traverses.

[20] The error associated with the traverse segment from *Bench marks B* to *F* is estimated to be 0.64 m which is equal to the closing error of the traverse from *Bench marks F* to *B*.

documentation on the circumstances under which the earlier level was formed. However, several conclusions can be made concerning these results. First, since all bench marks constructed for this study were recorded within a twenty-two day period, it is probable that 0.093 m is the maximum error due to variations in datum plane. Second, the difference in datum plane error between coves in the fiord and and those in Delabarre Bay would suggest that the irregular character of the fiord coastline does not significantly affect the magnitude of tidal range, at least within the boundaries of the field area. Finally, these results verify the potential use of surrogate water levels as datum planes in the absence of offical bench marks levelled from the national datum line.

5.2.2. Instruments

A combination of altimeter and surveyor's level was used in the measurement of shoreline altitudes. In 1984 only the altimeter was employed for such measurements, and it was concluded that the error margin was probably too great for any detailed reconstruction of sea levels in the field area. Sparks (1953) discussed the merits of the aneroid barometer for accurate field work. He concluded that observations should be taken during stable atmospheric pressure periods while avoiding strong winds (>12 mph), the passage of fronts, unstable air conditions and irregularly developing pressure systems. He suggested that if a minimum of ±1.5 m accuracy is required then a traverse should be no longer than one hour in duration, assuming suitable conditions. During 1948-1949, Sparks (1953) calculated that only 33% of summer days in north-west Scotland were suitable for this level of error margin. Consequently, as Gray (1983) concludes, "this method must trade accuracy against field time."

The pressures on time and expense have, in many studies in Arctic Canada, eliminated the possibility of using a more accurate instrument than the altimeter/aneroid barometer. Consequently researchers have concentrated on improving the accuracy of data recorded with this instrument; for example, correction procedures based on pressure and/or temperature changes (Mathews, 1967; Archer, 1968), the employment of a second instrument located at the opening and closing point of a traverse, or frequent rechecking (every 30 minutes) of the instrument at a base point (Moller and Sollid, 1972 [Norway]; Blake, 1975). To facilitate these corrections, the traverse must be closed and the closing error distributed along the traverse. A further suggestion, as a result of this study, relates to the instrument operator. Three people were asked to record measurements, using the same technique, at the same position on a landform and the resultant readings varied by two metres. To retain consistency, therefore, only one operator should record measurements along a single traverse.

A pilot study on the accuracy of altimeter measurements was carried out during the 1985 field season, using a 'Wallace & Tiernan' altimeter. The main objective was to test measurement accuracy against time taken to complete a traverse. Table 5-2 shows the results of this test. The altimeter traverses were opened and closed at the same point and the resultant error distributed throughout the traverse. Measured elevations at each station were then compared to the equivalent levelled data and a mean difference for all stations calculated. This value is also expressed as a percentage of the maximum altitude measured. The percentage error of 1-2% calculated for this study compares closely with error values cited for similar research from Arctic Canada [e.g. Löken, 1962*b* (Labrador, 1.3%); Archer, 1968 (Hudson Bay, 1-3%)].

Table 5-2: Results of a study to test measurement accuracy against time taken to complete a traverse.

Measurement accuracy versus traverse duration					
	T18	T18	T19	T4/5	T13
	0.5 hr	1.0 hr	3.0 hr	4.5 hr	6.9 hr
$\sum$ difference	04.073 m	04.345 m	11.590 m	20.631 m	37.009 m
Mean difference by station	0.679 m	0.869 m	0.966 m	1.289 m	1.234 m
Mean difference/max. elevation	1.2%	1.5%	2.1%	1.9%	1.5%

If duration of traverse is plotted against mean error, a direct relationship is found (Fig. 5-2). However, the rate of error increase during the first hour (77 cm/hr) is much greater than the rate of 6 cm/hr recorded for the subsequent six hours. If error is expressed as a percentage of the maximum elevation recorded, and these values are graphed with duration of traverse (Fig. 5-2), then for the sample traverses plotted, the trend shows a decrease in percentage error after a minimum time period of three hours.

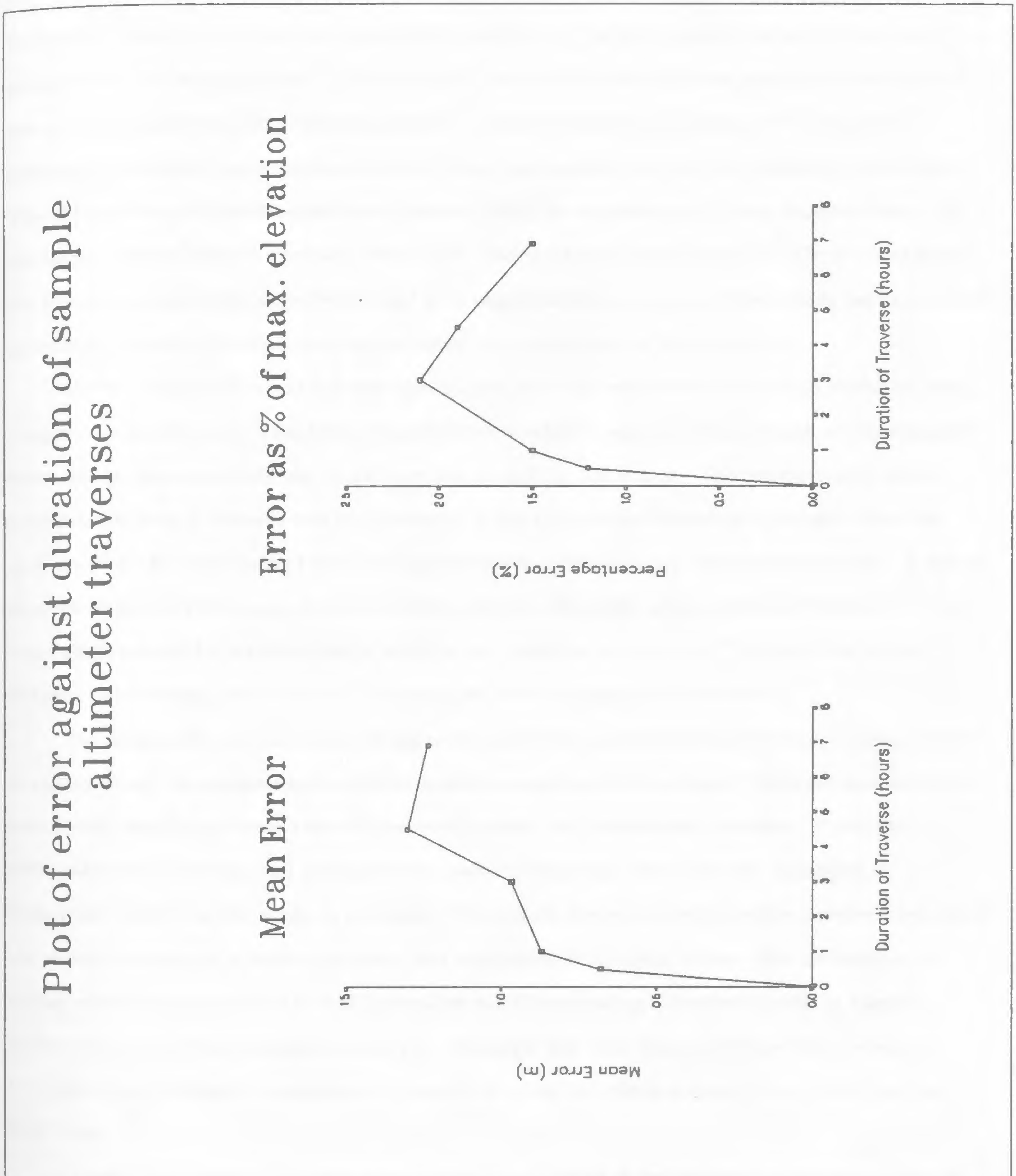

Figure 5-2: Plot of error values against duration of randomly selected altimeter traverses. Error values are expressed as a mean value and as a % of the maximum elevation recorded.

Both these graphs contradict the expected trend of error over time as suggested in the literature. Sparks (1953) and Johnson (1985) specifically caution against the use of traverses greater than one hour duration. The traverses used in this analysis were randomly selected and are not necessarily best-fit profiles in relation to their levelled equivalents. The data were acquired at different times during the field season and under various meteorological conditions. The applicability of Sparks' guidelines (Sparks, 1953) for altimeter use is not disputed here, only the assumption of limited accuracy once these conditions have been satisfied. These conclusions are based only on preliminary results and it is suggested that a more detailed study be carried out in order to present definitive conclusions based on a significantly larger data set.

Marine limits and raised beaches throughout the field area were surveyed in detail in 1985, using a surveyor's level. This type of instrument is reliable and accurate, though errors can still occur due to improper field use or unfavorable operating conditions. The magnitude of error accumulated over a traverse can be calculated if the same point is used to open and close the traverse or if the elevations of the opening and closing points of the traverse are known. A list of these closing errors for major traverses undertaken in the study area is given in Table 5-3. The mean closing error for all traverses is 0.184 m per traverse, or 7.8×10^{-5} m error per metre distance per traverse, or 3.72×10^{-3} m error per metre altitude per traverse.

The magnitude of this mean closing error would be considered large by Gray (1983) [>0.15 m closing error], in comparison to similar studies carried out in Scotland. However several factors may justify this larger error value for surveys carried out in northern Labrador. First, the collimation error in using the instrument is considered greater than 0.011 m suggested by Sutherland (1981) for his study in Scotland. The rough bouldery terrain makes precise levelling of the actual instrument a time-consuming and sometimes frustrating affair. The persistence of strong winds and generally unsuitable weather conditions during a traverse result in larger collimation errors than normally expected. Although this error can be reduced by keeping backsights and foresights approximately equal, on some occasions accuracy was traded against field time.

Second, it is assumed that there is a direct relationship between length of traverse and the size of the closing error. Some traverses within the study area were approximately 15 km in length and therefore an above-average error was expected. Although this was the case for several

Table 5-3: Information on major levelling traverses carried out in the study area.

Traverse closing errors and gradient indexes				
Traverse	Distance (m)	Max. elevation	Error (m)	Index (m/m)
1	3162.3	37.352	0.023	0.012
2	0726.5	70.585	0.099	0.097
3	0578.6	52.042	0.265	0.090
4	0872.3	68.110	0.379	0.078
5	0615.9	69.828	0.139	0.113
6	3538.8	37.868	0.258	0.011
7	0455.1	43.672	0.422	0.096
8	0302.8	34.373	0.031	0.114
9	5445.3	58.292	0.064	0.011
10	0620.5	52.126	0.661	0.084
11	0786.1	47.753	0.062	0.061
12	6082.9	86.248	0.002	0.014
13	8044.8	87.325	0.002	0.011
14	0925.0	16.540	0.247	0.018
15	0826.6	17.035	0.247	0.021
16	4866.8	57.020	0.297	0.012
17	4891.5	55.531	0.297	0.011
18	0261.3	57.772	0.081	0.221
19	1015.9	45.636	0.036	0.045
20	6415.2	67.733	0.359	0.011
21	3103.1	28.355	0.180	0.009
22	0409.2	25.543	0.033	0.062
23	0322.2	22.499	0.055	0.070
Total	54268.7 m	1139.24 m	4.239 m	—

traverses (for example, closing errors = 0.258 m [T6], 0.297 m [T16], 0.359 m [T20]), the relationship did not hold true for all traverses. Figure 5-3 shows a plot of error value against a gradient index for each traverse. The index value was calculated by dividing the maximum altitude recorded, by the distance covered in the traverse. [This index value is only an estimate of the traverse gradient as the traverse distance is not a horizontal straight-line measurement.] Therefore, in this instance, a low index value indicates a long traverse, whereas a high index value represents a traverse over a relatively small distance, recording a large altitudinal range (See

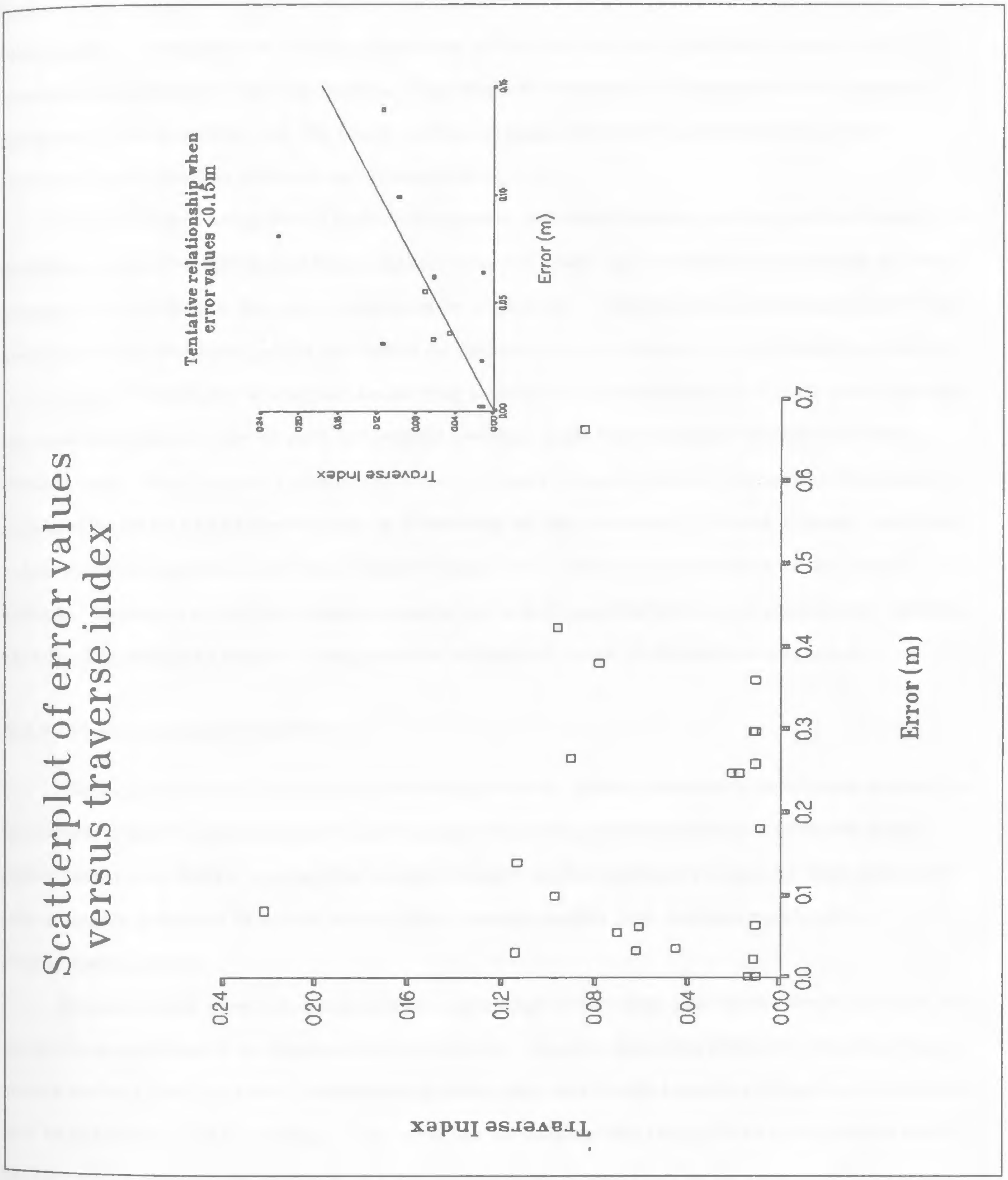

Figure 5-3: Scatterplot of error values against traverse index values. A tentative relationship between index value and error is described by a regression line for data points with error values <0.15 m.

Table 5-3 for comparison of values). Those traverses with steep gradients, >0.03 index value, tend to have a larger magnitude of error proportional to traverse gradient. This observed relationship is probably due to high collimation errors; that is, conditions not conducive to accurate employment of the instrument. The frequent occurrence of steep foreshores and back edges on raised shorelines and the rough surface of many of the steeper slopes along these traverses make levelling difficult and susceptible to error.

Finally, these closing errors include the datum line error discussed earlier in this chapter. A maximum error of 0.093 m has been suggested for this study and is almost a magnitude of error greater than the datum line error calculated by Sutherland (1981) for the Loch Long/Loch Fyne area of western Scotland [±0.01 m - based on Ordance Survey figures for bench mark accuracy].

It was decided not to correct the levelled altitudes by distributing the closing error through the traverse cumulatively at each instrument position; a practice normally carried out with levelled data. As the exact location and source of error along a traverse was not known, there appeared to be no obvious advantage in blanketing all the measurements with a mean correction value; thereby suggesting that the resultant figure was a closer approximation of the actual altitude. Instead, the levelled measurement is given and qualified with an averaged error margin thereby informing the reader of the potential altitudinal range of the surface measured.

5.2.3. Measurement Procedure

The largest source of error in the levelling of raised marine features is attributed to *what* is measured, *where* the measurement is taken and *how many* measurements are recorded (Gray, 1983; Sutherland, 1981). During this study all raised marine features recognized were measured, which mostly consisted of cobble beach ridges, marine-eroded drift terraces and boulder limits/washing limits.

Measurements were taken towards the back edge of features, near to the break in slope, but avoiding areas affected by slopewash or solifluction. In some cases the emerged strandline was poorly defined, and the exact relationship between the area levelled and the former sea level could not be determined with certainty. The error due to misidentification of the exact position on the feature which records the evidence of a former sea level is suggested to be ±0.5 m in Labrador (Johnson, 1985), or ±0.3 m in Scotland (Sutherland, 1981).

As regards the question of the number of measurements to be taken on a raised marine feature, this has not necessarily posed a problem in northern Labrador. The fragmented distribution of emerged shorelines and the difficulty associated with the spatial tracing of these features does not generally permit more than one reliable recording to be made on each raised marine feature or segment.

5.2.4. 'Relationship to sea level and tidal regime' error

Once the elevation of a raised marine feature has been recorded and the associated error margin calculated, there exists the problem of relating the altitude of this feature to the sea level it represents and to that part of the tidal regime at which it was formed. Generally, the relationship between modern coastal landforms and present sea level is used as a guideline for the study of relict shorelines. However, it is difficult to estimate how applicable these present conditions are to the past; variations in tides, bathymetry, mean wind speed and prevailing wind direction over time will affect any general comparisons made within the study area. Consequently no local present-day analogues were measured in order to quantify this relationship. Elsewhere however, attempts have been made to relate the altitude of actively forming shoreline landforms to contemporaneous mean sea level and to apply the calculated correction factors to the relict forms (Rose, 1981, p.329). Corrections made to mean sea level by Rose (1978) for present-day beach ridges [-2.0 m], rock platforms [-1.0 m] and drift platforms [-2.0 m], in Norway, provide a general measure for this relationship in northern Labrador. The error associated with this correction factor has been calculated for individual coastal landforms by Sutherland (1981) but as he points out, these figures are only relevant for his study, based on the variability of coastal landforms encountered in his study area [for example, beach ridges ±0.46 m; erosional terraces ±0.23 m]. Andrews (1970, p.79) suggests that a standard error of ±1 m should be applied to the corrected elevation of raised marine features in order to reduce variance and "to gain a set of conformable data". This study will adopt an error margin of ±0.5-1.0 m when converting the elevation of raised marine features to altitude above mean sea level, based on the calculations of Rose (1978) and Sutherland (1981), and on the suggestion of Andrews (1970).

5.2.5. Summary of Errors

Table 5-4 outlines the individual errors and their proposed magnitudes for this study. As instrument error includes the datum error, and because the majority of elevations used in this thesis were measured by surveyor's level, addition of categories (2)b and (3) gives an error margin of ±0.684 m for landform elevations quoted. This figure added to the observational error associated with the point of measurement on the landform (3), plus the error related to the correction factor for individual landform types (4), gives the total error associated with an assigned sea level altitude based on the evidence of raised marine features [total error = ±1.684-2.184 m].

Table 5-4: Summary of error values associated with the measurement and employment of shoreline data in this study.

Summary of errors related to shoreline data measurement and employment	
Error description	Magnitude
(1) Datum error	±0.093 m
(2) Instrument error - Altimeter	[a] ±0.838 m
(2) Instrument error - Surveyor's level	[b] ±0.184 m
(3) Observational error	±0.5 m
(4) 'Relationship to tidal regime' error	±0.5-1.0 m

The information in Table 5-4 does not include errors associated with misinterpretation of evidence nor gross errors accumulated during stages of a traverse; however, it is suggested that the consistent and systematic approach to data acquisition eliminated the possibility of these errors occurring. Variation in landform type (4) and the correlation of raised shoreline fragments are the major sources of error in the reconstruction of relict shorelines (Andrews, 1970; Sutherland, 1981). These errors are expected to be greater around large former ice bodies (Gray, 1983), where rapid rates of relative sea level change result in poorly preserved marine landforms. The fragmented distribution and poor preservation of raised marine features in the outer Nachvak Fiord area, and along the northern Labrador coast in general (Johnson, 1985), may indicate late deglaciation of the area, and a rapid relative sea level change. (Extensive sea-ice cover may also be responsible for reduced wave action and ineffective beach processes.) Consequently, the quality of the raised marine evidence is a major obstacle to the reconstruction of late-glacial and postglacial sea levels in the region. Hence, it is suggested that the error margins quoted in this study, especially those concerned with landform type, are a minimum estimate, based on similar research carried out under more ideal conditions, using superior data sets.

5.3. Shoreline correlation and representation

Under ideal circumstances, the process of correlating shoreline segments that relate to a single sea level stand, would be undertaken in the field through the lateral tracing of marine landforms. However, in northern Labrador, the sparse distribution of shoreline segments prevents more than very limited use of this approach to shoreline correlation. An alternative method that has been widely used in glacio-isostatic studies involves the plotting of shoreline elevation points in a vertical plane running parallel to the direction of the assumed ice centre. Shoreline segments are then resolved into a series of inferred shorelines by visual or statistical means. This method relies upon two assumptions; first, that there is no curvature to shoreline isobases[21] in the area under consideration, and second, that isobases of different shorelines are parallel to one another. If the first assumption is violated then points projected into a single plane will plot in the wrong position relative to each other. The resultant error becomes greater as isobase curvature increases, as shoreline gradient increases and as distance from the projection plane increases. When the second assumption is violated points from any one shoreline will be wrongly plotted relative to other shoreline points. This will result in a misrepresentation of shoreline gradients, the magnitude of error being proportional to the variation in isobase trends.

As this alternative method requires that the vertical plane, into which the points are plotted, is drawn normal to the shoreline isobases, it assumes that shoreline isobase trends are known. Unless such information is available, it may not be possible to determine isobase trends until the segment elevations are correlated on the basis of the shoreline diagram. Gray (1983) reports that, in most Scottish studies, the isobase trend of a well developed shoreline is used as a guide for the trend of other shorelines, and hence the problem is avoided in the research

[21] Definition of an *isobase* is a line joining points of equal postglacial emergence and refers to simultaneously formed features (Andrews, 1970).

procedure. Other studies have rotated the line of projection by increments of 5-15° until a projection plane is found that minimizes the residual variation of points along the inferred shoreline. Consequently the best-fit projection plane can only provide an estimate of shoreline isobase trends.

5.3.1. Correlation of shoreline segments

The method of shoreline segment correlation used in this study is a variation on the approaches outlined above. A marked shoreline, approximately 25 m aht, was recognised in Adams Lake valley and subsequently traced east along the northern shore of Delabarre Bay. Elevation measurements on the shoreline were recorded at the intersection of levelling traverses and selected shoreline segments. The location and elevation of these points were input to a *SURFACE II* trend surface computer program (Sampson, 1978) which produced a first degree trend surface, fitted to the data by means of a least-squares regression analysis. The solution to the linear regression equation was used to project the linear trend surface to all parts of the study area. The decision to accept a shoreline segment as a correlative to the observed shoreline was dependent on the segment elevation range[22] overlapping with the predicted elevation range of the linear surface[23] at that location. The resultant data set for this shoreline, Shoreline D, comprises the initial observations from Adams Lake valley and Delabarre Bay, and the predicted correlatives from the remaining parts of the field area.

The correlation of segments for the next highest shoreline, Shoreline E, was initiated using a data set composed of segment elevations which were observed to be the next highest shoreline segment above Shoreline D segments, on levelling traverses. These data were input to the linear trend surface program and a best-fit linear surface produced. The solution to the first degree linear regression equation was used to extrapolate the linear surface to remaining parts of the study area. Identification of correlatives to Shoreline E segments was conditioned by three

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[22] The elevation of the shoreline segment combined with the error margin associated with the measurement of this elevation, that is, ±0.684 m for this study.

[23] The predicted elevation of the linear surface combined with the standard error of the forecast, $SE_{\hat{y}}$. This latter value is the combination of the following three variance sources: the standard error of the estimate, SE_y; the standard error of the mean of the dependant variable, $SE_{\bar{y}}$; and the standard error of the regression coefficient, SE_b. These values will vary according to the *fit* of the linear trend surface to the shoreline data.

constraints on the data. The first constraint, referred to as the elevation constraint, requires that the elevation range of an observed shoreline segment overlaps with the predicted elevation range of the linear trend surface at that location. The second and third constraints are an application of Cullingford's staircase and continuity constraints which were originally formulated to assist in the identification of shorelines on scatterplots of shoreline segments (Cullingford, 1977). The staircase constraint requires that if segment A occurs above segment B in a staircase, then they cannot be correlated. The continuity constraint requires that segments A and B must be correlated if they are morphologically continuous. If a shoreline segment satisfied these three constraints then it was correlated with Shoreline E and included in the data set. Failure to comply with these constraints at any stage during the analysis would result in the loss of segments from the data set.

This method of shoreline segment correlation was repeated until all possible shorelines were resolved. Some shoreline segments, identified in the field, did not fit the correlation constraints of any shoreline; other segments satisfied the constraints but the resultant data set was too small [<10 data points] for linear trend surface analysis. However, the complete set of shoreline segments is displayed on a scatterplot for later discussion (Fig. 5-4). Results of statistical analysis on the eleven shorelines, A to K, are given in Table 5-5, and the solutions to the linear regression equations are listed in Table 5-6.

Snedecor's *F*-ratio is a test for the significance of the linear correlation coefficient, r_{yx}. For shorelines C to K this coefficient is statistically significant above the 99.9% confidence level. Shorelines A and B have relatively low coefficient of determination values, r^2_{yx}, which suggests that the residual variance in the elevation values $S^2_{(y-\hat{y})}$ around the linear trend surface is similar to the variance (S^2_y) that originally existed in the elevations values. Therefore the explained variance in $Y\,(S^2_{\hat{y}})$ due to the linear regression equation, is small. The low values for the coefficient of determination $(r^2_{yx} = S^2_{\hat{y}}/S^2_y)$ can be explained by the horizontal nature of Shorelines A and B which is almost equivalent to a horizontal line through the mean Y value.

The correlation of shoreline fragments by this method assumes that straight isobases exist for the shorelines in the study area. The error introduced by the possible curvature of isobases is considered negligible as the curvature of isobases across such a small study area is potentially low, and the maximum distance of any shoreline segment from a projection plane drawn through the study area is <10 km.

Table 5-5: Statistical analyses of the linear trend surfaces for Shorelines A to K.

Statistical analysis of linear trend surfaces for each shoreline							
	SS_r	SS_d	r^2	SE_y	F	C.L.(%)	n
Sl-A	000.768	02.081	0.270	0.510	001.477	>50.0	11
Sl-B	000.142	00.524	0.214	0.274	000.952	>50.0	10
Sl-C	010.729	00.677	0.941	0.311	055.470	>99.9	10
Sl-D	278.254	10.585	0.963	0.710	276.023	>99.9	24
Sl-E	322.594	13.205	0.980	0.813	244.306	>99.9	23
Sl-F	295.203	07.358	0.988	0.904	180.544	>99.9	12
Sl-G	450.219	14.170	0.985	1.190	158.865	>99.9	13
Sl-H	250.368	06.124	0.988	0.661	286.178	>99.9	17
Sl-I	341.088	07.449	0.989	0.910	206.060	>99.9	12
Sl-J	495.159	06.978	0.993	0.934	283.860	>99.9	11
Sl-K	029.203	05.930	0.912	0.920	017.237	>99.9	10

Table 5-6: Solutions to the linear regression equation for Shorelines A to K.

Solutions to the linear regression equations			
	b_0	b_1	b_2
Sl-A	09.162693	-0.020268	-0.025607
Sl-B	15.354060	+0.011061	+0.012254
Sl-C	20.643686	-0.062200	-0.119764
Sl-D	33.184864	-0.280984	-0.583561
Sl-E	35.543495	-0.278217	-0.269758
Sl-F	43.368752	-0.422162	-0.231253
Sl-G	47.966946	-0.539868	-0.220613
Sl-H	50.658264	-0.318681	-0.243155
Sl-I	60.850174	-0.593819	-0.102804
Sl-J	72.578636	-0.734846	-0.063241
Sl-K	76.125824	-0.281550	+0.020160

The use of trend surface analysis in the investigation of shoreline data and the

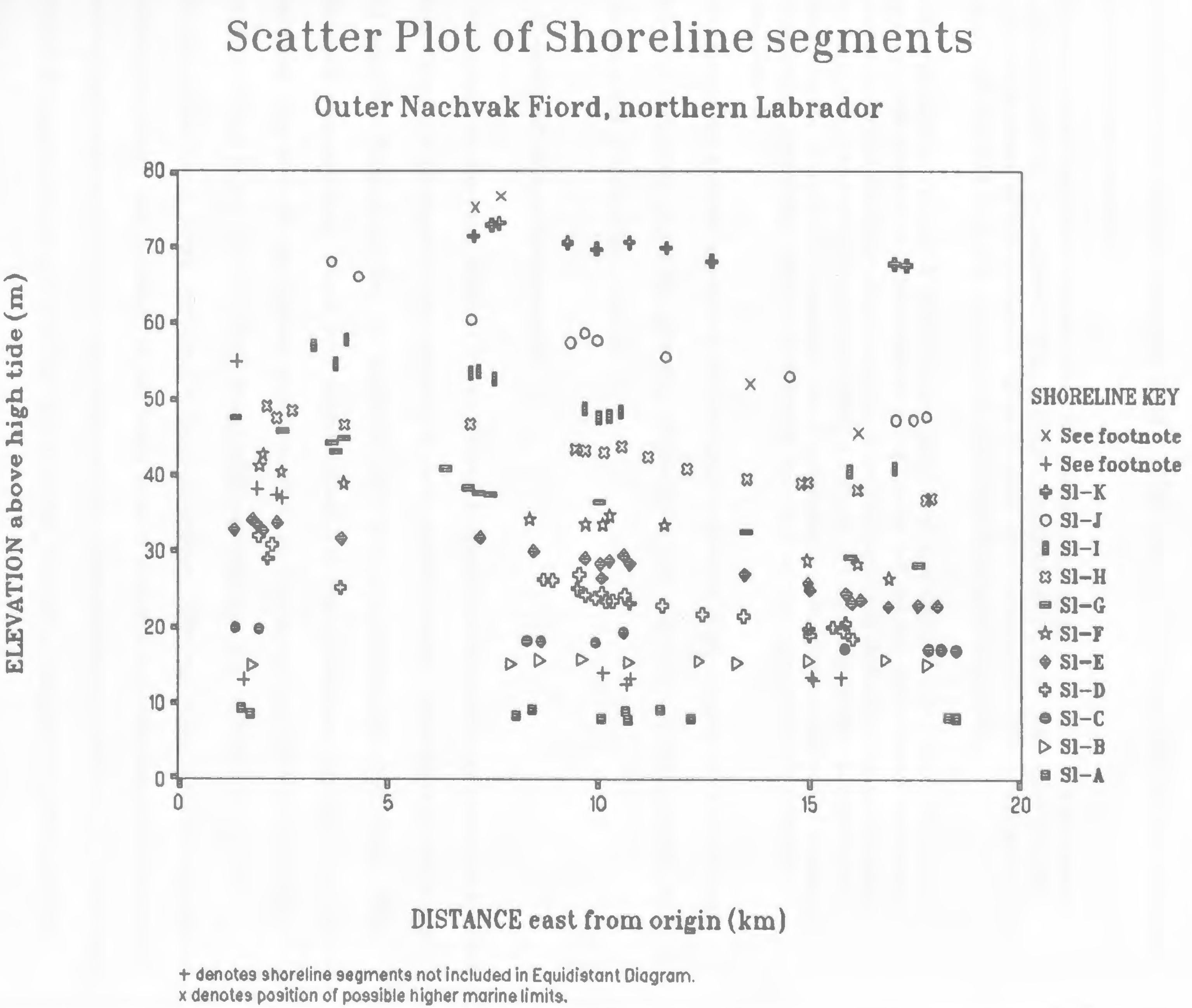


Figure 5-4: Scatterplot of all observed shoreline segments, projected in a plane normal to the mean isobase trend. The origin is located at grid reference *469000 6540000*.

reconstruction of shoreline planes has been widely discussed (Gray, 1972*a*; Cullingford and Smith, 1980). Although the major debate concerns the surface order that is most significant for a particular data set (for instance, linear, cubic, quadratic, *etc.*), the suitability of shoreline data for trend surface analysis has also been questioned. This type of analysis is valid for the shoreline data presented here because:

- Each shoreline segment is represented by a single elevation value. Groups of values along a segment are represented by a geographical mean sample value which is the arithmetic mean of the elevation values located at a mean geographical position. This approach ensures that the residuals for shoreline data are independent.

- Plots of residuals for each shoreline are presented in Figures 5-5 to 5-15. They are a general indication of the distributions of residuals around the linear trend surfaces. Autocorrelation between closely spaced shoreline segments does occur in some cases; for example, Shoreline D segments along northern Adams Lake valley in Figure 5-8. However, as Gray (1983) suggests, this is probably due to local conditions of beach formation. Generally, the set of residuals for each surface appears to be normally distributed.

No consistent pattern of residual distribution is obvious from Figures 5-5 to 5-15, and therefore no conclusions regarding shoreline displacement or shoreline formation can be deduced from the data set of shoreline residuals.

5.3.2. Trends of shoreline isobases

The trends of shoreline isobases were calculated using the solutions to the linear regression equations. Figure 5-16 displays the variation in these isobase trends. Shorelines A and B are omitted from this diagram as they are approximately horizontal within the study area. The pattern which emerges from Figure 5-16 demonstrates a gradual movement in isobase trend from approximately due west for the highest shoreline, Sl-K, to approximately SSW for Shoreline D. Shorelines C and H are anomalous to the general pattern and may reflect errors in the construction of the linear trend surfaces for these shorelines. The restricted areal distribution of data points due to the 'wedge' shape of the study area may result in a wide range of shoreline isobases for small fluctuations in the elevation data of a particular trend surface. Low shoreline gradients and large standard errors of the estimate (SE_y) would accentuate the effects of this inaccuracy.

The construction of an equidistant shoreline diagram requires that data points are plotted into a projection plane drawn normal to the isobase trends. In this study the isobases are not

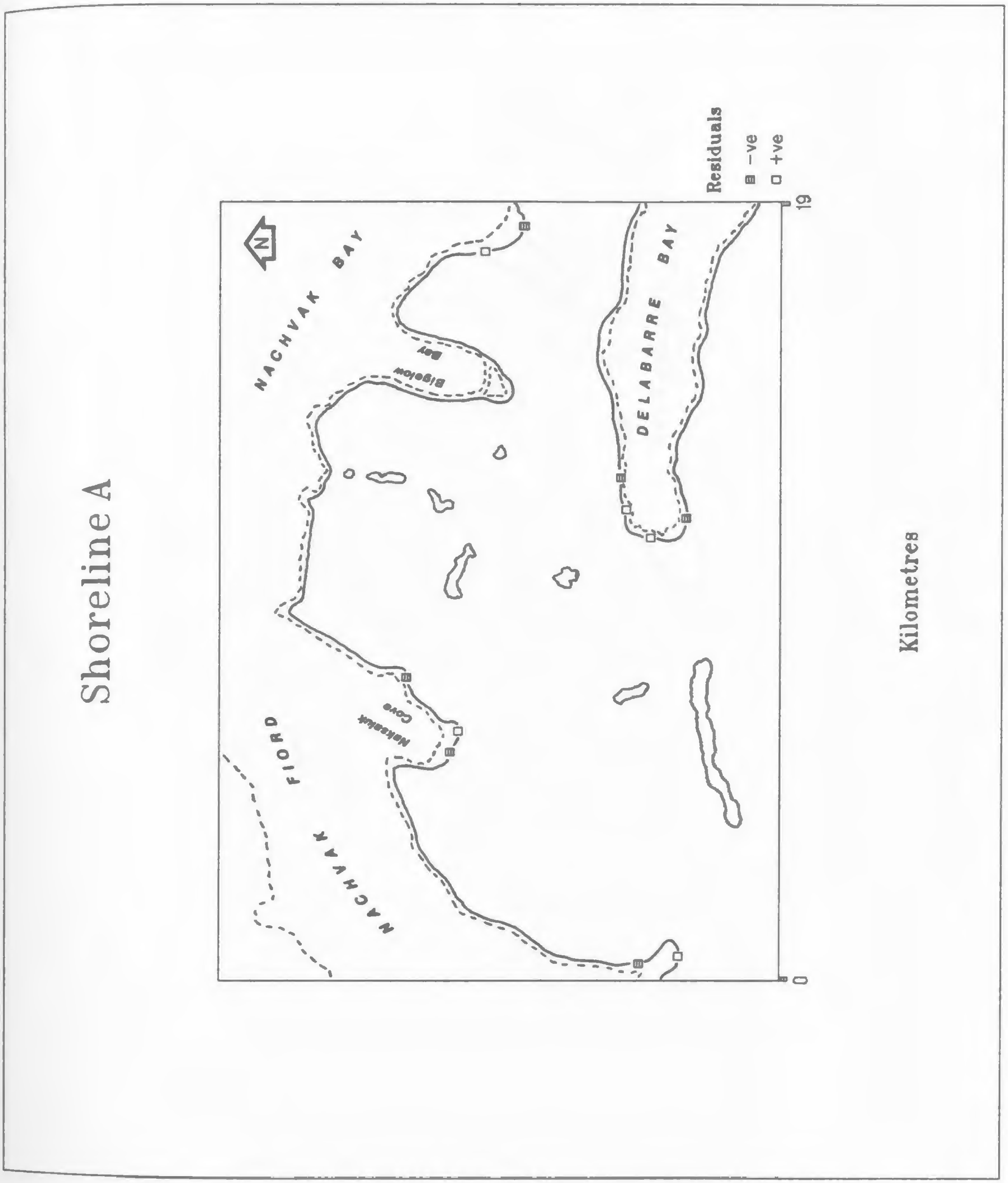


Figure 5-5: Plot of residuals for Shoreline A data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline A.

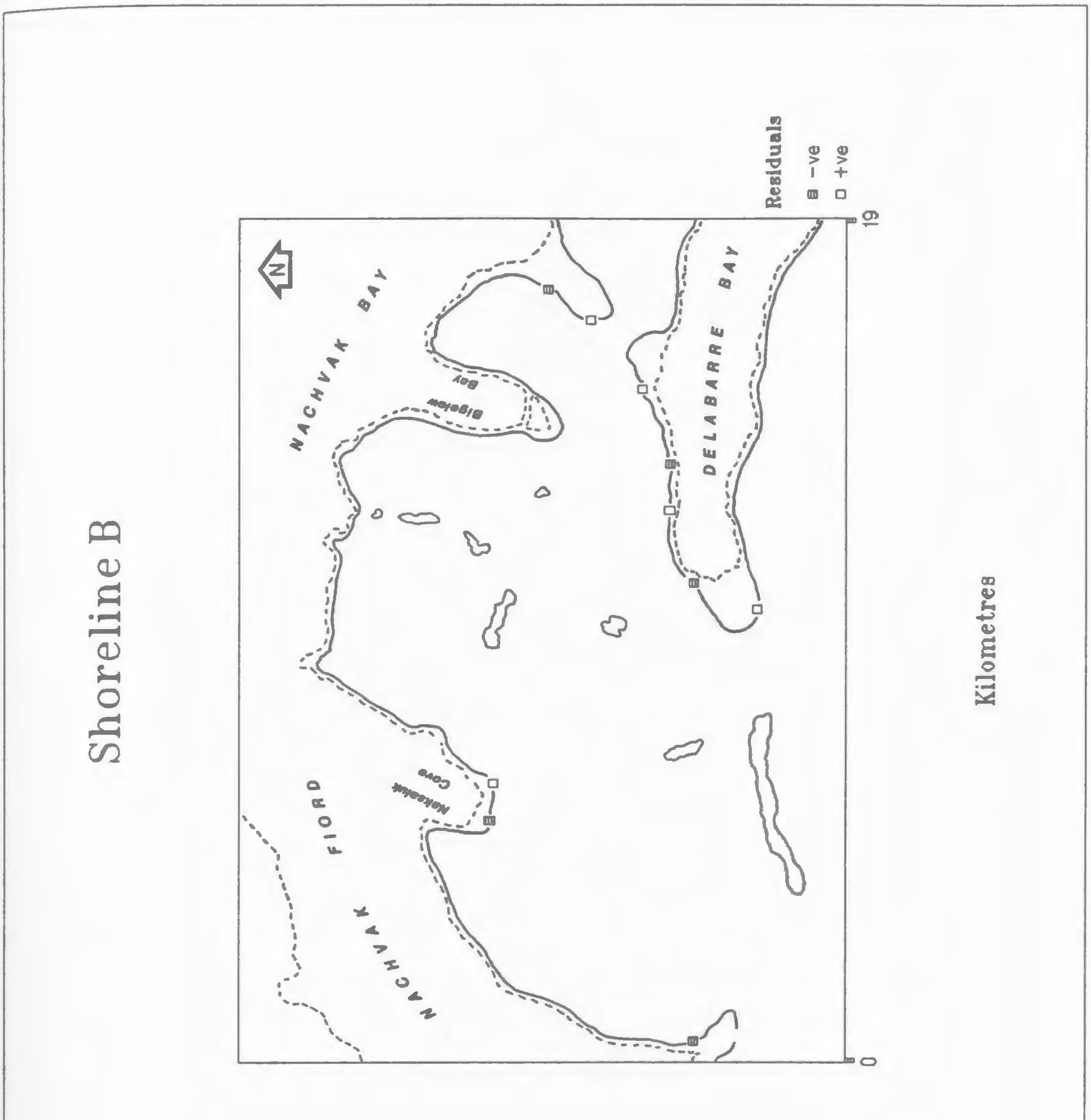


Figure 5-6: Plot of residuals for Shoreline B data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline B.

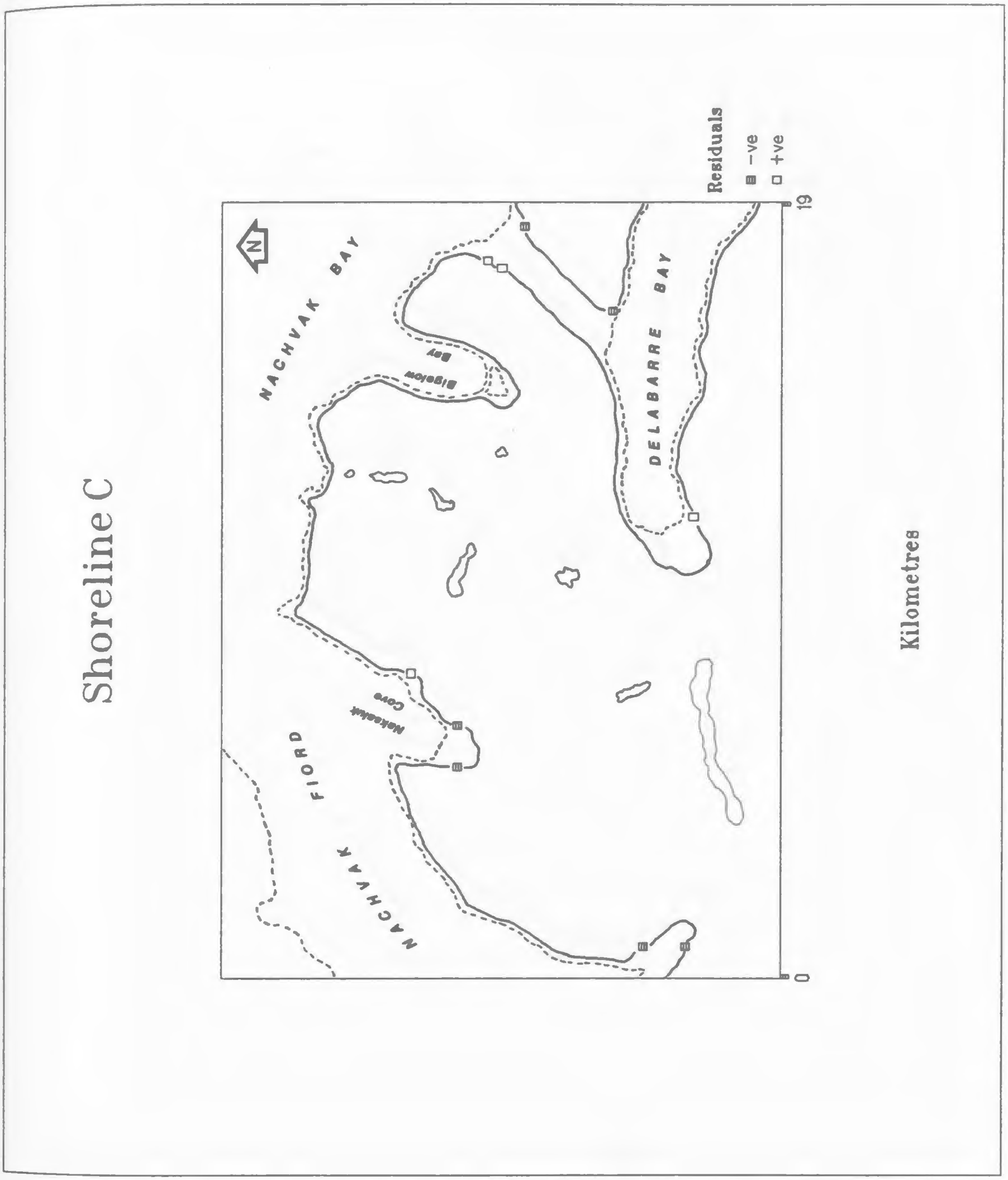


Figure 5-7: Plot of residuals for Shoreline C data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline C.

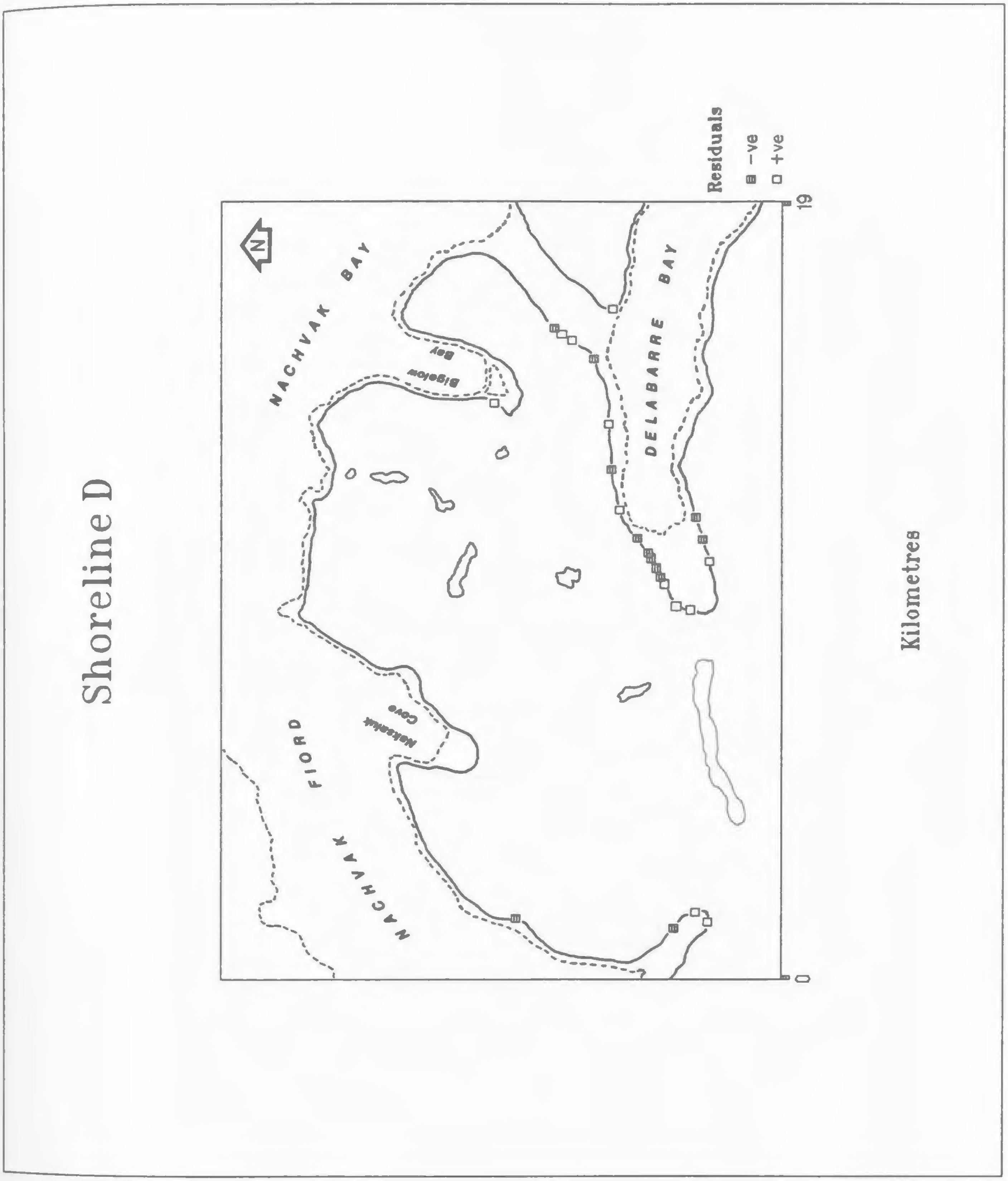


Figure 5-8: Plot of residuals for Shoreline D data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline D.

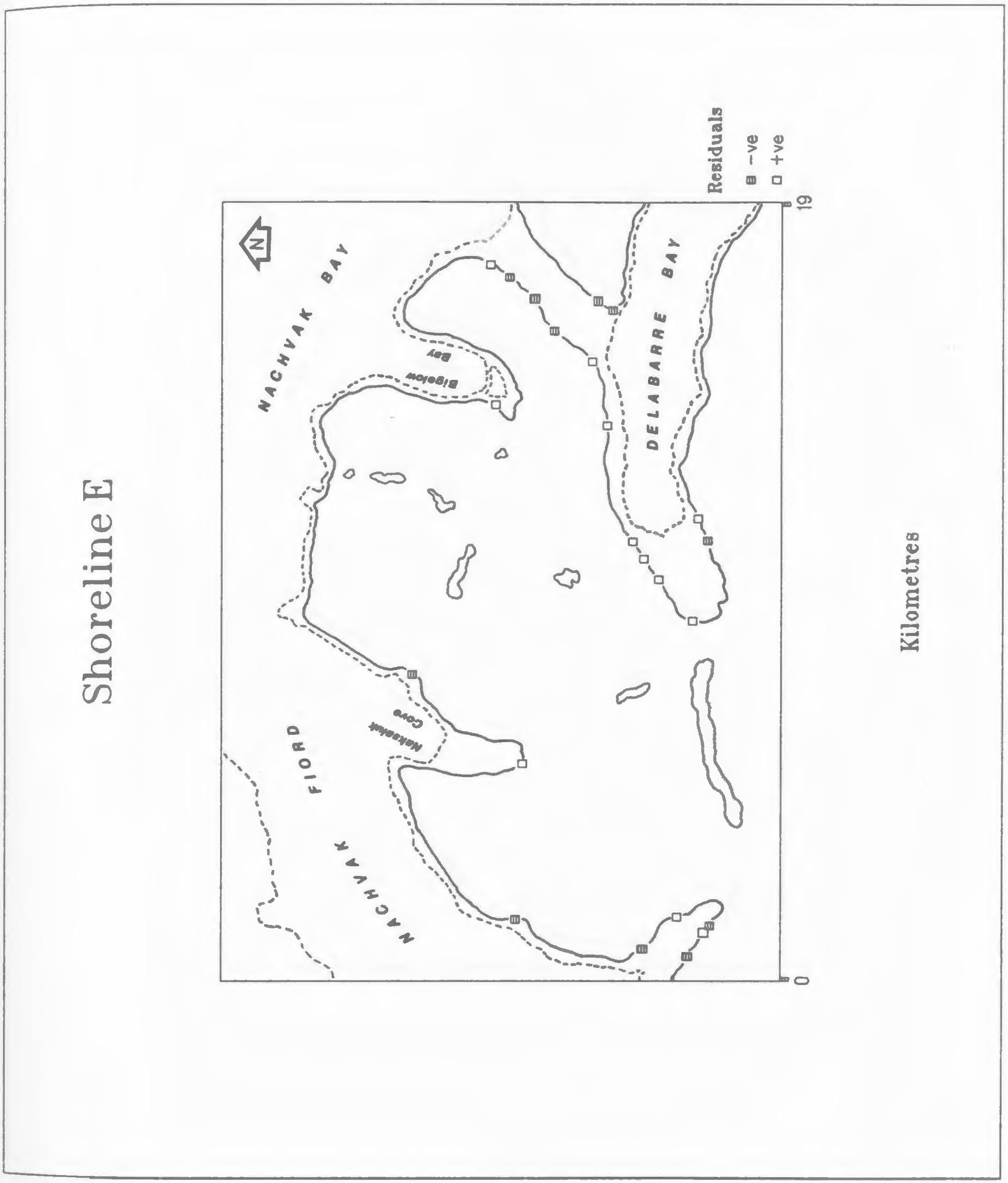


Figure 5-9: Plot of residuals for Shoreline E data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline E.

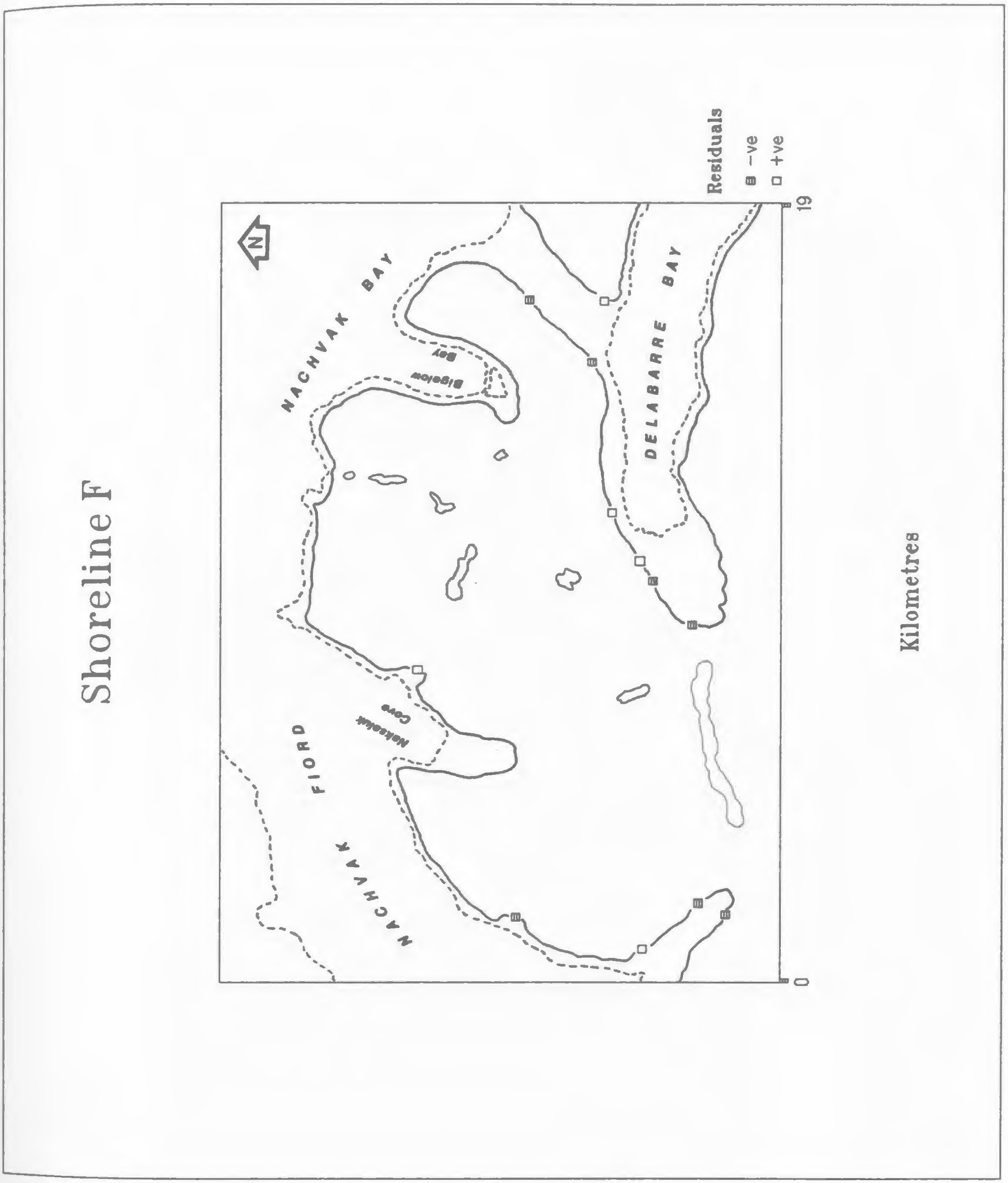


Figure 5-10: Plot of residuals for Shoreline F data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline F.

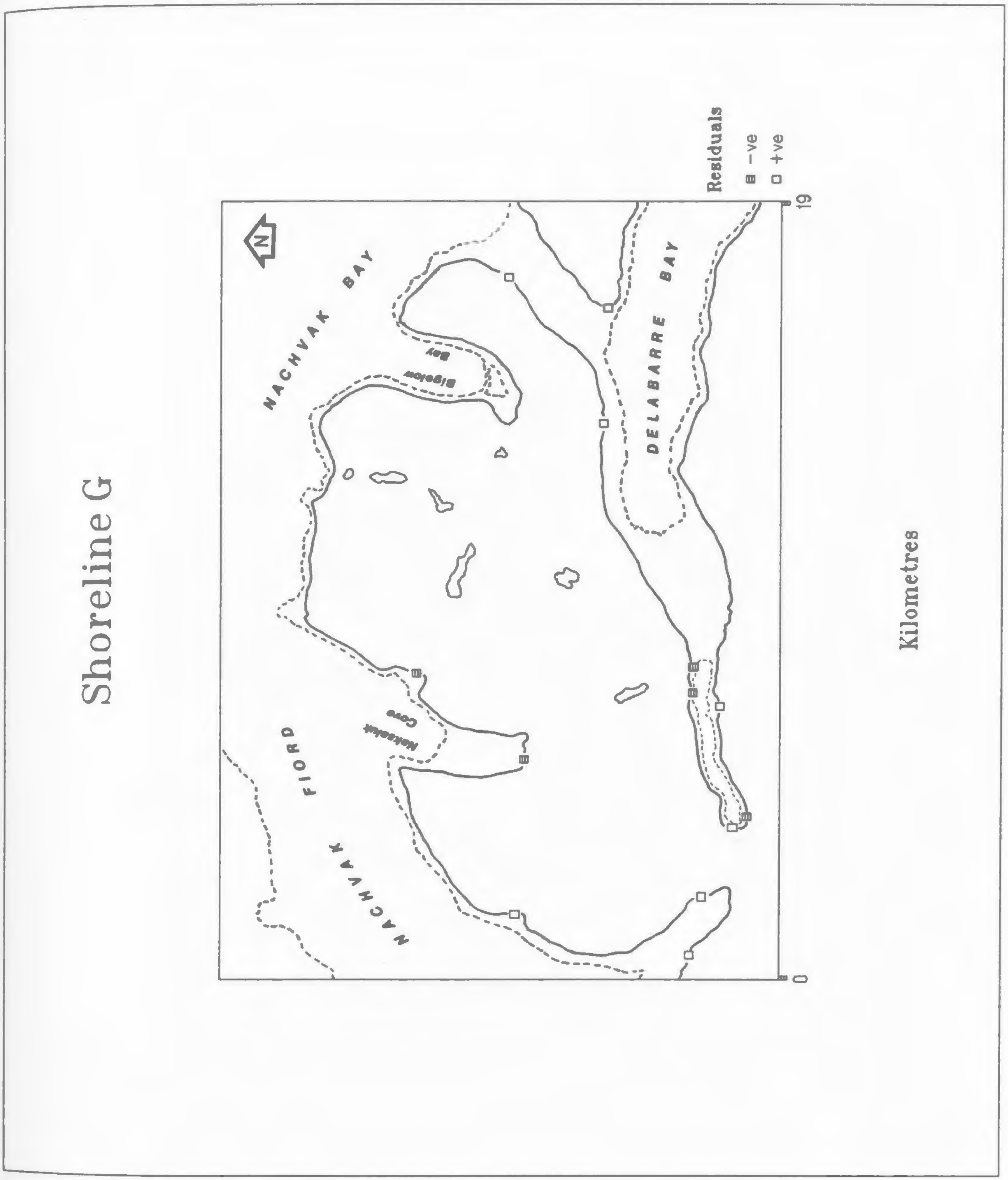


Figure 5-11: Plot of residuals for Shoreline G data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline G.

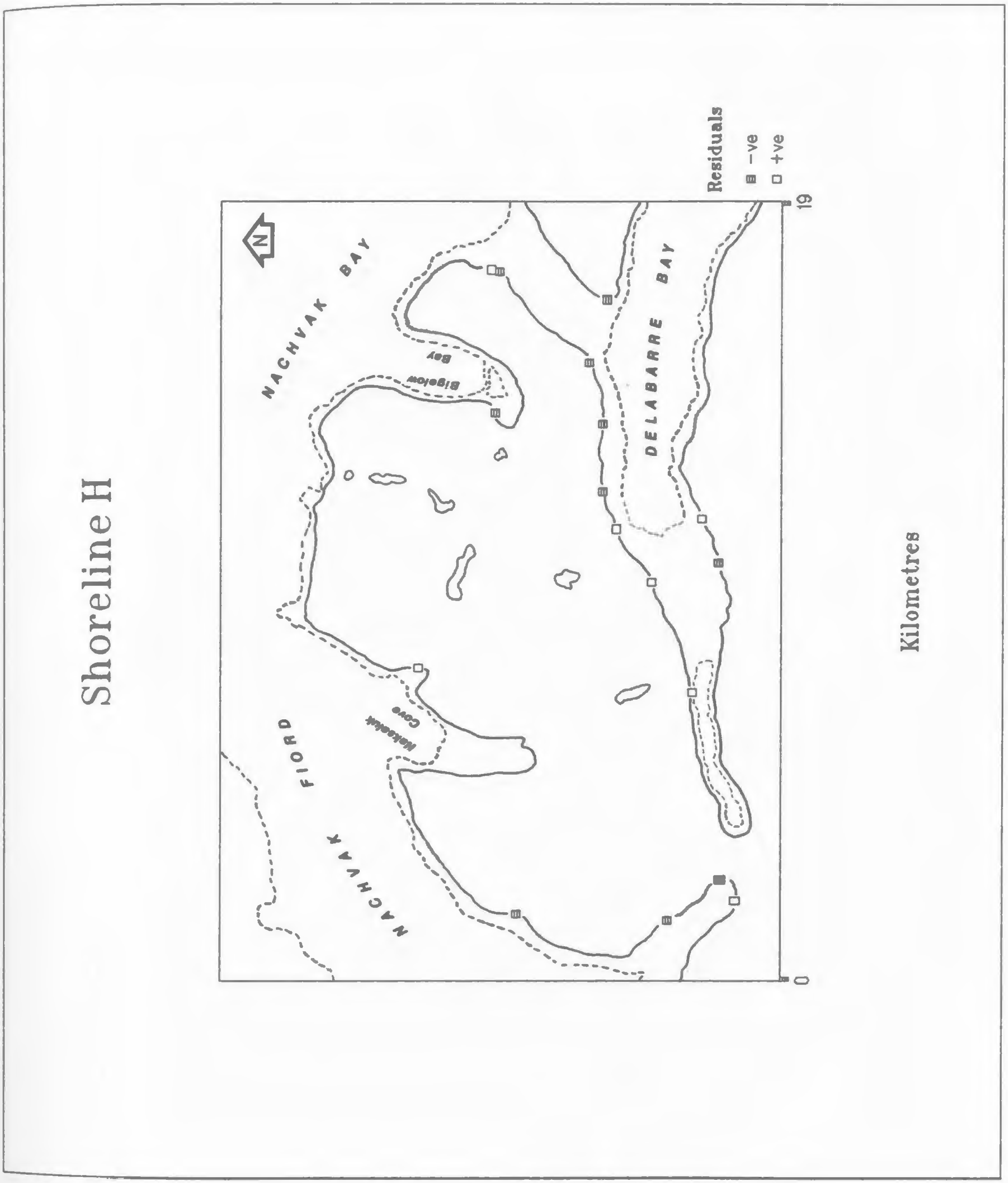


Figure 5-12: Plot of residuals for Shoreline H data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline H.

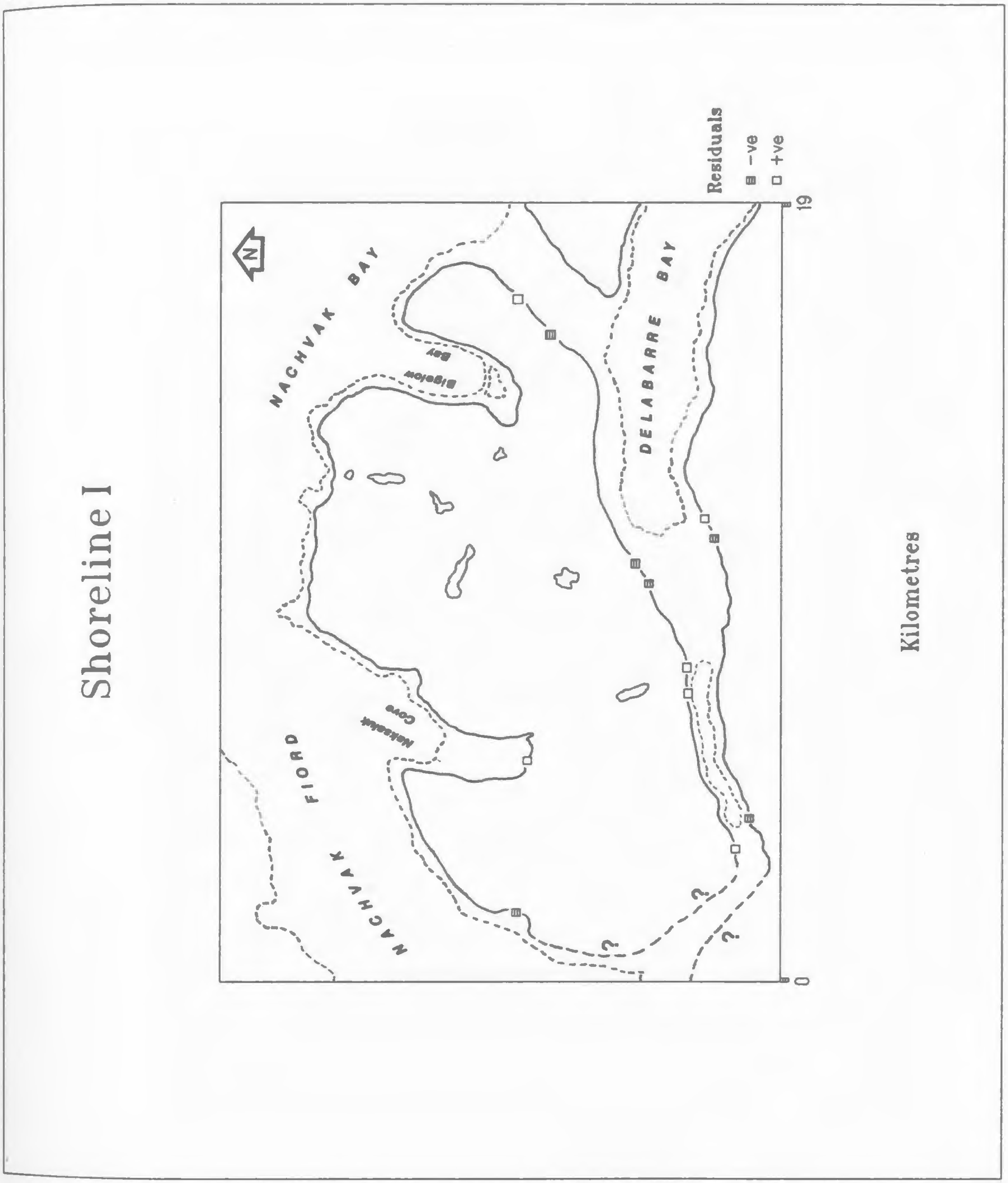


Figure 5-13: Plot of residuals for Shoreline I data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline I.

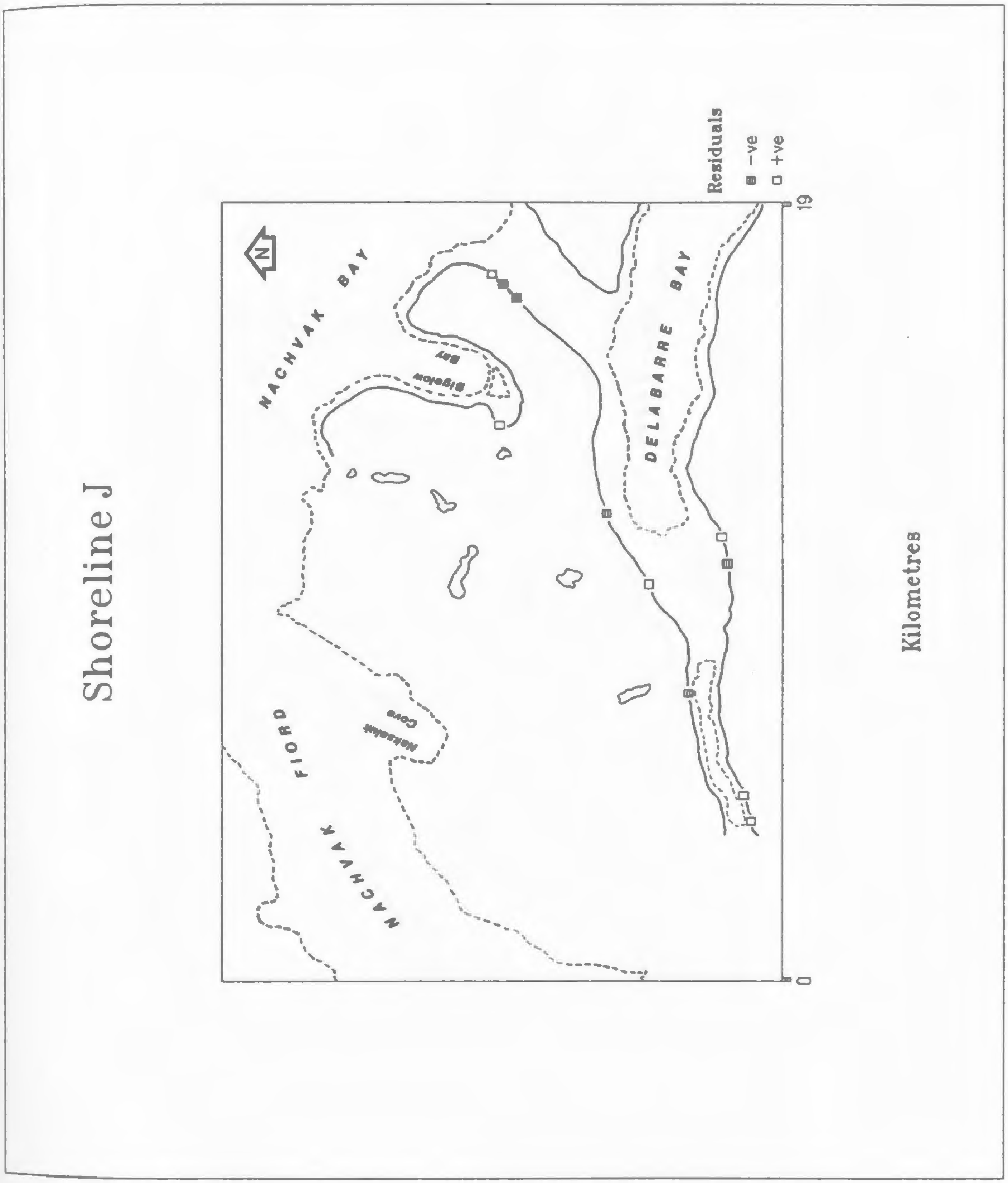


Figure 5-14: Plot of residuals for Shoreline J data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline J.

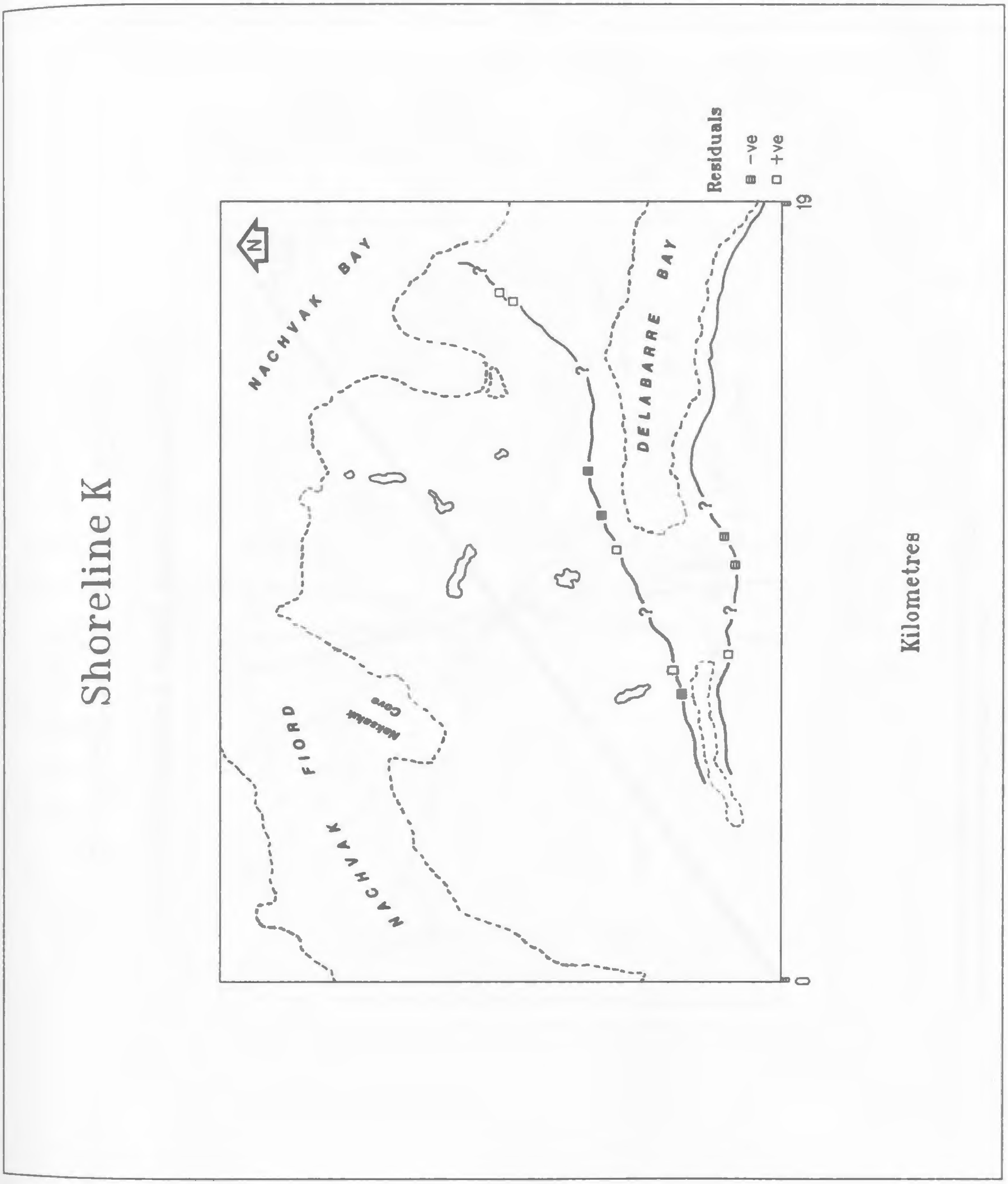
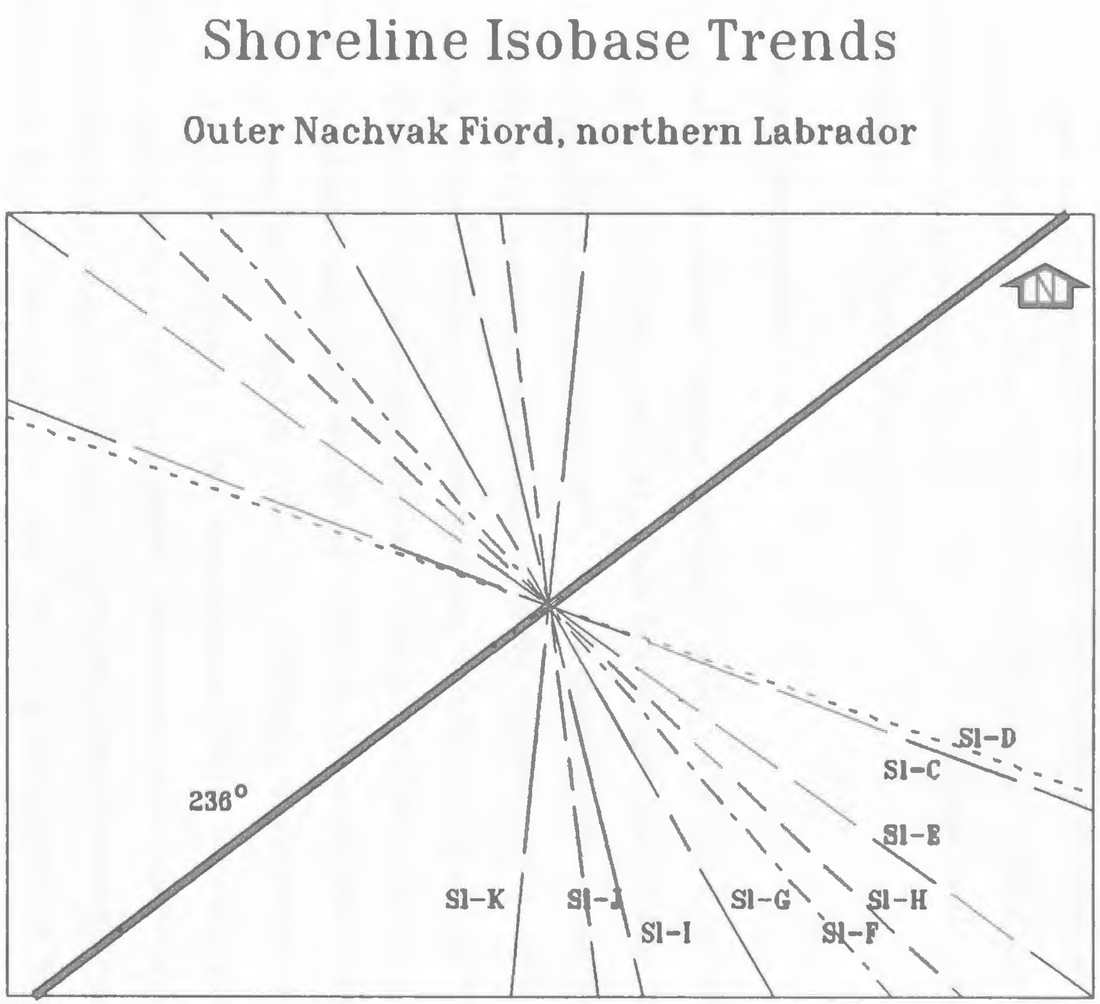


Figure 5-15: Plot of residuals for Shoreline K data. The dashed line represents the outline of the present coast, whereas the solid line represents the outline of the coast during the formation of Shoreline K.



The heavy line shows the direction normal to the mean isobase trend along which the equidistant shoreline diagram is drawn.

Figure 5-16: Diagram showing the trend of shoreline isobases for each Shoreline C to K. The orthogonal to the mean isobase trend is also shown.

parallel and therefore the projection plane is drawn normal to the *mean* isobase trend. The azimuth (from true north) of this projection plane is 236° in the direction of the assumed ice centre. Löken (1962*b*) used a projection plane of 205° azimuth which was approximately perpendicular to the mean shoreline isobases that he constructed for northernmost Labrador. He based his calculations on only two shorelines (Löken, 1962*b*, p.40-41), in contrast to the nine shorelines used in this study. Löken also found that his lower shoreline (Sl-3) had a more westerly trend relative to the highest shoreline, Sl-1 (SSW trend), which is a reverse pattern to the general isobase trend recognised here. The azimuths for Löken's projection plane and the projection plane proposed in this study are similar, the difference probably being due to the number of shorelines used to compute the mean position in each study.

5.3.3. Shoreline Representation

A scatterplot of shoreline segment altitudes is presented in Figure 5-4 and an equidistant shoreline diagram[24] is given in Figure 5-17. Both figures represent shoreline data which have been plotted into a projection plane bearing 236°. As this plane is drawn normal to the *mean* isobase trend, the shoreline points that are associated with the extreme isobase trends are plotted in the wrong position relative to each other, and therefore the presented shorelines have reduced gradients. The decrease in shoreline gradient, due to this arbitrarily chosen plane, can be calculated from the difference between the true slopes of the linear trend surfaces and the slopes of the projected shorelines (Table 5-7). The large differences in gradient for Shorelines C, D and K illustrate the extreme variances of these shoreline trends from the mean shoreline trend. The remaining shorelines have compatible gradients represented by the equidistant diagram.

The scatterplot of shoreline segments records the location and elevation of all observed segments along the projection plane (Fig. 5-4). The 'x' symbol marks the position of those features which had an ambiguous origin. These features were found at the upper limit of levelling traverses; however, the effects of solifluction obscured their original morphology and therefore it

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[24]Definition of an ***equidistant shoreline diagram*** is a graph of shoreline segment elevations drawn in a plane normal to the local system of isobases. The Y axis represents elevation above a known datum plane (usually present sea level) and the X axis represents distance. This diagram shows the extent and gradient of inferred shorelines and the differential emergence of the shoreline since its formation. Synonyms include strandline diagram (common in Scandinavia) and height-distance diagram (British studies).

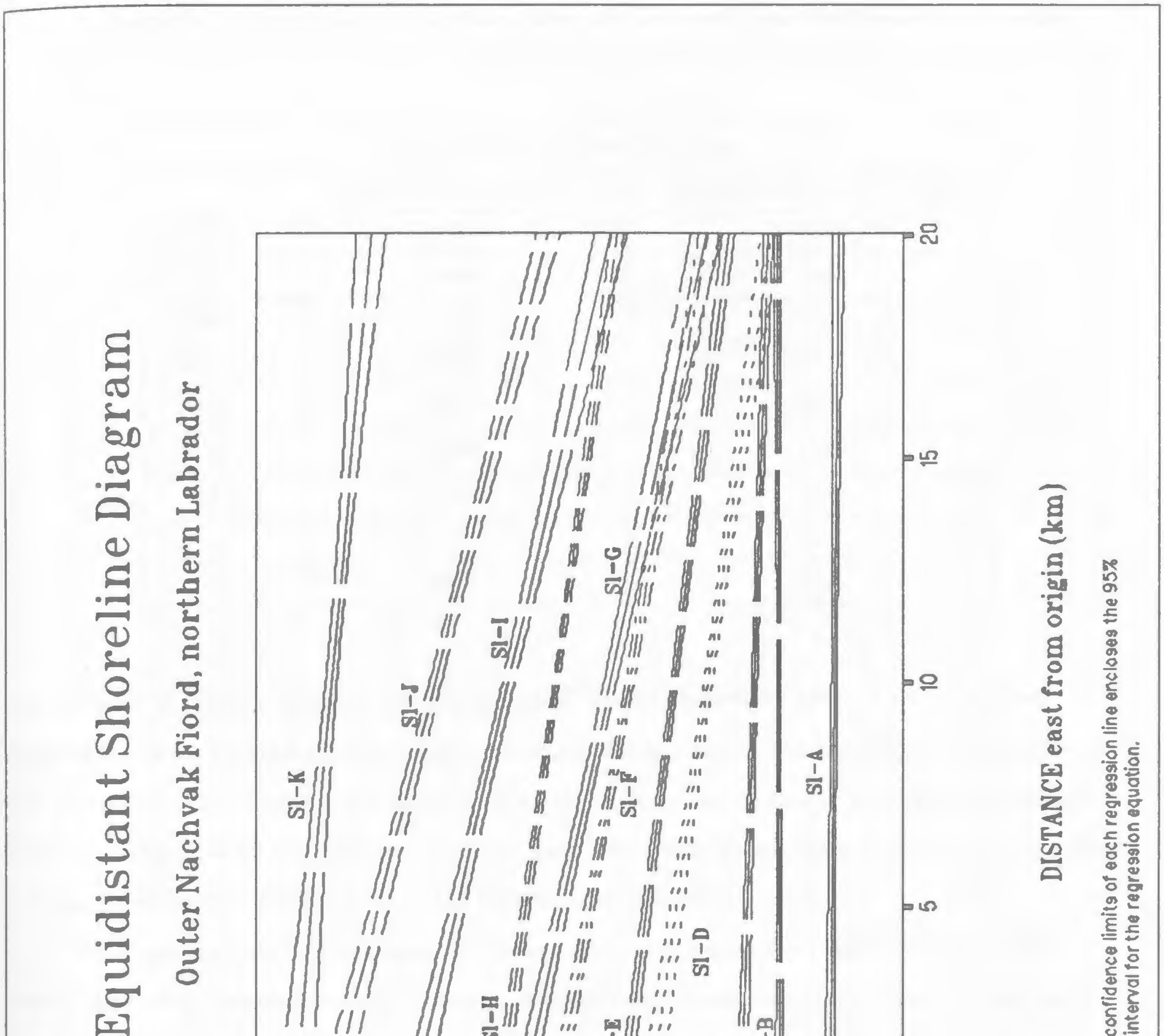

Figure 5-17: Equidistant shoreline diagram representing Shorelines A to K projected into a plane normal to the mean isobase trend.

Table 5-7: Comparison of shoreline slopes as calculated from the linear trend surface and the projected shorelines in Figure 5-17.

Comparison of shoreline slopes		
	Linear surface slope	Projected shoreline slope
Sl-C	1:3600	1:6140
Sl-D	1:800	1:1296
Sl-E	1:1400	1:1517
Sl-F	1:950	1:1023
Sl-G	1:800	1:844
Sl-H	1:1180	1:1333
Sl-I	1:780	1:759
Sl-J	1:700	1:722
Sl-K	1:3000	1:2065

was difficult to classify them as the marine limit[25] in the immediate area. The '+' symbol denotes the occurrence of shoreline segments which did not satisfy the correlation constraints of any shoreline. The sequence of segments at an elevation of *ca.* 13 m aht probably represents a horizontal shoreline at this altitude; however, only seven data points were recorded and therefore a linear trend surface could not be calculated for this shoreline.

Although the shoreline gradients in the equidistant diagram are slightly in error, the significance of the regression lines, drawn as best-fit lines through the data points, is very high (>99.9%, Table 5-8). This suggests that the shorelines resolved from the sample data set have >99.9% probability of occurring in the study area. However, the small number of observations on each shoreline (represented by 'n' in Table 5-8) restricts the inferences that can be made concerning these results.

If factors responsible for beach formation are constant and if conditions of continuous emergence occur, then it is expected that a single beach would be traced from the marine limit to

[25] *Marine limit* is defined in this study as the maximum elevation of observed marine evidence relating to higher stands of sea level, above present high tide. The term does not necessarily apply to Late Wisconsinan and Holocene sea level stands.

Table 5-8: Statistical analysis of the projected shorelines in the equidistant shoreline diagram, Figure 5-17.

Statistical analysis of regression lines/shorelines in the Equidistant Diagram							
	b_0	b_1	r^2	SE_y	F	C.L.(%)	n
Sl-A	09.06511	-0.02484	0.256	0.485	003.093	>85.0	11
Sl-B	15.22025	+0.01141	0.159	0.265	001.513	>70.0	10
Sl-C	20.02238	-0.08144	0.850	0.463	045.245	>99.9	10
Sl-D	31.85552	-0.38589	0.889	1.210	175.381	>99.9	24
Sl-E	35.00250	-0.33319	0.949	0.902	393.494	>99.9	23
Sl-F	43.64604	-0.48887	0.975	0.868	474.454	>99.9	12
Sl-G	48.11590	-0.59235	0.968	1.160	333.022	>99.9	13
Sl-H	50.41693	-0.37508	0.971	0.700	508.719	>99.9	17
Sl-I	61.38060	-0.62818	0.964	1.124	266.595	>99.9	12
Sl-J	71.68366	-0.69274	0.981	1.038	456.993	>99.9	11
Sl-K	75.50919	-0.24210	0.814	0.905	034.888	>99.9	10

the modern shore. However, such conditions are rare, and normally the rate of emergence is not constant. Still-stands or readvances during the recession of a glacier reduce the rate of emergence (Walcott, 1970, 1972), and hence shorelines experience sustained periods of wave wash and resulting beach formation[26]. Repeated still-stands may produce corresponding shorelines in the proglacial landscape. The elevations of these shorelines above present sea level may reflect the residual uplift of the earth's surface following the glacial still-stand, the effects of eustatic sea level change and the decrease in gravitational attraction between a deformable sea level surface and the land-based ice mass (Clark, 1976, 1980; Dyke, 1979*b*; Peltier and Andrews, 1976). Within a small area, the gradients of these shorelines probably represent the differential uplift of the earth's surface in response to isostatic recovery. As isostatic recovery decreases with distance from the ice centre, shorelines are tilted away from the ice centre, and the shoreline gradient reflects the proximity of the ice centre during its formation. The equidistant shoreline diagram in

[26]This is a simplified description of the relationship between ice margins and the formation of shorelines. Andrews (1970) suggests that former shorelines are the consequence of several different processes but that climatic change may be a prime factor (p.95).

Figure 5-17 can therefore be used to infer glacio-isostatic events from the characteristics of the shorelines presented.

Shorelines J to C display a range in gradients from *1:700* to *1:3600*. Shorelines J, I and H, Shorelines G, F and E, and Shorelines D, C, B and A show a direct relationship between shoreline gradient and shoreline altitude, as would be expected in a general glacio-isostatic recovery model. However, the inverse of this relationship occurs between Shorelines H and G and Shorelines E and D, suggesting reversals in the uplift rates associated with isostatic recovery. Such episodes of renewed downwarping of the earth's crust are associated with glacial readvances (e.g. Parry and Macpherson, 1964; Andrews, 1966). During these periods, the lowest raised shoreline (Shorelines H and E, respectively) was tilted towards the ice centre, in the study area.

5.4. Discussion

This section describes the individual shorelines constructed and presented in the preceding sections. A model of relative sea level change is proposed for the study area and is then extrapolated and subsequently tested for the Nachvak Fiord region. A relative chronology of sea level change is suggested and a radiocarbon dated shoreline is used to calculate a rate of emergence over the last 9000 years for the study area.

5.4.1. Shorelines

Shoreline K (67-73 m aht) is the highest recorded shoreline in the study area. It is geographically restricted to central and eastern Adams Lake valley, Delabarre Bay and Valley of the Flies (Fig. 5-15). The shoreline is a composite of the observed marine limits in this area, and therefore the correlation of these segments into one shoreline may be invalid, even though the 'fit' of the shoreline to the data is statistically significant (Table 5-8). An equally valid interpretation would suggest that each segment, or combination of segments, represents an individual shoreline and hence they are metachronous features. Consequently, Sl-K on Figure 5-17, does not show the true extent and geometry of sea level during formation of these segments. As no equivalent shoreline data were observed east of the field area (beyond Gulch peninsula is the Labrador Sea), neither hypothesis could be tested. Therefore Sl-K data may represent one shoreline or several shorelines which exist as the marine limit in the locations where they were observed.

Shoreline K data points were recorded from marine terraces cut in drift. As mentioned previously, some terraces were observed above Sl-K data (*Y02* and *Y03* in Table C-1; Fig. 5-4) but they had an ambiguous origin and therefore could not be positively identified as the marine limit along the traverse.

Shoreline J (47-68 m aht) occurs in Adams Lake valley, Delabarre Bay, Valley of the Flies and Bigelow valley (Fig. 5-14). It exists as the marine limit in western Adams Lake valley and Bigelow valley, but elsewhere it lies below Sl-K data. In Valley of the Flies this shoreline is represented by the uppermost beach on a raised gravel/sand spit (Fig. 5-18). This feature has been modified by subsequent lower sea levels and numerous beach ridges were recorded below Shoreline J. Sand dunes occur along the back edge of this beach (Sl-J) and result from wind erosion of the sandy beach material in this exposed area. In central Adams Lake valley this shoreline is represented by the planed surface of the western part of Adams Lake moraine. The scatterplot of shoreline segments (Fig. 5-4) suggests that segments *Y01* and *Y04* (Table C-1) which were described as having an ambiguous origin, would be correlatives to Sl-J segments if they were classified as marine features.

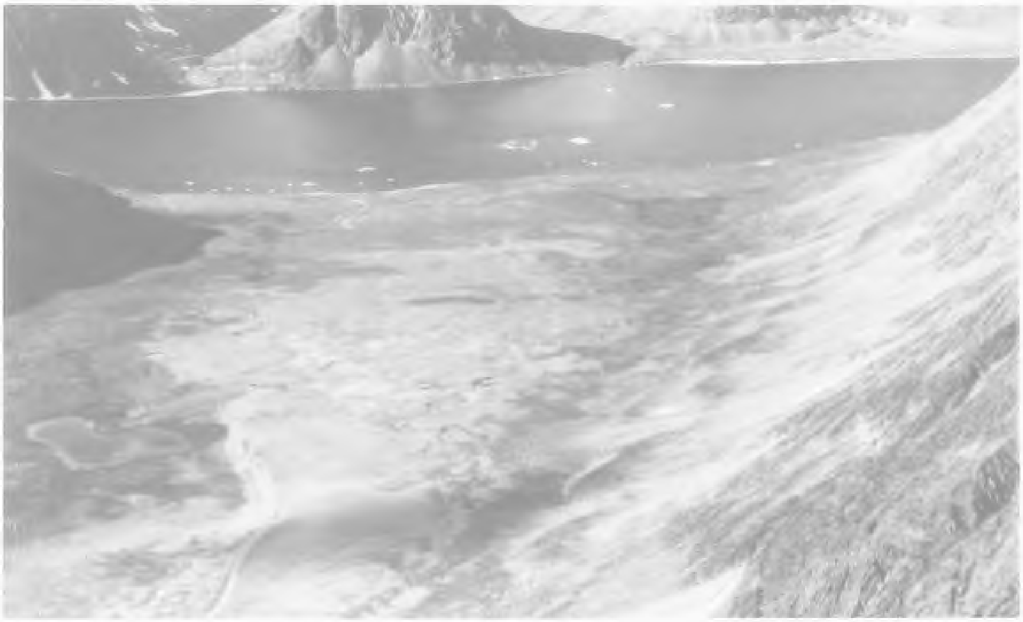

Figure 5-18: Photograph of the raised gravel/sand spit in Valley of the flies.

Shoreline I (40-58 m aht) occurs extensively throughout the study area (Fig. 5-13). During the formation of Sl-I most of the study area stood as an island off the mainland coast. Sl-I segments in Naksaluk valley and Shrule Cove (Figs. 2-2 & 5-19) are the marine limits in these areas. Segment *X11* (Table C-1; Fig. 5-4) was identified as the marine limit in northwestern Tinutyarvik valley but failed to satisfy the elevation constraint for Sl-I segments. *X11* is a washing limit with numerous boulders strewn along its surface. The back edge of this limit is defined by a steep bedrock outcrop. The use of washing limits as reliable indicators of the marine limit has been discussed by Gray (1972*b*), and they have been used extensively in Scandinavia (Stephens and Synge, 1966; Synge, 1969) where such limits are "particularily impressive". Synge (1969) presented many examples of washing limits, some of which had an elevation range of 2.4-5.0 m (p.196). If such ranges are typical for washing limits, then the water level they represent cannot be determined with certainty within an altitudinal range of two or three metres. Therefore it is suggested that the marine limit in northwestern Tinutyarvik valley is represented by the washing limit at *X11* which would have a residual value of 2-3 m below the trend surface of Shoreline I.

During the formation of Shoreline H (37-49 m aht), most of the study area was joined to the mainland by a narrow isthmus located between Tinutyarvik and Adams Lake valleys (Fig. 5-12). The well-preserved beach ridge which appears to cross the floor of Tinutyarvik valley can be seen in Figure 5-20. The central portion of this ridge has been eroded by the river flowing north down the valley to Nachvak fiord. The corresponding beach in western Adams Lake valley has been obscured/eroded by the alluvial fan deposits originating from a moraine-dammed lake to the south of Quartzite moraine (Fig. 5-21). Elsewhere this shoreline is marked by marine cut terraces (in drift) and beach ridges. A good example was observed in Bigelow valley (Fig. 5-22).

Shoreline G (28-48 m aht) is recorded by well-defined beach ridges in many parts of the study area (Fig. 5-11). In western Adams Lake valley, Sl-G is marked by two beach ridges and a terrace eroded into the alluvial fan deposits (Fig. 5-21). This would suggest that the alluvial deposits were laid down prior to the formation of Sl-G but following the Sl-H sea level. Segments *G01* in Naksaluk valley and *G13* in eastern Valley of the Flies are distinct beach ridges with marked crests (Figs. 5-23 & 5-24).

The distribution of Shoreline F segments (26-43 m aht) (Fig. 5-10) suggest that during the period between the formation of Sl-G and Sl-F, Adams Lake changed from a sea inlet to a freshwater lake. Segment *F13* in eastern Adams Lake valley was traced across the valley floor and is the highest of three well-defined beach terraces in this area. All three beaches display similar characteristics. Large boulders are frequently found at the back edges of the beaches and at the bottom of the foreslopes. Occasional boulders are perched along the beach crest which is predominantly composed of sub-rounded pebbles and cobbles[27]. There appear to be the remnants of a lagoonal area behind the crest of the uppermost beach, an observation which was made on many of the well-defined beaches in the study area.

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[27]These beach crests have been used by early residents of Labrador as dwelling sites evidenced by tent rings and other structures. A total of fifty observations were made on archaeological sites in the study area. The majority of these sites was located on raised beach deposits, at elevations as high as 50 m aht. [The archaeological information recorded was submitted to the Newfoundland Museum, St. John's.] Fitzhugh (1980) has documented pre-, middle and late Dorset sites and neo-eskimo sites in the Nachvak Fiord region.

Figure 5-19: Photograph of segment *I07* [SI-I] in Shrule Cove - marine limit.

Figure 5-20: Photograph of segment *H17* [SI-H] in Tinutyarvik valley.

Figure 5-21: Photograph of alluvial fan deposits in western Adams Lake valley. Note segment *G06* [Sl-G] and the terrace eroded into alluvial deposits.

Figure 5-22: Photograph of segment *H12* [Sl-H] in Bigelow Bay.

Figure 5-23: Photograph of segment *G01* [Sl-G] in Naksaluk valley.

Figure 5-24: Photograph of segment *G13* [Sl-G] in eastern Valley of the Flies.

Shorelines E (23-34 m aht) and D (19-31 m aht) are extensively represented in the study area (Figs. 5-9 & 5-8). Well-defined Sl-E beaches were observed in Naksaluk valley (Fig. 5-25), Adams Lake valley and Bigelow valley. Shoreline D, which was used as a reference shoreline in the correlation of all beach segments, is the most distinctive raised shoreline in the study area. Segment *D15* in Tinutyarvik valley represents a marked beach level which is oriented east-west across the valley. Segments *D01-D14* form part of an extensive shoreline that was traced from Adams Lake valley to Delabarre Bay. Segment *D19* is a well-preserved beach in Bigelow valley.

Figure 5-25: Photograph of segment *E23* [Sl-E] in Naksaluk valley.

Evidence relating to Shoreline C (17-20 m aht) is relatively sparse along the outer coast (Fig. 5-7) although Sl-C segments are marked shoreline features where they occur in the study area.

Shorelines B (15.5 m aht) and A (9 m aht) are recognised as horizontal shorelines (Figs. 5-6 & 5-5). The series of segments recorded at an elevation of approximately 13 m aht (Fig. 5-4) may also be classified as a horizontal shoreline, even though it is not formally recognised as a shoreline because of limited data points. These shorelines occur throughout the study area and are recorded by beach ridges and marine-cut terraces at these elevations.

5.4.2. Glacio-isostatic model

The correlation and synthesis of shoreline data from the study area are used to construct a model of relative sea level change which can be presented as primarily due to glacio-isostatic events in the fiord region. Any gravitational effect on sea level will emphasize differential rebound, not obscure it, since gravity decreases with the square of distance from the ice margin, and isostatic uplift decreases by some lesser power function of distance from the former ice centre. The equidistant shoreline diagram summarizes the general model. If periods of shoreline formation are regarded as fluctuations in the general glacio-isostatic recovery process, then each of the shorelines, C to K, indicates a slow-down or still-stand in ice recession during deglaciation. Furthermore, the repeated downwarping of the crust is evidence of renewed glacial activity on two occasions during this period. In summary, a tentative, stepwise deglacial model with two glacial readvances is proposed for the Nachvak Fiord region.

This model can be tested by extrapolating the shorelines up-fiord in order to demonstrate the pattern of deglaciation to the west of the study area. This test will check the validity of the shorelines constructed from data in the study area, and will isolate probable sites of glacial still-stands or readvances, from the pattern of marine limit elevations along the fiord coastline. Figure 5-26 shows the location of selected sites which were used to test the model. Elevations on shorelines east of Ivitak Cove were recorded by the author in 1984. Shoreline elevations in Ivitak Cove are reported in Evans (1984) and Evans and Rogerson (1986). Information on shorelines west of Ivitak Cove were supplied by J. Gallagher (personal communication, 1986). Table C-2 (Appendix C) is a comparison of the predicted shoreline elevations calculated from the linear trend surfaces and the actual shoreline elevations as recorded by altimeter or surveyor's level. There is close agreement between these two sets of data considering that the majority of recorded elevations was obtained by altimeter [$\pm$1.338 m error in this study] and that the potential error margin associated with the extrapolation of the trend surfaces increases with distance from the study area.

A summary of the test data is given in Figure 5-27. The stepwise westerly decrease of the marine limit along the fiord suggests that fiord ice retreated in contact with the sea and that only those areas beyond an ice margin recorded evidence of sea level stands during deglaciation. Furthermore the western limit of a shoreline represents the maximum easterly position of a glacial still-stand, or readvance, associated with the shoreline.

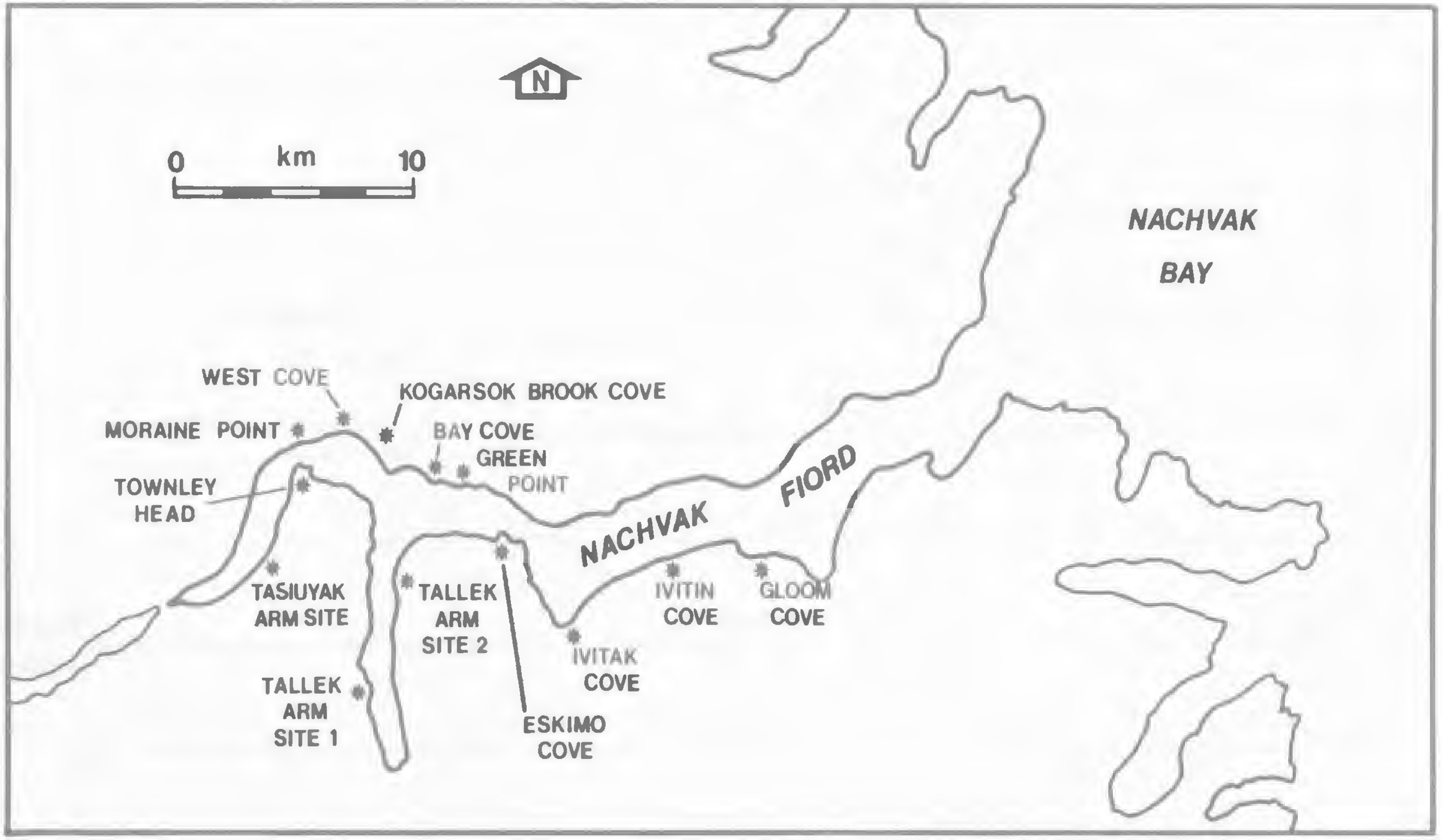

Figure 5-26: Location of sites selected to test the glacio-isostatic model in the fiord region.

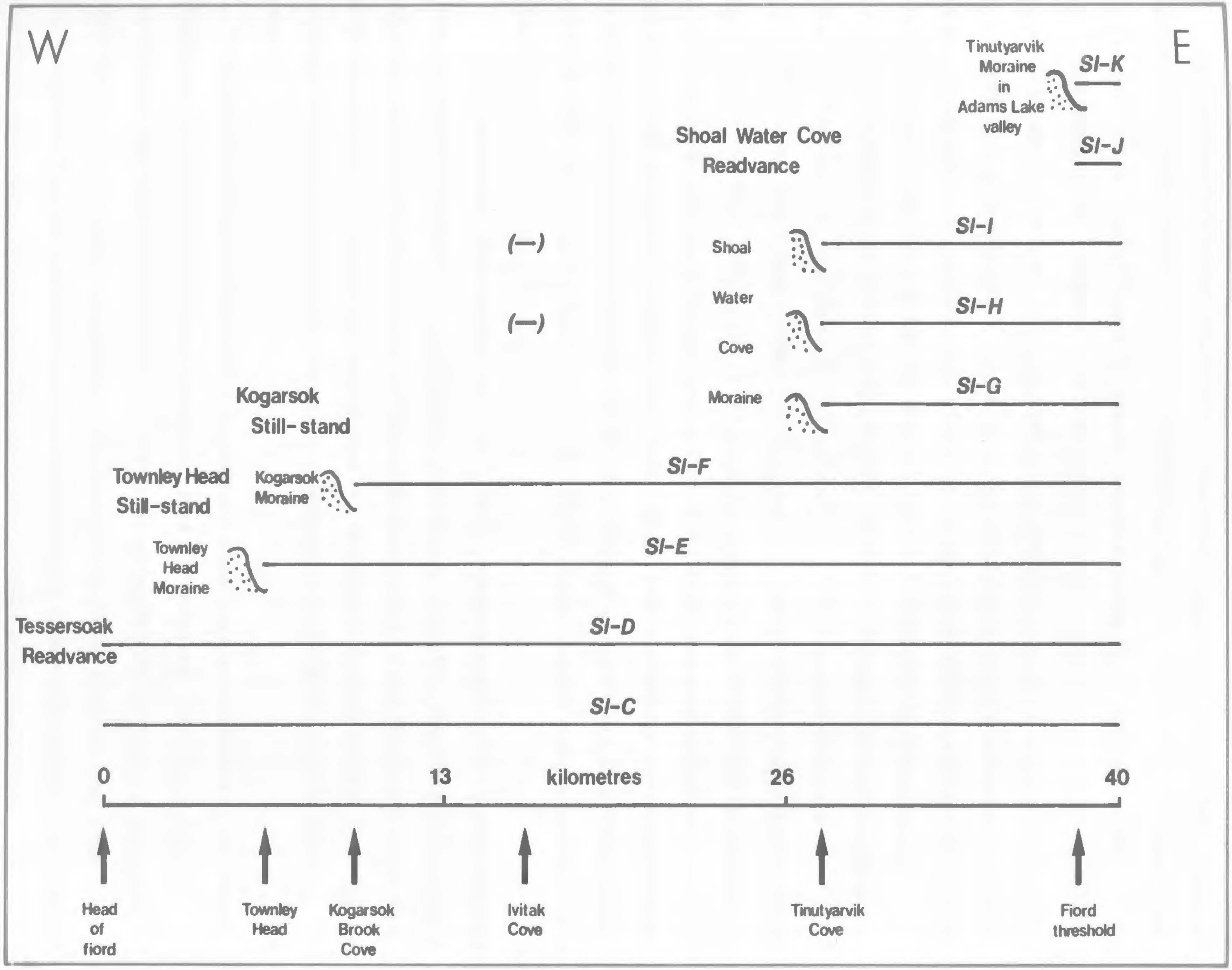

Figure 5-27: Diagram showing the pattern of shoreline distribution in the fiord and associated ice margins.

As mentioned previously, the existence of Shoreline K is questionable and therefore the most easterly ice margin that can be postively identified from the data is situated west of Bigelow Bay and probably on the fiord threshold. Recession from this margin was interrupted on two occasions, during which Shorelines I and H were formed. The westerly extent of deglaciation is unknown because the subsequent glacial readvance eroded the evidence for these former shorelines in the fiord, west of Tinutyarvik Cove. However the preservation of two shorelines at 67 m and 53 m in Ivitak cove is interpretated here as evidence of glacial retreat west of Ivitak Cove during this phase. Evans (1984) and Evans and Rogerson (1986) interpretated these shorelines as evidence of ice-dammed glacial lakes formed during the glacial maximum, the Nachvak phase. The 67 m shoreline is situated distal to the Nachvak phase lateral moraine in the lower McCornick valley (Base Camp moraine) and is reported to plane its southern face. As Shoreline I postdates the maximum glacial phase in the fiord, this evidence is consistent with the proposed model. The 53 m shoreline is located on both sides of the Base Camp moraine and is interpretated by Evans and Rogerson (1986) to be a glacial lake shoreline formed when ice stood to the north of the Base Camp moraine. However no moraine associated with this ice limit was identified. This study suggests that the 53 m shoreline is marine in origin and is a correlative to Shoreline H.

The downwarping of the earth's crust due to this ice readvance resulted in Shoreline H being tilted towards the ice centre. The subsequent sea level stand (Shoreline G) would have submerged the evidence of Shoreline H west of Ivitin Cove if the readvancing ice had not reached as far east as Tinutyarvik Cove. Therefore the preservation of a Shoreline H segment in Ivitak Cove implies that fiord ice prevented the possible erosion, or modification, of this shoreline by the Sl-G sea level.

The geomorphological evidence for this readvance phase may be represented by the Shoal Water Cove moraine which is located immediately west of Tinutyarvik Cove (Fig. 4-5), and therefore this glacial phase is termed the Shoal Water Cove readvance. However, it should be noted that the limit of this readvance phase may have existed at any point between Ivitin Cove and Tinutyarvik Cove as the highest beach terrace recorded in Gloom Cove (Shoreline D) was backed by a steep bedrock slope where the preservation of higher terraces is unlikely.

The western limit of Shoreline F occurs between Tallek Arm and Kogarsok Brook Cove (Fig. 5-27). Fiord ice stood in this locality during the deposition of the Kogarsok moraine (Fig. 4-5) and therefore Shoreline F was formed during the Kogarsok still-stand.

Shoreline E extends as far west as Townley Head which is the location of a moraine that extends at least halfway across the fiord at this point. The bouldery crest of the moraine is exposed at low tide (Figs. 5-28 & 5-29). It is suggested that this moraine, the Townley Head moraine, was deposited contemporaneously with the formation of Shoreline E during the Townley Head still-stand.

Evidence of Shoreline D was recorded as far west as Tasiuyak Arm, the most westerly site selected for the testing of this model[28]. Shoreline D was formed during a readvance phase [Tessersoak readvance] of the receding ice sheet and therefore should be associated with a moraine system. R.J. Rogerson (personal communication, 1985) reported the position of distinct moraines south and west of Nachvak Lake and tentatively described them as 'Sheppard moraine' equivalents. Löken (1962*b*) described a set of moraines close to the drainage divide west of Eclipse Channel in northernmost Labrador and referred to them as the Sheppard moraines. He tentatively correlated these moraines with the Kangalaksiorvik readvance phase and the formation of a marked shoreline, 'Sl-3'.

Shorelines A and B are thought to represent major transgressive events during the postglacial period. Löken (1962*b*) associated a horizontal shoreline, 'Sl-4', at 15.5 m aht with a major transgression in northernmost Labrador.

5.4.3. Shoreline isobase trends

The azimuth of shoreline dip can be used to determine the location of former glacial centers and the migration of such centers during glacio-isostatic recovery. Although many broad assumptions must be made regarding the geometry of former ice sheets and the outline of the continental mass, there has been some success in the identification and location of ice centres which existed during the last glaciation. Barnett and Peterson (1964) projected the azimuths of orthogonals to various shoreline isobases from the Labrador-Ungava peninsula and concluded that

[28] J. Gallagher (personal communication, 1986) described a horizontal shoreline at approximately 60 m in this area but its origin has not been determined.

Figure 5-28: Photograph of Townley Head moraine, between high and mid-tide level, looking north from Townley Head. The 'x' marks the viewpoint of Figure 5-29.

Figure 5-29: Photograph of Townley Head moraine at low tide, looking back towards the viewer in Figure 5-28. The photograph was taken near the centre of the fiord.

the intersection of these orthogonals indicated the centre of glacio-isostatic recovery for the Laurentide ice sheet in this region. Their interpretation was supported by the field observations of Ives (1960) which suggested that an area near Schefferville was the site of one of the last glacial remnants of the Laurentide ice sheet although, as Barnett and Peterson (1964) point out, the centre of last ice need not coincide with the centre of ice dispersal. Andrews and Barnett (1972) produced a similar map showing the dips on the postglacial uplifted surface of eastern and central Canada which tentatively indicate a multi-domed Laurentide ice surface. One centre of postglacial uplift is located in central Labrador-Ungava, near the Schefferville area.

Location of the centre of postglacial uplift at a specific time during the postglacial period requires that a system of synchronous shorelines is used to project the ice centre. Herein lies one of the problems in this method, since it may not be possible to determine a system of synchronous shorelines across a large area. In Figure 5 of Barnett and Peterson (1964), projections of orthogonals from Löken's (1962*b*) isobase system in northernmost Labrador are compared with orthogonals from glacial lake shoreline isobases in central Labrador-Ungava. They acknowledge that the former shorelines are much older and therefore are not expected to coincide with the ice centre calculated from the glacial lake shorelines. This study corroborates the observations of Barnett and Peterson, and tentatively describes the migration of the ice centres that isostatically affected the Nachvak Fiord region.

It is suggested that there is a possible correlation between Shorelines G and D of this study and Sl-1 and Sl-3 of Löken's study (1962*b*), respectively. This correlation is based on their association with readvance phases in the region, and to some degree, on the similarity with respect to shoreline gradients in each case. Acknowledging the large potential error inherent in such a correlation, the assumption is carried one step further and it is suggested that the intersection of the orthogonals to the isobases of Shoreline G and Sl-1 is the location of the ice centre which existed during the formation of this synchronous shoreline. The position of the intersection point is located approximately 60 km east of the George River and 120 km west of the head of Saglek Fiord (Fig. 5-30). The position of the ice centre during the formation of Shorelines J and I would be located to the north of this area based on the orthogonals of these shorelines. Similarily, the ice centre associated with the formation of Shorelines F, H and E would be to the south, and possibly east, of this location as suggested by the direction of dip on these shorelines.

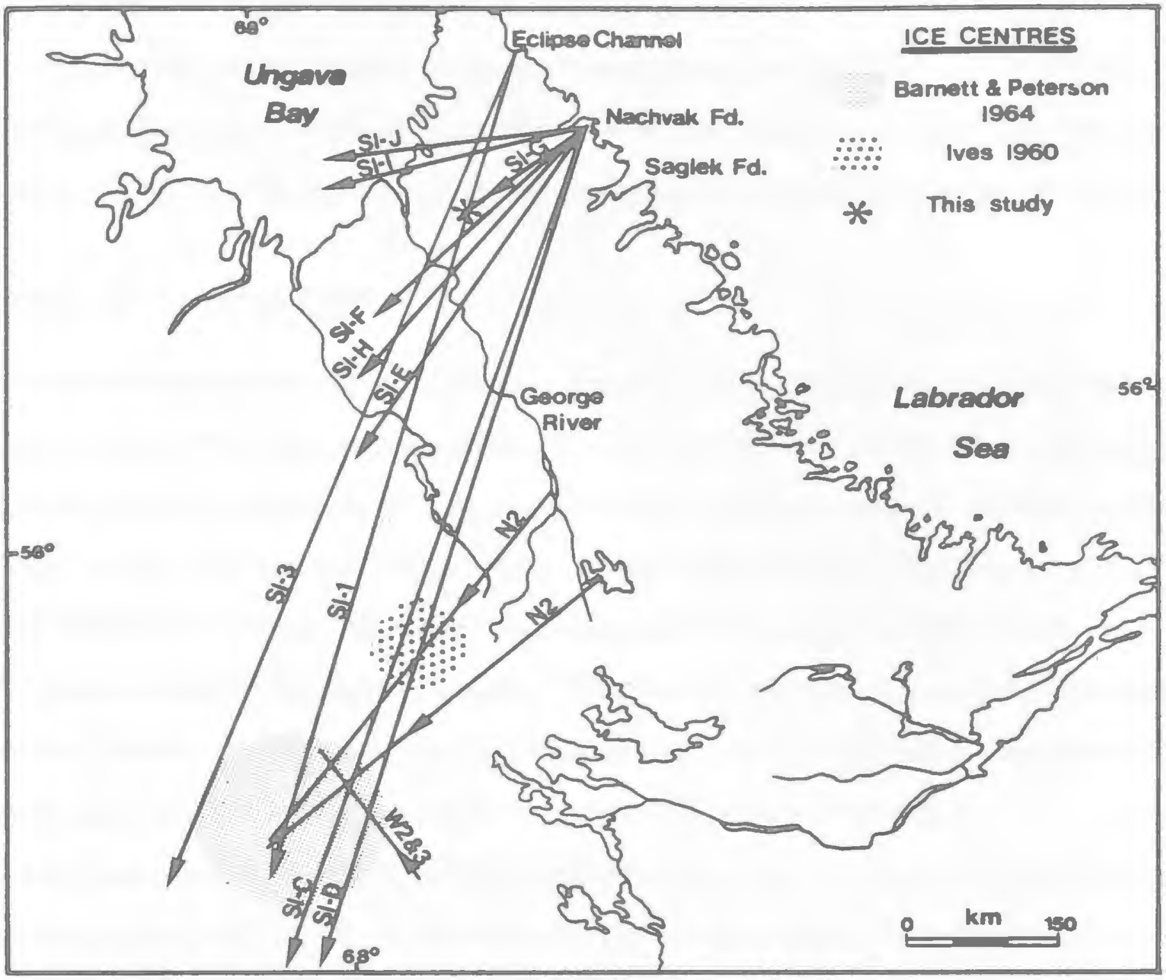

Figure 5-30: Map showing predicted ice centres in the Labrador-Ungava peninsula based on this study and various other sources. *After* Barnett & Peterson, (1964).

These proposed ice centres suggest that large volumes of ice prevailed in the area of the George River and Whale River, to the southeast of Ungava Bay, during this period of deglaciation. Support for such a conclusion can be drawn from the observations that the youngest shoreline of this series, Shoreline E, was formed while ice stood against Townley Head moraine in Nachvak Fiord, and that large ice-dammed lakes existed to the west of the Labrador-Ungava drainage divide in late-glacial time (Ives, 1960). The persistance of ice and lake water in this area would alter the isostatic recovery pattern of northern Labrador relative to that of the central Labrador-Ungava peninsula.

The orthogonals to Shoreline D and Sl-3 isobases do not intersect within the area outlined in Figure 5-30. However the azimuths associated with Shorelines C and D both pass through the proposed ice centre of Barnett and Peterson (1964) in the Schefferville area which refers to a period after the glacial maximum and before the final wastage of the Laurentide ice sheet.

Results of this study cannot evaluate the relative importance of the various ice centres, nor can they determine the history of such ice centres throughout the late-glacial period. The tentative interpretations and conclusions outlined above only relate to the Nachvak region and the ice centres which affected isostatic recovery during specific periods of shoreline formation.

5.5. Rates of emergence

A river-eroded section through Shoreline D deposits was located in Adams Lake valley (Site AA; Figs. 3-9 & 3-10). The site is situated near to shoreline segment *D12* on Traverse 11 and is at an elevation of 22.5 m ±0.8 m aht. The section is interpreted to represent a transgressive sequence with marine clay overlying beach sands. A barnacle fragment (*Balanus balanus*, identified by W. Blake Jr.) from the marine clay was radiocarbon dated at 9170 ±100 years BP (GSC-4161) (see Section 6.2 for further details). This marine clay unit is tentatively associated with Shoreline D based on the elevation of the stratigraphic section 4 m below segment *D12*, and therefore this radiocarbon date is suggested to represent the age of Shoreline D.

The importance of this dated shoreline is such that the relative chronology of glacio-isostatic events described previously can be given some numeric dating control. Shoreline D was formed during a period of renewed downwarping of the earth's crust which is attributed to the Tessersoak readvance of Laurentide ice. The date of Shoreline D formation can also be tentatively assigned to this readvance phase, although the exact relationship between the readvance maximum and the shoreline cannot be ascertained. The formation of Shoreline C and the postglacial horizontal shorelines, A and B must be younger than 9000 years BP because of their lower elevations above high tide. Löken (1962*b*) identified a horizontal shoreline at 15.5 m aht in northernmost Labrador and suggested that it may be correlated with the maximum level of the Tapes transgression in Scandinavia which was dated at 4700 years BP (Faegri *in* Holtedahl, 1960, p. 425 & 455). Horizontal Shoreline B in the study area is approximately 15.5 m aht, and may be assigned a similar age.

The rate of emergence for the last 9000 years can be calculated for the study area by using Shoreline D as a reference level. The arbitrarily chosen origin at grid reference *469000 6540000* is used as the control point for this calculation, based on the elevations of the shorelines at this location[29]. Since the formation of Shoreline D at 9170 ±100 years BP, there has been 31.86 m of emergence which represents an average rate of 3.5 m/ka. This figure conflicts with the value of 1.25 m/ka extrapolated from Baffin Island data for the northern Labrador coast, by Andrews (1970, Fig. 7-8). This may relate to a greater ice mass melted from northern Labrador, relative to that from Baffin Island.

Evidence of land submergence occurring at present in the study area was recorded by the recent erosion of fossil storm beaches (with old Inuit dwelling sites on their surfaces) and the submergence of peat below the high tide level, particularly evident in the Ivitak Cove area (Rogerson, personal communication, 1986).

5.6. Summary

One of the main objectives of this chapter was to give a detailed and comprehensive report on the shoreline data recorded in the study area. An important aspect of this report is the quantitative assessment of the consistency of the data. Similar studies carried out in Scotland are used as a standard by which to compare data quality. Major discrepancies are examined in the context of the extreme field conditions under which the data were recorded. In general, data consistency compares favourably, and could be regarded as an interim standard for shoreline data acquisition in northern Labrador until larger scale studies are completed.

The eleven shorelines presented in this chapter are the result of computer-assisted shoreline segment correlation. Their accuracy is dependent on data quality because they are constructed as best-fit planes to the individual data sets. The results of statistical analyses on these linear surfaces are highly significant and therefore it may be concluded that these planes are accurate reconstructions of former sea levels. The extent and geometry of the shorelines are the bases of a glacio-isostatic model proposed for the study area. The occurrence of low gradient shorelines (Sl-H & Sl-E) above higher gradient shorelines would appear to be inconsistent with the view that a higher shoreline is the integration through time of uplift that affected that shoreline plus those below it. By convention, therefore, such shorelines could be regarded as the result of erroneous identification and correlation. However, the shoreline gradients are derived from data which have

[29] The exact elevation of each shoreline at this point is given by the value b_0, the intercept of the regression line with the Y-axis on Figure 5-17. These values are listed in Table 5-8.

common standards of definition and measurement and to omit them would be to misrepresent the results of the analysis. Since the axis of deformation for a glacial readvance need not be identical with that for an earlier glacial phase, and local sea level in the fiord may be perturbed by the proximity of ice during a readvance, the conventional approach may be in error. It lies beyond the scope of this thesis to analyse the geophysical factors which could result in the situation described. Tentatively, it is reasonable to hypothesize that readvances may be responsible for these irregularities.

The extrapolation of the shoreline model to the adjacent fiord area provides a test for the shoreline sequence and allows refinement of the model with respect to the pattern of glacio-isostatic recovery in the region. The development of a relative deglacial chronology in association with a late-glacial and postglacial emergence history is an important step towards the construction of an absolute chronology of late-Pleistocene events.

The identification of the Shoal Water Cove, Kogarsok and Townley Head moraines appears consistent with the proposed sequence of events in the glacio-isostatic model. The exact location of the Tessersoak Readvance is unknown but is clearly visible in the pattern of late-glacial emergence. An attempt is made to locate major ice centres which affected the Nachvak Fiord region, and although there is supporting evidence for these proposed centres, the procedure is prone to error and therefore the results are speculative. Radiocarbon-dated shells from sediments associated with Shoreline D suggest that this shoreline was formed approximately 9 ka BP, and that *ca.* 30 m of land emergence has occurred since this time in the study area.

Chapter 6
Quaternary chronology

6.1. Introduction

A tentative relative chronology of glacial events was suggested in Section 3.7 on the basis of morphostratigraphical, lithological and geochemical evidence from the study area. In this chapter a multiparameter dating approach is used to provide relative ages for landform and lithostratigraphical units and to correlate stratigraphies from different geographical locations. Radiocarbon dates on shells from fossiliferous sediment units provide absolute ages for events within the practical range of radiocarbon dating, and minimum ages for events beyond this range. Amino acid ratios in fossil shells are used to construct a relative dating framework for glacial and marine events. Absolute ages, based on amino acid ratios, are tentatively proposed, and discussed in the context of aminozones previously developed for northern Labrador and Baffin Island. Pedogenic indices which demonstrate the degree of soil development are used to relatively date moraines and their associated drift sheets. A tentative absolute soils-based chronology is discussed for the study area.

6.2. Radiocarbon dates

Six radiocarbon dates are reported for shell samples from fossiliferous units in the study area (Table 6-1). The oldest dates of 38900 ±1420 years BP and 40600 ±1640 years BP (GSC-4204) were obtained from the outer and inner fractions, respectively, of a thick specimen of *Hiatella arctica*, collected from the glaciomarine diamicton of unit VA1 in southern Valley of the Flies (see Section 3.4.4, Fig. 6-1). *Hiatella arctica* shells from unit VA2 which stratigraphically overlies the diamicton, provided a date of 37800 ±1610 years BP (GSC-4189). This unit is interpreted to comprise the reworked and winnowed sediments from the underlying diamicton, and therefore the overlap in dates for these units is consistent with this interpretation.

Table 6-1: Radiocarbon dates reported for fossiliferous sediment units in the study area. See Figs. 6-1 & 6-2 for location of units.

Radiocarbon dates			
Unit	Species	Lab#	^{14}C Date
VA1	*Hiatella arctica*	GSC-4204	40600 ±1640
VA1	*Hiatella arctica*	GSC-4204	38900 ±1420
VA2	*Hiatella arctica*	GSC-4189	37800 ±1610
VD1	*Mya truncata*	GSC-3950	33700 ±710
AA3	*Balanus balanus*	GSC-4161	9170 ±100
AB1	Gastropod fragment	TO-305	9820 ±70

Radiocarbon dating of *Mya truncata* specimens from unit VD1 gave an age of 33700 ±710 years (GSC-3950) (Table 6-1). This unit is also considered to be part of the diamicton, but on the basis of the above date, it appears to postdate the deposition of unit VA1, although not necessarily unit VA2. The four dates reported above, while finite, lie beyond the practical limit of radiocarbon dating of shells (30,000 years according to Terasmae, 1984), and therefore should be considered minimum dates for the units they represent. Clarke *et al.* (1972) demonstrated that shells recovered from the 'Salmon River sands' in Nova Scotia which produced radiocarbon ages of 38,000 years BP are, in fact Sangamon interglacial age, based on stratigraphic and palaeoclimatological evidence. They suggested that such apparently finite radiocarbon dates should be regarded as minimum estimates until they can be verified by dates on associated carbonaceous material, or checked by alternative methods.

Only 1% modern carbon is required to produce an apparently finite age for material that is actually infinitely old. The outermost part (20%) of the shell specimens recovered from the study area were leached in order to avoid such contamination. Shells from unit VD1 had their outer and inner fractions dated (GSC-4204), and although both dates are in good agreement, they differed by almost 12% including the full range of error, and therefore may reflect differential replacement by modern carbon.

Two shell samples were radiocarbon dated from sections in Adams Lake valley (Table 6-1, Fig. 6-2). At 23 m aht, a river eroded section (Site AA) is interpreted to reveal fossiliferous

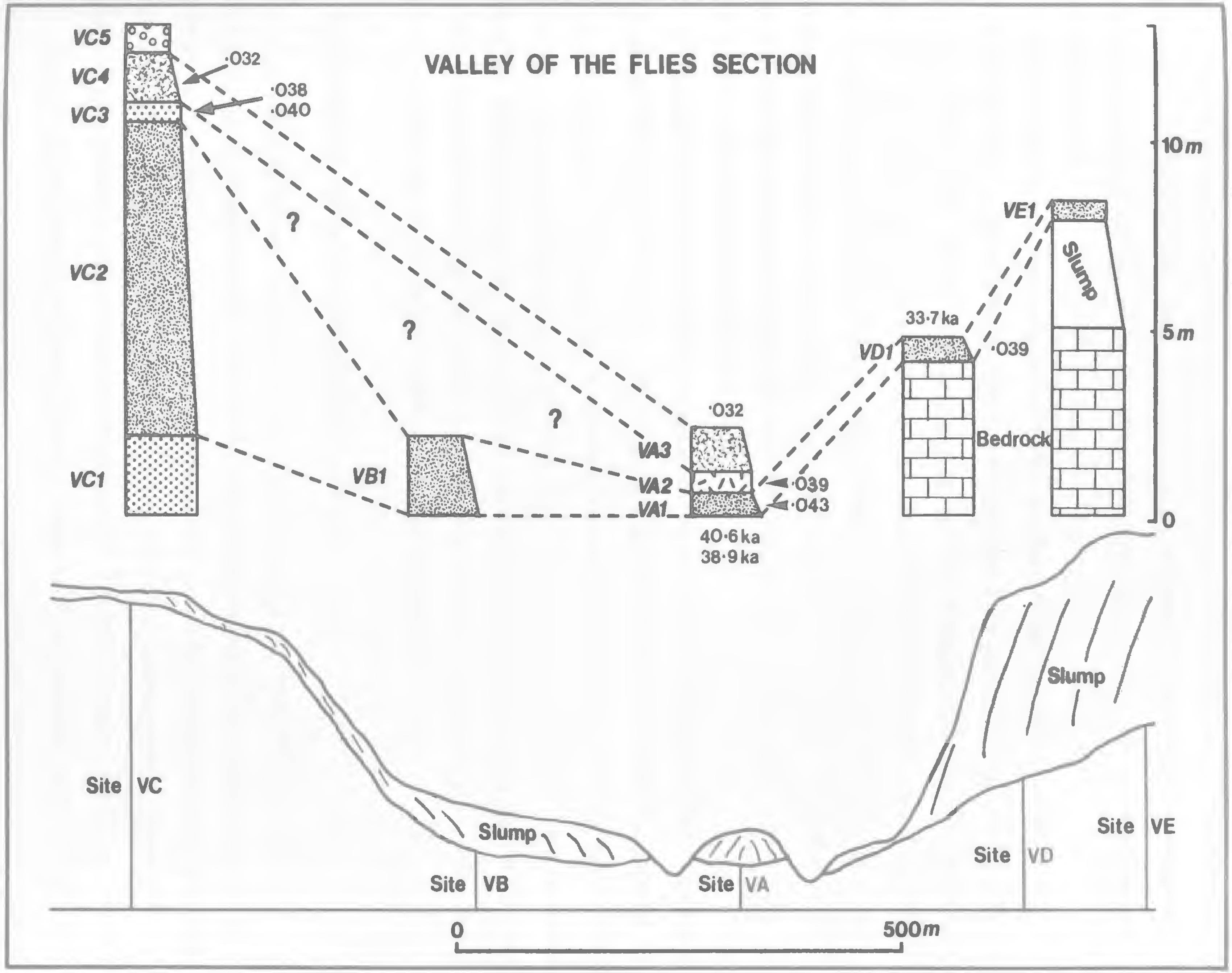

Figure 6-1: Stratigraphic sections of fossiliferous units with their radiocarbon dates and total amino acid ratios in Valley of the Flies.

marine clays overlying beach sands, with the former dated at 9170 ±100 years BP (GSC-4161). In the bottom of Adams Lake valley, at 8-10 m aht, a gastropod fragment from a bluish-grey muddy unit (Site AB) was radiocarbon dated at 9820 ±70 years BP (TO-305) using accelerator mass spectrometry.

The younger date at approximately 9 ka BP is related to the transgressive event which deposited the marine clay at Site AA, and formed Shoreline D in the study area. The muddy nearshore deposit at site AB is underlain by a diamict which is similar in texture and colour to that in Valley of the Flies. Unit AB1 may consist of reworked 'old' shells from this underlying unit because of the similarity between the assemblage of shell fauna collected from it and those identified from the Valley of the Flies diamicton (see Sections 3.4.3 & 3.4.4). The radiocarbon date reported here confirms that unit AB1 contains Holocene shells, but these need not be the only, nor even the majority of shells in the unit.

A limited absolute chronology of events in the study area can be reconstructed on the basis of these six ^{14}C dates. In summary, radiocarbon dating suggests that the Adams Lake glacial phase which deposited a glaciomarine diamicton in Adams Lake valley and southern Valley of the Flies, occurred $\geq$ 38 ka BP (mean value for the four ^{14}C dates from the diamicton and reworked unit in the Valley of the Flies section). Marine depositional events at approximately 9 ka BP and 10 ka BP were responsible for the formation of Shoreline D and the deposition of a marine clayey unit which drapes the lower part of Adams Lake valley, respectively.

6.3. Amino acid racemization results

Over the last fifteen years aminostratigraphic correlation has been used as a relative chronostratigraphic framework for Quaternary events on Baffin Island [see Miller (1985) for an overview of the multidisciplinary dating program carried out by the Institute of Arctic and Alpine Research, University of Colorado]. More recently amino acid racemization dating has been employed in Quaternary studies in northern Labrador (Clark, 1984; Johnson, 1985).

This technique involves the quantification of amino acid racemization which accompanies fossilization of organisms. Amino acids occur as the building blocks of proteins and as structural components in all living organisms. During diagenesis, these proteins hydrolize to release low molecular weight polypeptides and free amino acids. Loss of amino acids from the fossil can occur

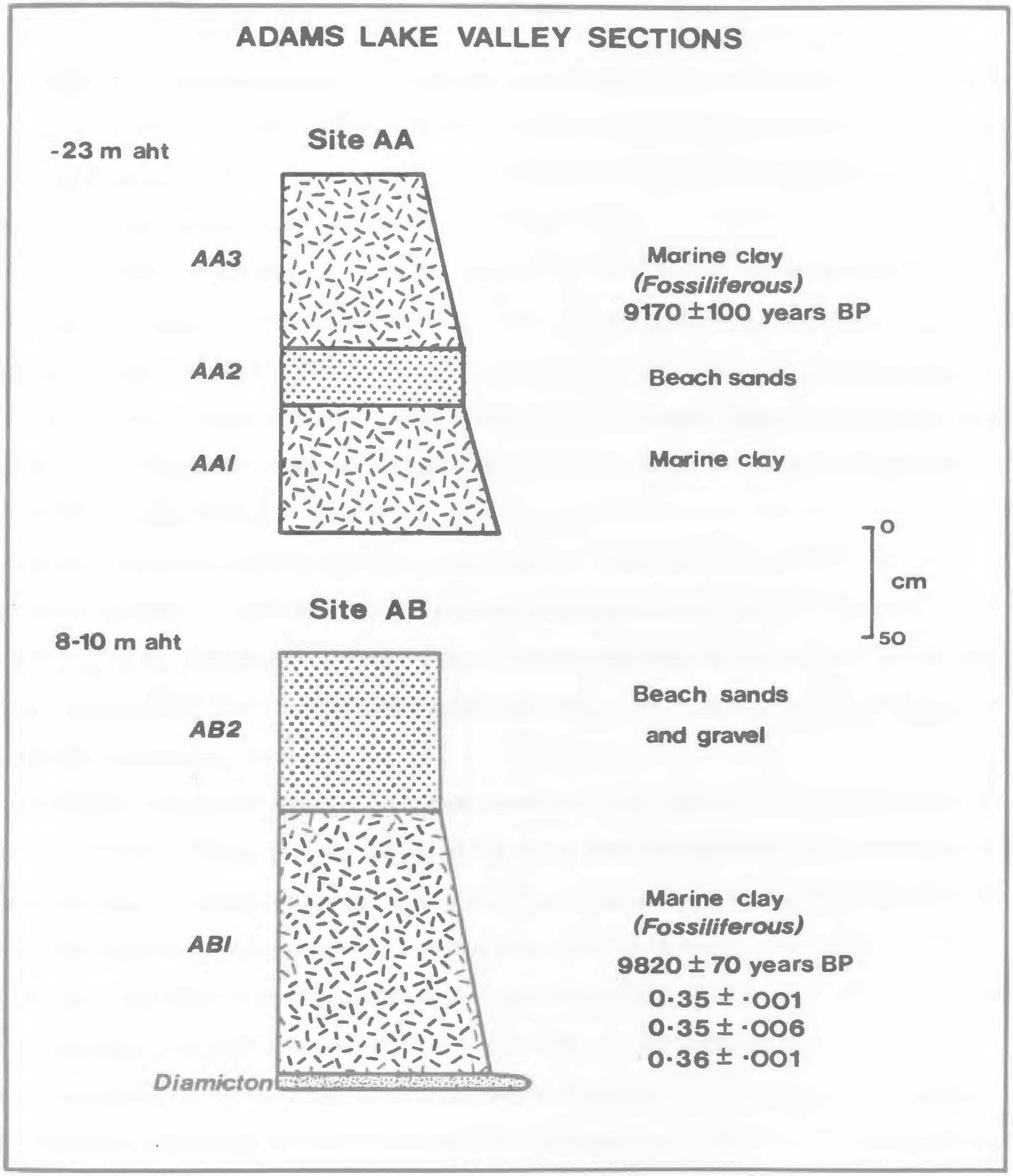

Figure 6-2: Stratigraphic sections of fossiliferous units with their radiocarbon dates and total amino acid ratios in Adams Lake valley.

by decomposition, diffusion, or leaching. In addition to these reactions, conversion of the original L-amino acids (*levo*, or 'left-handed') into a mixture of D- (*dextro*, or 'right-handed') and L-amino acids occurs. Molecules containing one chiral, or asymmetric carbon are called chiral molecules and exist in two stereoisomer forms called enantiomers (mirrow image isomers). During the racemization process, L-enantiomers are converted to D-enantiomers. Amino acids that contain two chiral carbons may exist as four stereoisomers, two enantiomers and two diastereomers (non-mirrow image isomers). However, during diagenesis only one of the chiral carbon atoms undergoes interconversion (Hare *et al.*, 1980; Wehmiller, 1982).

The D/L ratio determined for a fossil is regarded as a measure of the age of the fossil (Rutter *et al.*, 1985 and references cited therein). This value is usually calculated for the **total** amino acid mixture and for the **free** amino acids in a sample. The rate of racemization (or epimerization[30]) in a fossil is dependent upon the Effective Diagenetic Temperature (EDT), the sample type, the amino acid analysed, and the effects of contamination, or other diagenetic factors (Rutter *et al.*, 1985).

A major limitation of the technique results from the temperature dependency of the racemization process; the racemization rate doubles for approximately each 4 ° C increase in EDT (Rutter *et al.*, 1985). Consequently the technique is most accurately employed for the correlation and relative age dating of equivalent strata which have experienced similar temperature histories and diagenetic conditions.

Racemization rates are strongly dependent upon the type of material in which the amino acid is imbedded. For example, rates are faster for wood, intermediate for bones and slowest for shells. Furthermore, racemization rates for a given fossil material are highly species-dependent. This has been repeatedly demonstrated in both molluscs (Miller & Hare, 1975; Miller & Hare, 1980; LaJoie *et al.*, 1980; Wehmiller, 1980) and foraminifera (King & Hare, 1972; King & Neville, 1977). In deposits with mixed assemblages of fragmented, non-identifiable shells, amino acid ratios are speculative as relative age indices. However, Andrews *et al.* (1985) have calculated linear discriminant functions which characterize different species on the basis of amino acid ratios and which can taxonomically identify unknown species. This approach may be adopted to ensure that correlation and dating of deposits are carried out with the same species.

[30] The term *epimerization* is generally used to describe the configurational change involving compounds, such as isoleucine, with two asymmetric carbon atoms.

Amino acids racemize at different rates; for instance, aspartic acid racemizes rapidly whereas leucine and isoleucine have much slower rates. Correlation studies are most successful where there is a wide variation in fossil amino acid D/L ratios between the units to be correlated, and therefore the characteristic racemization rates of specific amino acids can be used to discern closely-spaced events. In high latitudes where low temperatures severely retard racemization rates, aspartic acid is used to determine D/L ratios. Similarily, leucine and isoleucine are most useful for fossils from warmer climates.

6.3.1. Relative age estimates based on amino acid ratios

Aminostratigraphy is at a very early stage in northern Labrador and therefore correlation must be based on Baffin Island aminostratigraphy and its associated chronological interpretations. The extrapolation of amino acid ratios over such a distance (400 km) may only reveal tentative correlations, although Miller (1985) proposed that such a correlation is valid considering the similarity in mean temperatures between the two regions. However, the traditional assumption that closely-spaced localities with the same present temperature can be interpreted in terms of similar Effective Diagenetic Temperatures (EDT), has been questioned by the results of Wehmiller and Belknap (1982) who discovered discrepancies between D/L ratios in *Mercenaria* and U-series coral ages along the Atlantic coast of the United States. McCartan *et al.* (1982) attributed these discrepancies to steep temperature gradients and local diagenetic conditions during glacial intervals. Clark (1984) justified the correlation of Baffin Island aminostratigraphy with his amino acid ratios from northern Labrador on the basis of the statistical significance of total alloisoleucine/isoleucine (aIle/Ile) ratios from radiocarbon-dated Holocene deposits in both areas. However the ranges of Clark's total aIle/Ile ratios for samples radiocarbon dated at 9-34 ka BP overlap, and therefore are not statistically significant. Consequently, correlations proposed by Clark are considered speculative.

Shell samples collected in the study area were forwarded to W.D. McCoy at Morrill Science Center, University of Massachusetts, for amino acid analysis. The reported aIle/Ile ratios for the free-fraction and total hydrolysates are given in Table 6-2(A). The question marks (?) in this table indicate samples where the free aIle was very difficult to measure accurately, but was definitely not 'not detectable' as is common in Holocene samples (McCoy, personal communication, 1986).

The ranges of total aIle/Ile ratios for shell samples from fossiliferous units in Valley of the Flies and Adams Lake valley are plotted on Figure 6-3. Three groups of samples are identified on the basis of their characteristic total aIle/Ile ratios and on the stratigraphical significance of the units from which the shells were sampled. There is a general trend of decreasing ratios up-section for the Valley of the Flies samples (Fig. 6-1), although the range of ratios overlap in some cases.

The stratigraphy of the sample sites are described in Section 3.4. Sedimentological, textural and lithological evidence suggests the following sequence of events for the deposition of sediments in southern Valley of the Flies:

1. Deposition of a marine sand unit. *Unit VC1*
2. Deposition of a glaciomarine diamicton during local ice advance through Adams Lake Valley and into southern Valley of the Flies. *Units VA1, VB1, VC2, VD1 and VE1*
3. Deposition of marine sand unit, possibly reflecting ice recession and deglaciation of site. *Unit VC3*
4. Deep gully cut through sediments - origin uncertain (subaqueous/subaerial?).
5. Subaqueous reworking and winnowing of diamictic sediments exposed in gully. Deposition of a gravel and shell lag unit in base of gully. *Unit VA2*
6. Subaqueous debris flow mantles underlying sediments. Occassional shell fragments found in fining upward sequence (*unit VA3*), but coarse shell lens concentrated along upper part of debris flow unit (*unit VC4*).
7. Deposition of beach sand and gravel over section. *VC5*
8. Emergence

Amino acid ratios provide a relative chronological framework for the above sequence of events and allows positive correlation of texturally-similar, fossiliferous units. *Hiatella arctica* specimens from the diamicton in Valley of the Flies, units VA1 and VD1, gave total aIle/Ile ratios of 0.043 $\pm$.005 and 0.039 $\pm$.004, respectively. *Hiatella arctica* specimens from the reworked deposit (unit VA2) overlying the diamicton at site VA have similar ratios to shells in the diamicton - 0.039 $\pm$.005. The marine sand overlying the diamicton at site VC contains thin, fragile shells in contrast to the thick, robust shells of the diamict, and therefore are not considered to be reworked from it. These shells gave total aIle/Ile ratios of 0.04 $\pm$.005 and 0.038 $\pm$.006. Specimens of *Mya truncata* from the debris flow unit, VA3 and VC4, have ratios of 0.032 $\pm$.001

Table 6-2: List of reported and predicted aIle/Ile ratios and associated radiocarbon dates for shell samples collected in northern Labrador and Baffin Island.

(A) AIle/Ile ratios and Radiocarbon dates for shell samples

Site	Lab #	Free	Total	Lab #	^{14}C Date	Species
VA1	AGL-471	0.18±.02	0.043±.005	GSC-4204	40600±1640	*Hiatella arctica*
VA1	AGL-471	0.18±.02	0.043±.005	GSC-4204	38900±1420	*Hiatella arctica*
VA2	AGL-470	0.17±	0.039±.005	GSC-4189	37800±1610	*Hiatella arctica*
VA3	AGL-478	0.18±.02	0.032±.001	***Predicted***	***26500±***	*Mya truncata*
VC3	AGL-472	0.17?	0.040±.005	***Predicted***	***36500±***	*Mya truncata*
VC3	AGL-473	0.20±.05	0.038±.006	***Predicted***	***34000±***	*Hiatella arctica*
VC4	AGL-474	0.17±.03	0.032±.003	***Predicted***	***26500±***	*Mya truncata*
VD1	AGL-479	0.22?	0.039±.004	GSC-3950	33700±710	*Hiatella arctica*
						Mya truncata
AA3	***Pred.***	—	***0.018±***	GSC-4161	9170±100	*Balanus balanus*
AB1	AGL-475	0.17	0.035±.001	***Predicted***	***30300±***	*Mya truncata*
AB1	AGL-476	0.18	0.035±.006	***Predicted***	***30300±***	*Balanus*
AB1	***Pred.***	—	***0.019±***	TO-305	9820±70	Gastropod fragment
AB1	AGL-477	0.14	0.034±.001	***Predicted***	***29000±***	*Hiatella arctica*

(B) AIle/Ile ratios and associated radiocarbon dates for shell samples collected by Clark (1984) from northern Labrador

New	AAL-3558	0.14±.008	0.016±.003	GX-8240	$34200\pm^{2100}_{1600}$	*Mya truncata*
Old	AAL-3558	0.16±.03	0.026±	GX-8240	$34200\pm^{2100}_{1600}$	*Mya truncata*
New	AAL-3443	0.15±.03	0.018±.005	GX-8241	$28200\pm^{1200}_{1000}$	*Mya truncata*
Old	AAL-3443	0.22±	0.042±	GX-8241	$28200\pm^{1200}_{1000}$	*Mya truncata*
Old	AAL-2437	0.18±.03	0.027±	GX-8241	$28200\pm^{1200}_{1000}$	*Mya truncata*
New	AAL-3442	Not detect.	0.013±.001	GX-9293	9110±470	*Mya truncata*

(C) Mean amino acid ratios and associated dates determined by Miller (1985) and Szabo *et al.*, (1981) for aminozones on Baffin Island

	Free	Total	^{14}C Date	U-series date	Amino acid date	Aminozones
New	Not detect.	0.0129±.0007	8-11 ka	$\geq$ 6 ka	13 ka	Eglinton Member
New	0.14±.02	0.020±.004	$\geq$ 40 ka	---	44 ka	Loks Land Member
New	0.30±.02	0.029±.004	$\geq$ 54 ka	$>$ 70 ka	110 ka	Kogalu Member

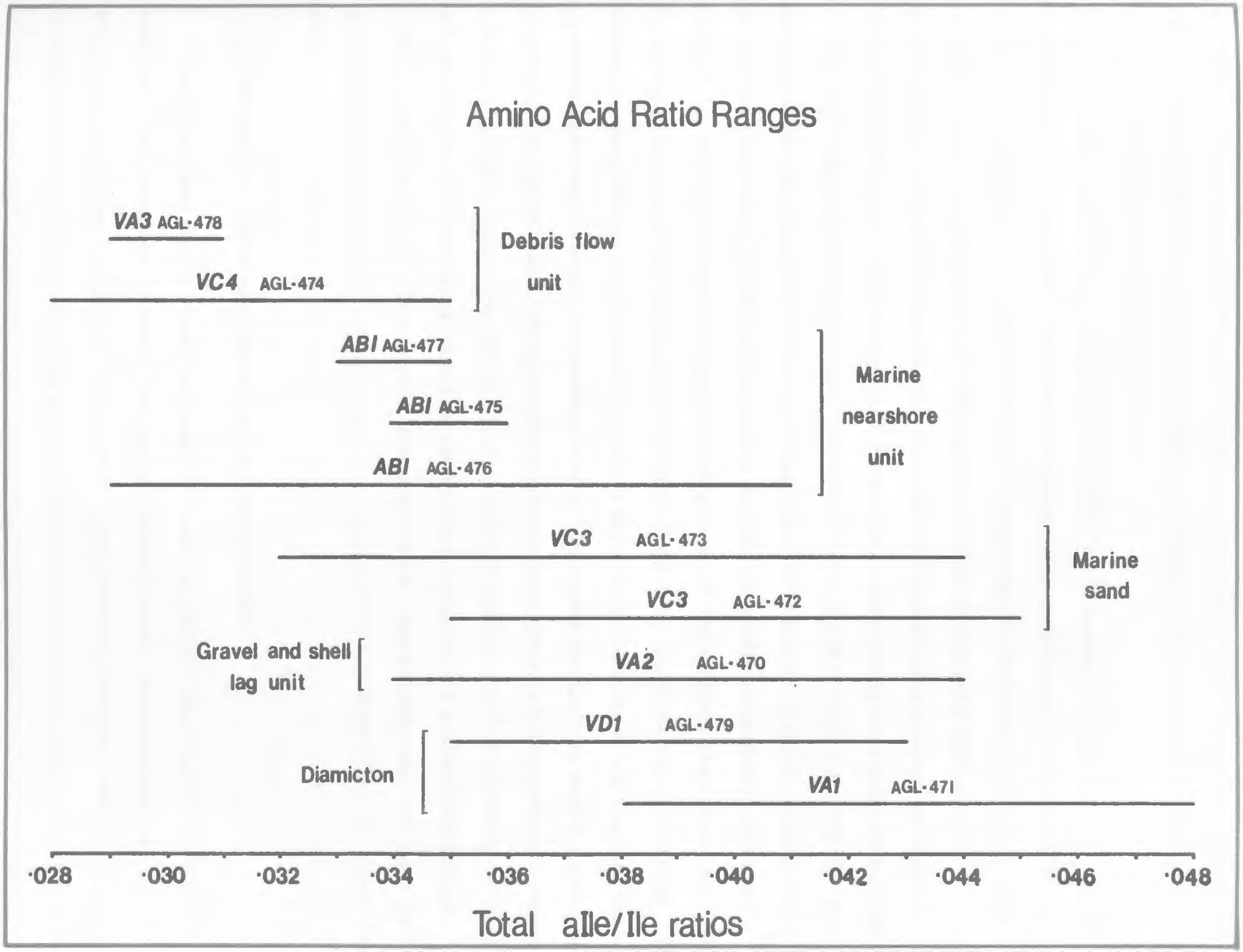

Figure 6-3: Diagram showing the plotted ranges of total aIle/Ile ratios for shell samples from the study area.

and 0.032 ±.003, respectively. Three shell samples were analysed from unit AB1 in Adams Lake valley. They gave ratios of 0.035 ±.001, 0.035 ±.006 and 0.034 ±.001 for *Mya truncata*, *Balanus* sp. and *Hiatella arctica*, respectively.

These total aIle/Ile ratios support the tentative ^{14}C chronology for local depositional events in southern Valley of the Flies. The oldest, fossiliferous unit is the marine diamicton which postdates the deposition of the marine sand that stratigraphically underlies it. The ages of the sand unit overlying the diamicton and the diamicton are not significantly different, but sedimentological and fossil evidence suggest that the sand unit represents a more quiescent depositional environment. The diamicton in Adams Lake valley is suggested to be younger than the equivalent diamicton in Valley of the Flies based on D/L ratios of shells reworked into the overlying marine clay [unit AB1]. This may suggest gradual recession of the ice margin west along Adams lake valley. As the ice front retreated, distal depositional environments may have experienced a decrease in sediment influx and energy regime, and therefore the well-sorted, fining-upward, sand unit overlying the diamicton in Valley of the Flies may reflect such conditions. The subsequent deep gullying of the section in Valley of the Flies is unexplained, but may represent a major submarine erosive event, or subaerial erosion. The gravel and shell lag deposit overlying the diamicton along the base of the gully represents reworked and winnowed diamictic sediment as evidenced by the similar total aIle/Ile ratios of the shell specimens and the similarity in the shell assemblages. The overlying subaqueous debris flow represents the youngest fossiliferous deposits in the Valley of the Flies section. The shell samples collected from this unit have the lowest total aIle/Ile ratios in the section, and may even postdate the deglaciation of Adams Lake valley.

6.3.2. Numeric age estimates based on amino acid ratios

Regression analysis was carried out using the four shell samples from sites VA and VD which had radiocarbon dated control. The purpose of this analysis was to estimate the age of those samples which had no dating control, and to predict the aIle/Ile ratios for the two Holocene dated samples based on the relationship between amino acid ratios and ^{14}C dates in the control samples. A total aIle/Ile ratio of 0.0111 ±.001 (AAL-3448) determined by Miller (1985) for modern specimens of *Hiatella arctica* was also used in the regression. The total hydrolysate ratios were used in preference to the free-fraction ratios as the former displayed more consistent and computable results. The results of the regression analysis are given in Table 6-2(A), and plotted on Figure 6-4.

It is stressed that the absolute dates predicted from the control samples are extremely tentative, based on radiocarbon age determinations which are probably minimum estimates, and on D/L ratios which are mean values. Radiocarbon dates for the Valley of the Flies diamicton suggest that it was deposited $\geq$ 38 ka BP. The predicted age of the marine sand overlying the diamicton is 34-36.5 ka, suggesting that this site was deglaciated by this time. Predicted dates for the Adams Lake valley diamicton are between 29 ka BP and 30.3 ka BP, or approximately 9 ka following deposition of the diamicton in southern Valley of the Flies. The subaqueous debris flow unit which mantles the Valley of the Flies section is predicted to date 26.5 ka BP.

The four control samples have a mean ^{14}C date of 38$\pm$ ka BP and a mean aIle/Ile ratio of 0.19$\pm$ (free) and 0.04$\pm$ (total). The free-fraction aIle/Ile ratio is similar to, but slighter greater than those determined for the Loks Land Member on Baffin Island (0.14$\pm$), and the Iron Strand diamicton north of the study area (0.14$\pm$)[31] (Table 6-2(B) & (C)). However, the aIle/Ile ratios in the total hydrolysates of the four samples are much higher than those reported for Miller's (0.02$\pm$) and Clark's (0.016$\pm$) ^{14}C dated equivalents. Such a contrast in ratios is not considered to be the result of intergeneric comparisons, that is *Hiatella arctica* versus *Mya truncata*. Such differences appear to be insignificant, as illustrated in data from the study area (compare AGL-472 and AGL-473 from sample site VC4; Table 6-2(A) and Fig. 6-4). It is also unlikely that these four control samples are older than the Loks Land Member, or the Iron Strand diamicton because of the similarity in their free-fraction ratios, and the significant difference between the latter and the mean ratio calculated for the older Kogalu Member (0.3$\pm$) on Baffin Island (Table 6-2(C)).

McCoy (personal communication, 1986) suggested that these higher total aIle/Ile ratios in the study area may be due to warmer summer temperatures in the Nachvak Fiord region. This hypothesis was tested by using the age-prediction equation formulated by Miller (1985) to compare the reported ^{14}C age of sample VD1 (AGL-479) with a predicted age based on an

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[31] Two sets of ratios are given for Clark's samples as a modification of the sample preparation technique revealed that the 'old' method overestimated amino acid ratios (see Miller, 1985). The 'new' method is considered to be more accurate and reliable and those results are quoted here (Clark, 1984).

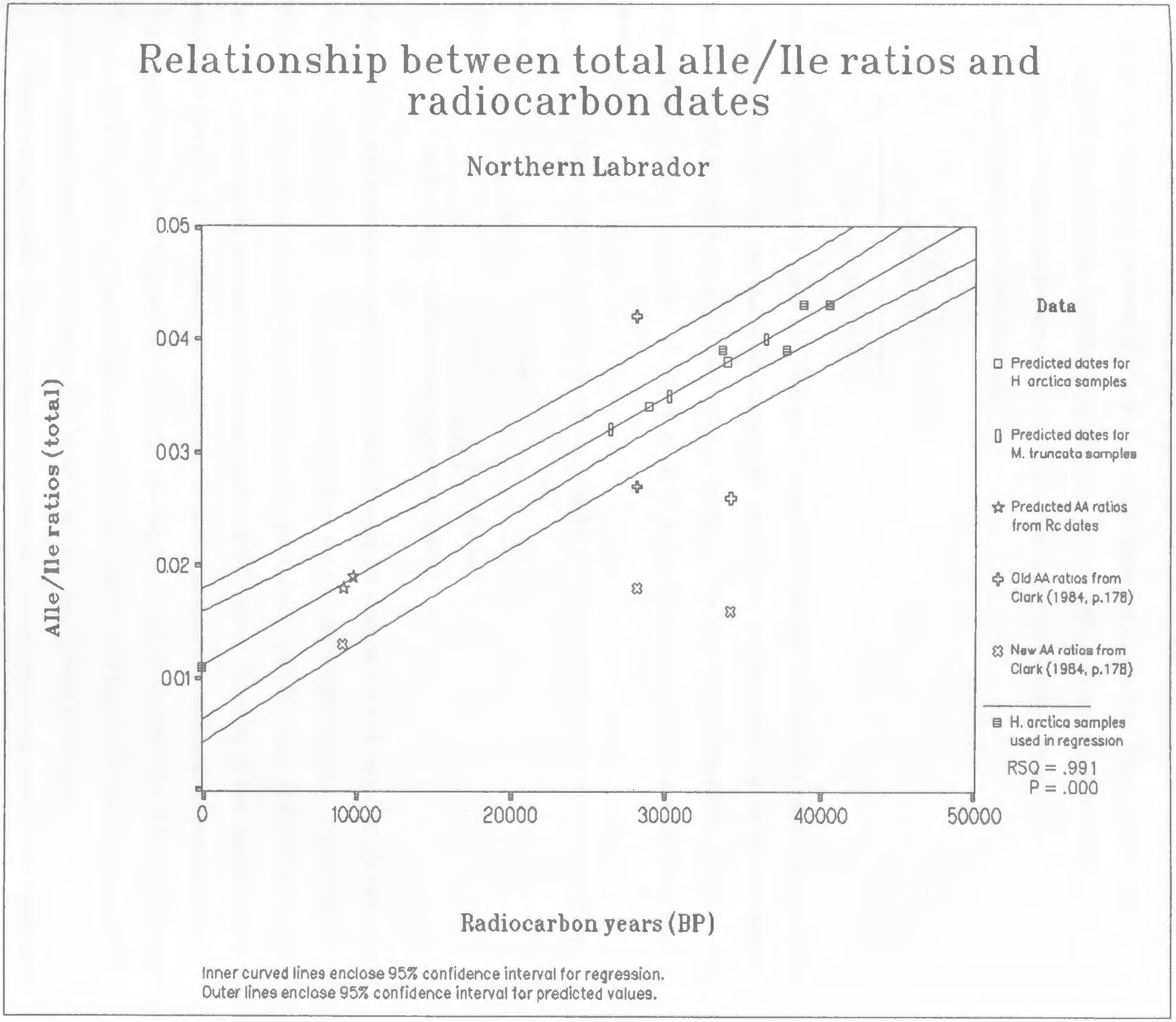

Figure 6-4: Relationship between total alle/Ile ratios and radiocarbon dates for shell samples from the study area.

Effective Diagenetic Temperature of -9 °C. This was the temperature that Miller (1985) considered to be the best estimate, bearing in mind the age of the Baffin Island aminozones. Clark (1984) calculated a similar EDT of -8.3 °C for northern Labrador. The equation was calibrated for the isoleucine epimerization reaction rate in *Mya truncata* (Miller, 1985), and therefore the following results involving *Hiatella arctica* are estimated values.

Using the age-prediction equation:

$$t = \frac{\ln\left(\dfrac{1 + D/L}{1 - 0.77\, D/L}\right) - 0.0194}{1.77\ 10^{\left(16.45 - \frac{6141}{T}\right)}}$$

where D/L is the aIle/Ile ratio of 0.039 and T is the EDT of 264 °K, then the predicted age of the sample is 180.6 ka BP. As stated previously, the age of this sample should not be greater than that estimated for the Loks Land Member ($\geq$ 40 ka) or the Iron Strand diamicton (33-50 ka BP, Clark (1984, p.184)). Therefore, the EDT of -9 °C is considered too low for samples from the study area.

A local EDT value was calculated using Miller's (1985) temperature-prediction equation:

$$T = \frac{6141}{16.45 - \log \dfrac{\ln\left(\dfrac{1 + D/L}{1 - 0.77\, D/L}\right) - 0.0194}{1.77\, t}}$$

where D/L is the predicted total aIle/Ile ratio of 0.018 for sample AA3 (Table 6-2(A)) and t is the ^{14}C date of 9170 years BP. This sample was chosen for the calculation because it has a genuine finite ^{14}C date, and because it has a similar age to the sample used by Clark (1984) (9110 ±470 years BP), on which he calculated his representative EDT value. Solution of the equation gave an estimated EDT of -1 °C. Using the traditional assumption that this Holocene temperature is representative for older temperature histories, the predicted age of sample VD1 was recalculated with an EDT of 272 °K. This resulted in an estimated age of 37.4 ka which is more compatible with the assigned ages of the Loks Land Member and the Iron Strand diamicton.

The mean age of the Valley of the Flies diamicton, as predicted from amino acid ratios and an estimated EDT of 272 °K is 38.7 ka BP. The corresponding diamicton in Adams Lake valley has a predicted age of 31.7 ka BP, representing a 6 ka age difference. The marine sand unit overlying the diamicton in Valley of the Flies is estimated to be 37.4 ka BP, while the debris flow unit is dated at 28.1 ka BP. These ages correspond closely to those derived from a simple regression analysis of the amino acid ratios with the ^{14}C dated control samples.

As mentioned previously, these predicted dates are extremely tentative. The calculated EDT of -1 °C for the study area is much higher than Clark's (1984) and Miller's (1985) values and therefore implies a very steep post-depositional temperature gradient north of Nachvak Fiord. This gradient may be realistic between Baffin Island and northern Labrador, but it cannot be explained for the 50 km distance between Kangalaksiorvik Fiord and Nachvak Fiord. There appear to be gross inconsistencies in total aIle/Ile ratios between these two areas which require further investigation beyond the scope of this study.

6.4. Pedological Studies

6.4.1. Relative soil-based chronologies

Relative dating techniques carried out on Quaternary deposits assume that post-depositional modification is mainly dependent on the time elapsed since deposition. The development of soil on Quaternary deposits is influenced by five factors, described by Jenny's (1941) soil equation, such that

$$\textit{Soil Development} = f\,[\,T, C, P, V, R\,]$$

where

T = time
C = climate
P = parent material
V = vegetation
R = topography

If the four factors, climate, parent material, vegetation and topography, are considered constant over time then the degree of soil development should reflect the time since initiation of soil formation. All studies using soil properties as relative or absolute age indices rely on this assumption, although Locke (1985) suggests that it is certainly false in some cases. However as he further points out, the technique appears to work, at least empirically, and therefore "with further refinement of the technique.....and with a continuous awareness of the operational assumptions, weathering studies [pedological studies] will continue to be a useful tool for terrestrial Quaternary stratigraphy...."

The justification for employing this assumption in pedological studies is well documented, both in technique and in results obtained, particularly in the eastern Canadian Arctic (Isherwood, 1975; Andrews and Miller, 1976; Birkeland, 1978; Evans and Cameron, 1979; Clark, 1984; Evans and Rogerson, 1986). In each study there appears to be a significant relationship between soil properties and the perceived age of the deposit.

The most commonly used age-dependent properties of soil are (1) depth of solum; (2) the intensity of oxidation of the B horizon [using the colour-development equivalents of Buntley and Westin (1965)]; (3) various properties associated with chemical alteration in the soil profile (e.g. pH, % organic matter, clay mineralogy, particle size distribution, unstable mineral abundance and weathering, etching of minerals and soil chemistry analysis). All these properties have varying degrees of potential as relative dating techniques; however, the depth of solum as an indication of the extent of oxidation appears to be a relatively robust age-dependent property, and combined with a fast and simple field technique, it has been widely recommended by previous users (Evans and Cameron, 1979; Andrews and Miller, 1980; Clark, 1984; Evans, 1984)

The depth of solum was recorded in this study from soil pits located on selected landforms. Pits were dug, where possible, to below the Cox/Cn boundary and each horizon classified according to the nomenclature of Birkeland (1974, 1978). Colours (moist) described for the horizons were judged against a Munsell Soil Colour Chart (Table D-1, Appendix D).

Comparison of pedological data between field areas is necessary if relative chronologies are to be constructed for glacial events in a region such as the Torngat Mountains of northern Labrador. In doing so however, it is critical that the original assumption of the technique is not violated, and hence false correlations proposed. To reduce possible error in extrapolating and comparing results, it is essential that the factors affecting soil formation are similar in selected sites and that these same factors have remained constant at individual sites. The following guidelines were adhered to with regard to site selection in the study area.

1. Topographical variation between sites was reduced by restricting soil studies to the crests of moraines. This ensured well-drained sites, free from the influx or removal of material by slope solifluction. Areas showing signs of cryoturbation were avoided.

2. Primarily, only tills with similar composition were compared. According to Birkeland (1974) and Bockheim (1980), the rate of increase in solum thickness is positively correlated with clay content of the parent material. In addition, till lithology and geochemistry may directly influence the rate of oxidation. Bockheim (1980) however, found that the colour development of the B horizon approaches a maximum through

time and then decreases thereafter. This phenomenon may be due to a depletion in weatherable iron-bearing minerals in the solum (Yaalon, 1971).

Differentiation of glacial episodes based on the degree of soil development is therefore restricted if till composition varies dramatically with each glacial event. Evans (1984) reported anomalous solum depths in his soil-based glacial chronology of the Selamiut Range, Nachvak Fiord, and concluded that variation in till composition was responsible for these results. Hence it is important to consider parent materials in any comparison of soil property data.

As mentioned in Section 3.4, till texture in Tinutyarvik and Naksaluk drifts is finer, with a higher clay content than elsewhere in the study area. As the rate of soil formation increases with increased clay content, then age determinations based on pedogenic indices derived from Tinutyarvik and Naksaluk drifts will be overestimated relative to age determinations from other drift sheets.

3. Variations in vegetation cover on selected sites will influence the rates of soil formation through time. Although this factor may have minimal variation on a local basis, extrapolation across vegetational boundaries will affect the validity of results obtained. All of the soil sample sites in this study were cut where vegetation was noticeably lacking, or composed of grass or other low vegetation cover.

4. The effects of climate on soil formation are such that any comparison of pedological data should be restricted to study areas with similar climatic regimes. Rogerson *et al.* (1986*b*) have suggested that soil development rates in part of the Torngat Mountains may be significantly greater than many other Arctic localities, principally due to a longer growing season, warmer temperatures and higher precipitation values. Bockheim (1980) demonstrated a positive correlation between rate of change in oxidation depth and solum thickness with mean annual temperature. Precipitation will also affect soil formation but, as Birkeland (1978) has noted, in those areas where much of the precipitation falls as snow, exposed wind-swept sites (i.e. moraine crests) will receive minimal moisture through precipitation. Bockheim (1979) has suggested that altitude and proximity to the ocean will dramatically influence soil properties. Clark (1984) did not observe such a phenomenon in his pedological data from an area only 50 km to the north of Nachvak Fiord, although Evans and Rogerson (1986, 1987) recognized a significant decrease in solum depth in soils formed above the Ivitak moraine.

Table 6-3 gives the mean values for cambic B horizon thickness recorded from soil pits dug on those moraine crests that were considered suitable for pedological investigation in the study area. These results suggest that the moraine system in Kammarsuit valley east is the oldest of the five moraines, having 34 cm thickness of cambic B horizon in its soil profile. Kammarsuit valley west end moraine is considered younger than the former moraine system because it has a less developed soil profile represented by a thinner cambic B horizon. This conclusion is consistent with the evidence of former glacial lakes in the two valleys. It was previously suggested that the glacial lake in Kammarsuit valley west was dammed by ice in Kammarsuit valley east and that it was the retreat of this ice margin that caused the lake to empty (see Section 3.2.4). During ice recession a proglacial lake existed in Kammarsuit valley east, dammed by the end moraine at the mouth of the valley. Although there is no direct evidence to suggest that this proglacial lake formed prior to the lake in Kammarsuit valley west, it is a more likely sequence of events in the glacial history of the area. This suggestion implies that ice initially retreated from the moraine in Kammarsuit valley east and that moraine consequently displays a more developed soil profile.

Table 6-3: Mean values for cambic B horizon thickness on selected moraine crests in the study area.

Mean values for cambic B horizon thickness		
Moraine	Thickness	Observations
Adams Lake	29 cm	2
Kammarsuit valley west	29 cm	1
Kammarsuit valley east	34 cm	1
Quartzite	22 cm	5
Naksaluk	21 cm	8

Adams Lake moraine may be coeval with Kammarsuit valley west end moraine, based on the comparable thicknesses of cambic B horizon. On the basis of sediment/landform considerations (Chapter 3) both moraines, plus the Kammarsuit valley east moraine system, represent local glaciation in the study area and it appears likely that deglaciation, and subsequent initiation of soil development commenced during a similar time period.

In contrast, on the same bases, Quartzite and Naksaluk moraines were deposited during the last regional ice advance into the study area [*Nachvak glacial phase*; Evans and Rogerson, 1986]. The similarity in the degree of soil development on the two moraines, based on 13 soil pits, reflects the general retreat of regional ice from the area. It appears likely that ice receded from the distributary valleys (Adams Lake and Tinutyarvik valleys) prior to deglaciation in the fiord basin (Naksaluk valley was probably backfilled by the main fiord ice stream), as there is slightly greater mean soil development on Quartzite moraine. The fine clayey texture of regional drift is more conducive to soil development and therefore the reported differences in cambic B horizon thickness between Quartzite/Naksaluk moraine and moraines from elsewhere in the study area may be more significant than the actual value observed. Consequently the relative age difference between local and regional tills may be greater than the observed difference in cambic B horizon development.

6.4.2. Absolute soil-based chronologies

Once a relative chronological framework has been developed for an area, there is strong temptation to transform the results into an absolute chronology, particularly in situations where radiocarbon dating control is limited, or unavailable, or where the age of deposits are beyond the limits of the radiocarbon dating method. Two types of transformation can be used to estimate the age of a deposit by means of soil development indices.

In the first case, the measurable soil property is regarded as an indication of the duration of weathering of the parent material. Therefore an average rate of soil development is estimated by dividing the difference in soil property and parent material (*e.g.* depth of solum) by the absolute age of subaerial exposure of the parent material. The age of subaerial exposure of another parent material can be determined, for example, by dividing its depth of solum by the calculated averaged rate of soil development in the reference chronosequence.

Evans and Rogerson (1986) employed this technique for dating glacial events in the Selamiut Range, Nachvak Fiord. Their 'depth of solum' data, observed in soil pits on moraine crests in the area, revealed three distinct soil depths (Fig. 6-5) which were transformed to soil ages using an averaged rate of soil development.

The calibration of scale on soil development processes is largely dependent on the nature of the soil forming factors through time which may be exclusive to individual study areas. Although the assumptions employed in the construction of the absolute framework are identical to those considered in the development of a relative soil-based chronology, the postulated variability in these factors will play a more critical role when calculations of soil development rates are attempted. Vreeken (1984) formulated these cautionary remarks as the *Principle of Fallibility of Pedogenic Indices as Estimators of Soil Age* which is expressed as follows:

> "Using a property of soil to construct a time measure that, in turn, is used to show that soils differ because of age, involves a circular argument, unless some independent criterion of environmental history relevant to soils is brought in."

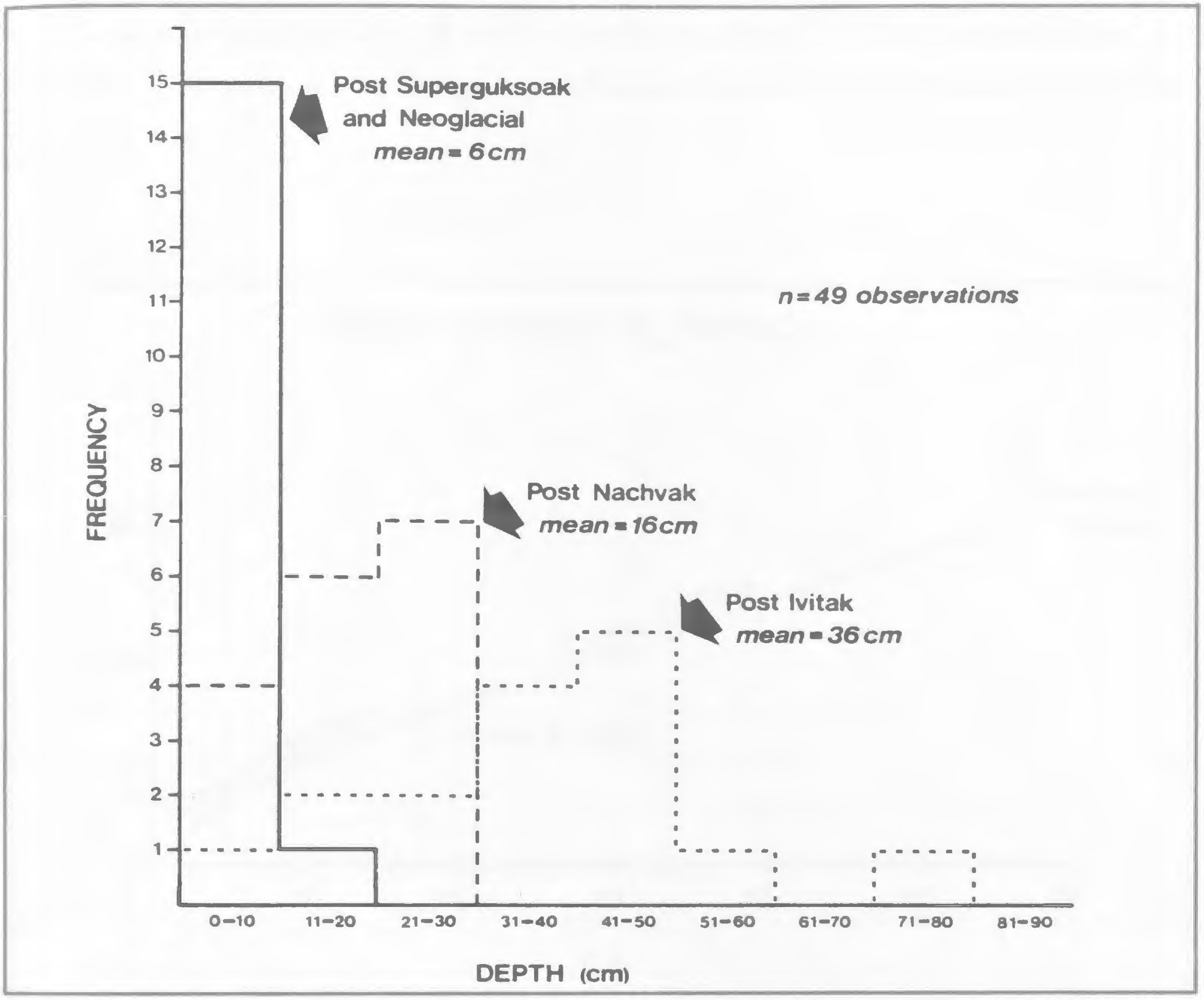

Figure 6-5: Soil depth - frequency histogram for soils measured to the base of the cambic B horizon, on moraine crests in the Selamiut Range, Nachvak Fiord (*from* Evans & Rogerson, 1987). The consistent trend of increasing solum depths with apparent ages of morphostratigraphic units suggests that this soil property may be a reliable index of soil age.

The second approach to age determination on the basis of soil development indices acknowledges that soil forming processes may vary in response to changing environmental conditions, and therefore a variable rate of soil development is suggested. Soils of different age before present (as determined by non-pedogenic criteria), with a partial overlap in soil history are interpreted as a time series. Each reference chronosequence typifies the soil development rate for the time period concerned, and the combined reference curve represents the variation in soil development over the complete time series.

Clark (1984) used this approach to date drift sheets in the Kangalaksiorvik Fiord/Iron Strand area of northern Labrador. He used two soil chronosequences, indirectly dated and with a large variance, to construct a soil development reference curve based on the depth of solum in soil profiles (Fig. 6-6).

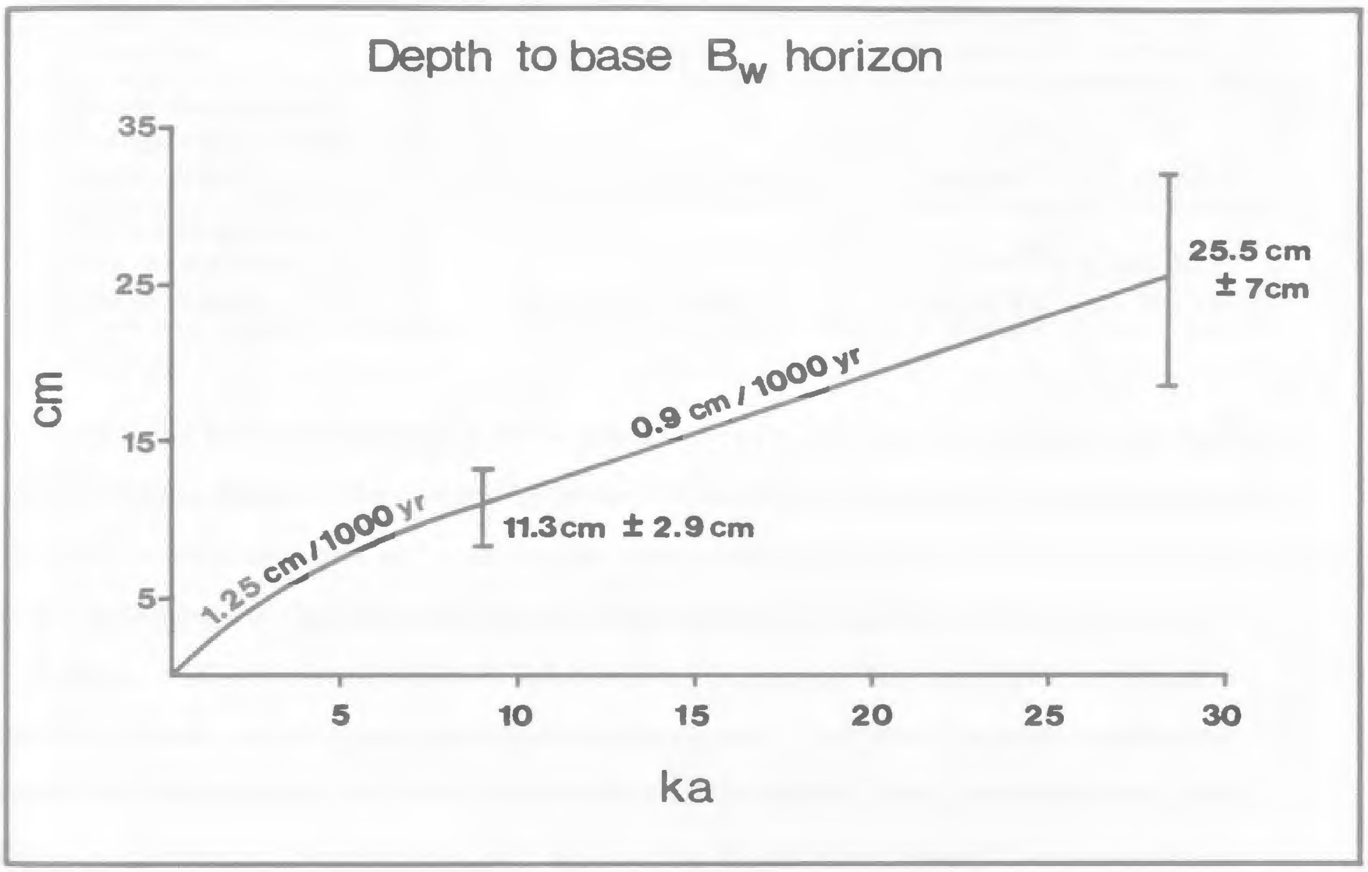

Figure 6-6: Plot of depth of solum to the base of the cambic B horizon against possible ages of drift sheets on which soils are developed (*from* Clark, 1984). The reference curve suggests a variable rate of soil development.

Vreeken (1975) questioned the normal practice of using these 'post-incisive' chronosequences as tools in a comparative study of soil development rates. Such interpretations assume that "macro-environmental conditions and local geomorphic regimen did not change significantly for the soil sites selected, during the time interval encompassed by the sequence" (Vreeken, 1975). Bockheim (1980) suggested that such assumptions must be validated through increased research on the changing effects of climate, vegetation and topography over time, on soil chronosequences.

Table 6-4: Soil development rates used in Quaternary chronologies in the eastern Canadian Arctic.

Soil development rates, eastern Canadian Arctic		
Location	Reference	Soil development rate
Nachvak Fiord, northern Labrador	Evans & Rogerson (1986)	1.0 cm ka^{-1} $\gg$40 ka
Kangalaksiorvik Fiord, northern Labrador	Clark (1984)	1.25 cm ka^{-1} $\leq$ 9 ka 0.9 cm ka^{-1} $>$ 9 ka
Broughton Island/ Kangetokluk Fiord, Baffin Island	Evans & Cameron (1979)	1 cm ka^{-1} $<$ 115 ka
Pangnirtung and Kingnait Fiords, Baffin Island	Birkeland (1978)	1 cm ka^{-1} $<$ 100 ka 0.25 cm ka^{-1} $\geq$ 100 ka

Table 6-4 lists the various soil development rates calculated for, or employed in absolute soil-based chronologies in the eastern Canadian Arctic. Evans and Rogerson (1986) suggested a linear soil development rate of 1.5 cm/ka for soils measured to the base of the Cox horizon, or 1.0 cm/ka to the base of the cambic B horizon in the Selamiut Range, Nachvak Fiord from comparisons with rates used on Baffin Island. They suggested that the effects of greater vegetation cover, precipitation, and annual temperatures in northern Labrador could result in a greater soil development rate than those used on Baffin Island. Evans and Cameron (1979) estimated a linear soil development rate of 1 cm/ka based on the solum[32] development of a chronosequence of four soils ranging in age from 8 ka to 115 ka. Birkeland (1978) proposed a linear rate of 1 cm/ka for the first 100 ka of soil development, decreasing to a rate of 0.25 cm/ka for soils older than 100 ka. Clark (1984) used rates of 1.25 cm/ka for the last 9 ka and a minimum rate of 0.9 cm/ka for earlier soil development. His results were calculated on the basis of two soils indirectly dated at 9 ka BP and >28.2 ka BP, respectively. Both Birkeland (1978) and Clark (1984) interpreted the solum as the soil profile to the base of the cambic B horizon.

[32] Presumably measured to the base of the cambic B horizon, but in one case to the base of the Cox horizon (Clark, 1987; Evans and Rogerson, 1987)

Calculation of soil development in outer Nachvak Fiord is not attempted due to restricted data. However it is suggested that an absolute soil-based chronology would require several different rates of soil development in order to compare pedological data from drift sheets of varied composition. On the basis of radiocarbon-dated shells from sediments associated with the local Adams Lake glacial phase (see Section 6.2), it is proposed that 36 cm thickness of soil (to the base of the cambic B horizon on Adams Lake moraine) developed in $\geq$ 30 ka (the predicted age of the glaciomarine diamicton in Adams Lake valley, which is regarded as a closer approximation of the date of deglaciation of Adams Lake moraine than the radiocarbon dates from the diamicton in Valley of the Flies). The subsequent advance of Nachvak phase regional ice to a point 4 km west of the Adams Lake moraine soil profile site would undoubtedly invalidate the assumption of monogenetic[33] soil development at this location, and hence disqualify the use of this site as a reference chronosequence for dating by average soil development.

The information available on soil development at the Adams Lake moraine site provides an opportunity to test the relevancy of the variable soil development rates employed by Clark (1984) in the Iron Strand/Kangalaksiorvik Fiord area, 50 km north of this study. For Clark's rates to be relevant in outer Nachvak Fiord, they must first predict the approximate age of Adams Lake moraine, and second they must distinguish the moraines of local origin from the younger moraines of regional origin. Table 6-5 displays the results of this test. The age of Adams Lake moraine was predicted relatively accurately, and Quartzite and Naksaluk moraines were isolated with a reasonable degree of confidence. In Table 6-5, the estimated ages of the moraines are given as maximum values because Clark's (1984) soil development rate of 0.9 cm/ka for soils older than 9 ka BP is considered to be a maximum value based on a minimum ^{14}C of 28.2 ka BP. The ages determined for regional till are probably overestimated relative to those for local till due to the contrast in till textures.

A seemingly more accurate estimate of the absolute age of Nachvak phase regional drift can be calculated by dating the soils on Quartzite and Naksaluk moraines by an average pedogenic index, based on the assumption that monogenetic soil development has occurred in the area since the last regional ice advance. The mean depth of solum to the base of the cambic B horizon divided by the linear rate of 1.25 cm/ka soil development, proposed by Clark (1984) for Late

[33]Soil monogenesis implies that the soil-forming factors remained constant during the course of soil formation.

Table 6-5: Estimated age of moraines in study area using depth of solum to the base of the cambic B horizon and Clark's soil development rates.

Estimated age of moraines using Clark's soil development rate				
Moraine	Till origin	Texture	Solum depth	Clark's rate
Adams Lake	Local	sandy	36 cm	$\leq$ 37 ka BP
Kammarsuit valley west	Local	sandy	42 cm	$\leq$ 43 ka BP
Kammarsuit valley east	Local	sandy	48 cm	$\leq$ 50 ka BP
Quartzite	Regional	clay/silt	30 cm	<30 ka BP
Naksaluk	Regional	clay/silt	25 cm	<24 ka BP

Wisconsinan soils, gives dates of 24 ka BP for Quartzite moraine and 20 ka BP for Naksaluk moraine. These estimated dates for Late Wisconsinan deglaciation correlate closely with the 17-23 ka BP date proposed by Evans and Rogerson (1986) for the Nachvak glacial event on the basis of morphostratigraphy and pedology in the adjacent Selamiut Range. These results also corroborate Clark's observations that a date of 9 ka BP is a minimum value for Late Wisconsinan deglaciation in northern Labrador.

The estimated ages of moraines derived using Clark's (1984) soil development rates appear consistent with morphostratigraphical evidence and radiocarbon dates from the study area. Although his rates resulted from research carried out in an area 50 km north of this study, there is no reason to suspect that the two areas have experienced contrasting palaeoenvironmental conditions. Similar guidelines were used for site selection, and both studies employed the field criteria and nomenclature of Birkeland (1974, 1978) for soil profile description. In addition, Clark visited the Nachvak Fiord area during the 1984 field season and verified that similar procedures were being used. Sample pits were dug on moraine crests in the Selamiut Range and there was general agreement on the depth of solum observed in the soil profiles. However, considering the wide range of solum depths which characterise the soil chronosequences used by Clark to develop his reference curve, the chronology of events suggested by the dates in Table 6-5 is regarded as tentative, and should be tested using independent sources of evidence.

6.5. Summary

Table 6-6 is a summary of the tentative chronologies of glacial events reconstructed from radiocarbon dates, ages predicted from regression analysis of amino acid ratios and ^{14}C dates (AA_{reg}, Table 6-6), ages predicted from Effective Diagenetic Temperatures for Nachvak Fiord (AA_{EDT}, Table 6-6), and soil development rates extrapolated from the Kangalaksiorvik Fiord area. These will be discussed in further detail in Chapter 7.

Table 6-6: Comparison of tentative absolute chronologies for glacial events in the study area, reconstructed from radiocarbon dates, amino acid ratios, and degree of soil development.

Comparison of absolute chronologies					
Glacial phase	Origin	^{14}C date	AA_{reg} date	AA_{EDT} date	Soils date
Adams Lake phase maximum	Local	33-41 ka BP	38 ka BP	38.7 ka BP	$\leq$ 50 ka BP
Adams Lake phase recessional	Local	—	29-30.3 ka BP	31.7 ka BP	$\leq$ 37 ka BP
Nachvak phase	Regional	>9 ka BP	—	—	20-24 ka BP

The radiocarbon age determinations for the Adams Lake glacial phase are possibly minimum estimates, but the soils-based chronology and amino acid ratios suggest that they may be close approximations of the true age of this event. The predicted amino acid dates are extremely tentative and are based on the assumption that the ^{14}C dated samples are valid control points. The calculated EDT of -1 °C used in the age-prediction equation is much higher than Clark's (1984) and Miller's (1985) values, and therefore implies a very steep temperature gradient north of Nachvak Fiord. This gradient may be realistic between Baffin Island and northern Labrador, but it cannot be explained for the 50 km distance between Kangalaksiorvik Fiord and Nachvak Fiord. There appear to be gross inconsistencies in total aIle/Ile ratios between these two areas which require further investigation beyond the scope of this study. Soil development rates used in the the soils-based chronology are extrapolated from Clark's (1984) study area. However, as soil development rates are closely associated with temperature regimes, a steep temperature gradient between the two study areas would invalidate any suggested correlations between soil data sets.

Chapter 7
Summary and discussion

7.1. Introduction

This chapter provides a summary of the major results from each type of evidence investigated in this study. The integration of these results forms a framework on which conclusions are presented on the glacial history and relative sea level change in the outer Nachvak Fiord area. These conclusions are critically examined and a summary of regional correlations are presented with respect to northern Labrador.

7.2. Glacial landforms, sediments and stratigraphy

Glacial landforms and sediments are described from outer Nachvak Fiord in Chapter 3. Observed and interpreted landform associations are the basis of a morphostratigraphy which is used to develop a tentative morphochronology. Interpretation of the depositional history of sediments which are genetically related to landforms lends support to the geomorphic evidence, and provides information on the style of glaciation. Analyses of the clastic and matrix components of sediments determine the origin and extent of glacial phases which further complement the morphostratigraphy, and aid in the construction of a lithostratigraphic framework.

The MI trimlines and moraines form the highest recorded glacial level in the outer Nachvak Fiord area. They slope eastward from a maximum elevation of 180 m aht in Adams Lake valley to 115 m aht in Valley of the Flies. Felsenmeer is widespread on the upland plateaus above this level and bedrock outcrops on the steeper slopes. Samples of MI drift are limited to high lateral moraine crests in Valley of the Flies due to the poorly preserved and soliflucted nature of MI moraines elsewhere in the study area. The clastic component of samples revealed mainly local lithologies with <5% glacial erratics from farther west in the fiord. Although values of <5% are insignificant with regard to the potential classification errors associated with the identification technique, it is proposed that the MI glacial level represents an extensive regional glaciation in the Nachvak Fiord area (Fig. 7-1). It is further suggested that during this glacial phase, ice advanced along the fiord and that Tinutyarvik and Adams Lake valleys acted as a distributary outlet for fiord ice. Correlative moraines may be those reported by Clark and Josenhans (1986) on the north side of Nachvak Fiord (Figs. 3-5 & 5-25).

MII moraine occurs below MI trimline on the southwestern flank of Naksaluk valley. It slopes northwards from 120 m aht to approximately 80 m aht in the Gurnot Lake area where it is truncated by the Naksaluk moraine. In Adams Lake valley it is associated with a moraine segment at approximately 116 m aht. The eastern limit of this glacial advance, termed the Adams Lake phase, is marked by the terminal moraine (21.5 m aht) in southern Valley of the Flies (Fig. 7-1). Adams Lake moraine, which slopes eastward from 86 m aht to 70 m aht in Adams Lake valley, is considered to represent a recessional stage of the Adams Lake phase (Phase 2, Fig. 7-1). Both MII drift and Adams Lake drift have predominantly local lithologies, although Adams Lake drift has a higher percentage of glacial erratics from up-fiord. These erratics are considered to be the reworked remnants of MI regional drift. The Adams Lake phase may represent a local valley glacier advance fed by ice from surrounding cirque glaciers. Stratigraphic sections exposed in southern Valley of the Flies suggest that ice associated with the Adams Lake phase advanced into a marine environment. Following retreat of ice from MII moraine to Adams Lake moraine in Naksaluk valley, an ice marginal lake formed, supported by ice in Adams Lake valley and possibly dammed by MII morainal deposits north of Gurnot Lake.

Naksaluk moraine, Quartzite moraine and Tinutyarvik moraine represent the limits of the Nachvak glacial phase in the study area (Fig. 7-1). The drift sheets bounded by these moraines are characterized by abundant glacial erratics from farther west in the fiord, by geochemical properties similar to those from a type area of Nachvak glacial deposits, and by a clay-rich texture which may have originated from the fiord. Naksaluk moraine truncates MII moraine of the Adams Lake glacial phase confirming a younger age for the Nachvak phase. While Nachvak ice receded from Naksaluk valley, an ice-supported lake developed, dammed in southern Naksaluk valley by Adams Lake moraine. A lake dammed by Quartzite moraine in Tinutyarvik valley also

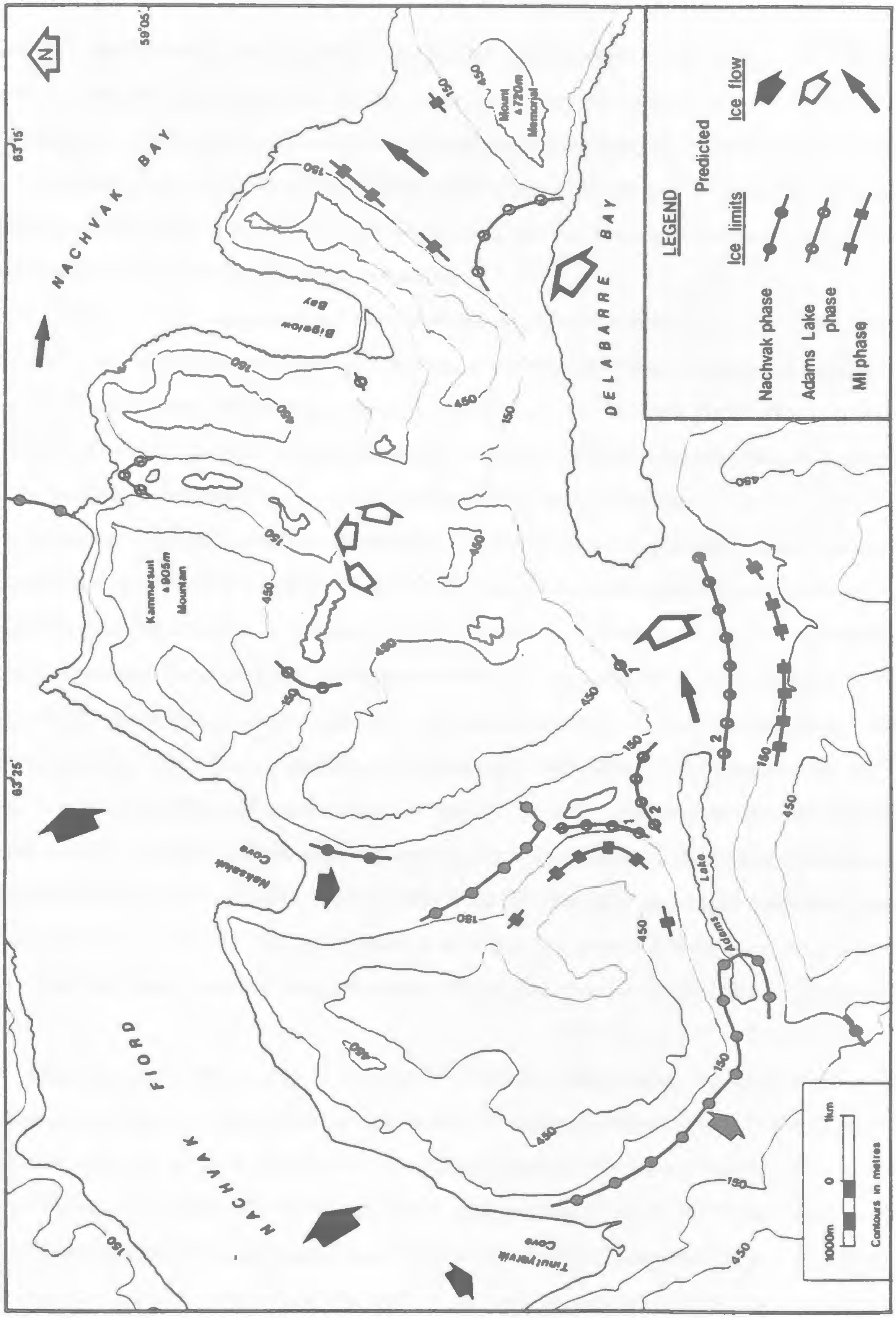

Figure 7-1: Summary of proposed ice flow patterns and ice limits in the outer Nachvak Fiord area.

formed during the Nachvak glacial phase, but subsequently emptied as the moraine was breached, causing an alluvial fan to be deposited in the southern part of the valley.

Lateral and end moraines occur in the Kammarsuit valley system. It is suggested that these moraines were deposited by local valley glaciers originating from cirques in the upland region. Till lithology and geochemistry support glaciation by local ice. Moraine-dammed lakes existed in these valleys as ice receded back into the mountains.

The similarity in the degree of soil development on moraines associated with local glaciation in Kammarsuit valley system and moraines deposited during the Adams Lake glacial phase suggests that these two events may have been coeval (Fig. 7-1). As both glaciations represent advances of local valley glaciers, it is considered probable that both areas responded to the same climatic factors which initiated cirque glacier development and expansion.

An extremely tentative absolute chronology of glacial events in the study area, based on rates of soil development on moraines, suggests that the Adams Lake glacial maximum, as represented by the end moraine in Kammarsuit valley east, occurred $\leq$ 50 ka BP. Similarly, a recessional stage from the maximum, as represented by Adams Lake moraine is tentatively dated at $\leq$ 37 ka BP. Radiocarbon dates obtained on shells from the glaciomarine diamicton which was deposited during the Adams Lake phase in southern Valley of the Flies suggest that the maximum occurred $\geq$ 38 ka BP (mean value of four ^{14}C dates). Absolute dates predicted from amino acid ratios on shells from the glaciomarine diamicton in Adams Lake valley tentatively imply that deglaciation of the valley occurred 29-32 ka BP. Employing similar methods, amino acid ratios were used to predict an age of 38-39 ka BP for the Adams Lake glacial maximum. In summary therefore, it is estimated that the Adams Lake phase occurred during the period 29-50 ka BP.

Although the above chronology is considered tentative, comparisons with glacial events from surrounding areas suggest some degree of consistency in regional chronologies. Clark (1984, p.184) described an ice advance in the Iron Strand/Kangalaksiorvik Fiord area which he tentatively dated at 33-50 ka BP and which had a proposed duration of $\leq$ 10 ka, based on the evidence of radiocarbon dates and amino acid ratios. Clark (1984) suggested that a continental ice sheet was responsible for depositing the drift sheet [Iron Strand drift] associated with this glacial phase. However, Iron Strand drift is distributed primarily along coastal regions in interfluves between the major valley systems, and drift lithologies and "...every other provenance indicator in all till samples examined from all drift sheets [Iron Strand drift] indicate extremely limited distances of transport" (1984, p.113). Therefore it is possible that this glacial phase may have consisted of local cirque and valley glacier expansion. The moraines mapped at 750 m asl and 940 m asl in Kangalaksiorvik Fiord and Ryans Bay, respectively, may relate to an MI glacial phase equivalent, since Clark presents no clear evidence that they are directly related to Iron Strand drift.

Evans and Rogerson (1986) tentatively dated an ice advance at approximately 40 ka BP on the basis of soil development on moraine crests associated with the late Ivitak glacial phase in the Selamiut Range. Although they suggested that this subphase was characterized by coalescence of local and regional ice, there is no definitive evidence of regional fiord ice associated with local cirque ice in their morphostratigraphic framework. The identification of large, lateral meltwater channels that have been cross-cut by the Ivitak moraine appears to be the only evidence of regional fiord ice, and it is possible that they were formed prior to the late Ivitak subphase. High-level glaciation equivalent to the regional ice limits of the Ivitak phase, as proposed by Evans and Rogerson (1986, Fig.10), most certainly occurred at some time in the past, but it is probable that the local Ivitak valley ice advance postdated this occurrence. As Evans (1984, p.39) remarks, "...on Ivitak Rigg and in Ivitak valley...morphological evidence appears to cover a broad span of chronological events". On the basis of the above presentations, it is tentatively suggested that significant local cirque and valley glacier advances occurred in northern Labrador between 29 ka BP and 50 ka BP.

The more extensive MI glacial phase must predate the Adams Lake glacial phase. Although this regional glaciation may be pre-Wisconsin in age, it is more likely an Early Wisconsinan advance of Laurentide ice which extended onto the Labrador Shelf. Clark and Josenhans (1983) originally suggested a correlation between an early Wisconsinan glaciation (>70 ka BP) and the upper till unit recognized on the Labrador Shelf. Evans and Rogerson (1986) proposed an early Wisconsinan age for the early Ivitak glacial phase in the Selamiut Range. This phase is associated with the deposition of till sheets up to 650 m asl only 20 km upfiord from Nachvak Bay and therefore may indicate an extensive regional glaciation. It is this subphase which may correlate with the MI glacial phase in the outer Nachvak Fiord area.

The lower degree of soil development on moraines associated with the Nachvak glacial phase, in comparison to Adams Lake phase moraines, suggests that they are younger than 32 ka BP (minimum date for local ice recession from Adams Lake valley). By using a linear soil development rate of 1.25 cm/ka, a tentative date of 22 ±2 ka BP was calculated for the Nachvak glacial phase maximum. Minimum dates for this event are provided by radiocarbon determinations on shells associated with a shoreline which formed subsequent to deglaciation of the fiord. These shells provided a date of 9170 ±100 years BP. Comparison of these tentative dates with estimated dates for the Late Wisconsinan maximum [18 ka BP (Short, 1981); 17-23 ka BP (Evans and Rogerson, 1986); 20 ka BP (Josenhans *et al.*, 1986)] suggests that the Nachvak glacial phase represents the Late Wisconsinan ice advance in the outer Nachvak Fiord area.

7.3. Fiord and lacustrine evidence

Chapter 4 examined the acoustic record of subbottom sediments from Nachvak Fiord and Adams Lake. The nature of the geophysical system on which this type of information is based, and the limitations of the acoustic profiling technique must be evaluated during the interpretation of such data. These considerations are stressed in view of the limited ground-truthing available for the stratigraphic interpretations.

The glacial - deglacial model constructed for Nachvak Fiord sediments is based on the acoustic properties of the units identified and the acoustic stratigraphical relationships between these units (Table 7-1). The lowermost unit, A, forms the impenetrable base of the fiord and defines the individual fiord basins. These basins are named Townley, Koktortoaluk, Ivitin and Outer Basins from west to east along the fiord. The bathymetrically-high barriers, composed of Unit A and separating the basins are named Kogarsok, Ivitak and Shoal Water Cove sills.

The lowermost basin-fill unit, B, is the thickest unit in the sequence, between 80 m and 130 m thick in the eastern basins. Unit B is interpreted to represent prograded deposition from sources located at Kogarsok and Ivitak sills. This sediment may have originated either from an ablating ice margin standing at the sills as ice receded west along the fiord, or from rivers draining the fiord hinterland and discharging into the basins near to the sills. Neither hypothesis can be supported with any definitive evidence.

Table 7-1: Summary description of Nachvak Fiord subbottom acoustic units. Thickness is expressed in metres and is provided for basins from west to east in the fiord.

Summary description of Nachvak Fiord acoustic units				
Unit	Thickness	Acoustic properties	Depositional style	Interpretation
F	3/6/8/10	Parallel reflectors in basin centres, upward transparency, point source reflectors	Uniform rain out of fine grained sediments, debris release from icebergs	Ice distal and modern sediments, typically fine grained with dropstones
E	<5/10/7/10	Parallel reflectors of variable strength, conformable with basin margins, slumping	Stream underflow/ sediment gravity flow	Ice proximal to ice distal sediments, variable sediment type
D_2	<5/15/15/20	Parallel reflectors, finely stratified, conformable, slumps	Cyclic sedimentation from blanket source	Deposition of silt and clay beneath an ice shelf
D_1	Variable	Acoustically stratified basin-fill unit, onlap with hummocks	Sedimentation from blanket source	Ponded sediments between hummocks
C	>20/30/50/35	Hummocky, lenticular reflectors in wedge or lens shaped unit	Subglacial deposition beneath grounded ice shelf	Glacial till and moraine complexes
B	?/80/130/80	Parallel to sub-parallel reflectors in basin-fill unit, acoustically variable	Stream underflow from ablating ice margin, or prograding deposition from sediment source	Ice proximal glaciomarine sediments or non-glacial deltaic sediments
A		Acoustically massive/ impenetrable		Till/bedrock

Unit C is considered to be glacial till, an interpretation supported by observation of this unit where it is exposed at the fiord bottom (Rogerson *et al.*, 1986*a*). This till unit occurs in all basins but is thickest in Ivitin basin (Table 7-1). Accumulations of till on Kogarsok and Shoal Water Cove sills have distinct moraine morphologies. It is tentatively suggested that Unit C was deposited beneath a partially grounded ice shelf which extended over the fiord basins. This conclusion is reached on the basis of the following observations: (1) The lower boundary of Unit C is generally conformable with the trend of acoustic stratification in the underlying deposit.

Although truncated Unit B reflectors are evident at the boundary, there is no indication that an outlet glacier advanced through the fiord eroding large amounts of material in its path. (2) Basin margins display evidence of marked deformation and erosion of strata which may be interpreted as grounding points, or areas which experienced excessive loading pressures. These points constitute an easterly dipping plane which intersects the surfaces of Kogarsok and Shoal Water Cove sills, and which may have been horizontal in an isostatically depressed subglacial environment. (3) Shoal Water Cove moraine has the characteristics of a 'lift-off' moraine and occurs above a probable grounding point for an ice shelf. Smaller 'lift-off' moraine ridges were observed along the surface of the till unit, west of Shoal Water Cove moraine, in Ivitin and Koktortoaluk Basins. These features may have formed in response to buoyancy changes or shifts in the grounding line location of an ice shelf.

Deposition of Unit C is considered to be time-transgressive between the fiord basins as Kogarsok moraine, with its distinct, asymmetrical end moraine shape, may reflect deposition against a grounded ice margin following disintegration of the suggested ice shelf in the eastern basins.

Ice shelf lift-off and the development of open water conditions beneath the ice are suggested by Unit D characteristics. This unit appears to be finely stratified as evidenced by the closely spaced, parallel acoustic reflectors and is subdivided into subunits based on the stratigraphical relationships of these reflectors with the underlying till unit. Subunit D_1 occurs in association with the hummocky surface of the till and infills the intervening depressions. Subunit D_2 mimics the upper surface of the till unit and may have been deposited from a blanket source. This stratified unit occurs in all basins but is thickest in the deeper Outer Basin, where open water conditions may have developed earlier. Alternatively, the repeated grounding of the ice shelf in Ivitin and Koktortoaluk Basins, as evidenced by the hummocky nature of the till unit in these basins, may have prevented thicker accumulation of this subunit. Further evidence of ice shelf buoyancy and thinning is reflected in the major slumps that occur at the basin margins within, and at the surface of this subunit.

Unit E is an acoustically varied and stratified unit which was deposited conformably on the surface of Unit D. It is postulated that ice shelf disintegration had occurred prior to deposition of Unit E, and that this sediment originated from stream underflows and/or sediment gravity flows generated at an ice margin in a proximal to distal location. The marked difference in thickness of this unit either side of Kogarsok moraine may suggest that Townley Head Basin was ice covered during this period, and further demonstrates the probable diachronous deposition within this and underlying units.

The uppermost unit in the fiord basins (unit F) records the transition from ice-distal and ice-rafted sediments to modern marine sediments. The upward decline in acoustic stratification and point-source reflectors within this unit suggests the waning influence of an active calving ice front and a decrease in ice-rafted material.

A wedge-shaped subunit of acoustically stratified sediment within Unit F in Koktortoaluk Basin may be related to the sudden emptying of an ice-dammed lake in Tallek Arm, or to the discharge of lake water down the Palmer River from ice-dammed lakes west of the drainage divide.

No groundtruthing of Nachvak Fiord acoustic and seismic records is available and therefore the interpretation of these records remains speculative. However, integration of the proposed depositional model with terrestrial glacial geology and geomorphology can provide a tentative framework for future discussion. The eastern limit of the last regional glaciation in the fiord, as suggested by landform and lithostratigraphic evidence from the study area, is interpreted to be at some point between Naksaluk Cove and the mouth of Kammarsuit valley east, the most likely location being on the fiord threshold. The till unit (unit C) recognized in the fiord subbottom sediments may be correlated with the last regional advance of ice through the fiord and therefore may be Late Wisconsin in age. The fiord threshold is the eastern limit of unit C in the fiord and although it has been proposed that this unit extends onto the Labrador Shelf (Clark and Josenhans, 1986; Josenhans *et al.*, 1986), there is no definitive evidence to support such a proposal in the Nachvak Fiord area.

Subglacial deposition beneath a grounded to partially-grounded ice shelf is proposed as a likely origin for unit C in the fiord. Naksaluk valley is interpreted to have been backfilled by ice from the fiord during the Late Wisconsin. Naksaluk moraine which was deposited during this advance has a characteristically low gradient with a steep proximal side. D.J.A. Evans (personal communication, 1986) compared the morphology of this moraine (from photographs) to those which he has observed along the fiords in northern Ellesmere Island, and which he has interpreted to be ice shelf moraines. Similarities with a suggested ice shelf moraine, 8 km north of Kangalaksiorvik Fiord (Clark, 1984, pp. 52-53), support this observation. Tinutyarvik and Quartzite moraines are also associated with the advance of regional fiord ice, but display typical valley glacier lateral and terminal moraine morphologies. Relative dating of moraines in Naksaluk and Tinutyarvik valleys, on the basis of the degree of soil development, tentatively suggests that Tinutyarvik valley, acted as a distributary outlet for fiord ice, and was deglaciated earlier than the fiord basin. Therefore, the point at which ice receded from Tinutyarvik valley may indicate the transition from a grounded to partially-grounded ice shelf in the fiord, and the deposition of an ice shelf moraine by ice backfilling Naksaluk valley. The clay-rich tills which comprise the regional drift sheets in the study area are likely correlatives to unit C in the fiord basins.

A rough estimate of ice thickness off Naksaluk Cove was calculated to be 350 m $\pm$25 m based on the depth of unit C below present sea level and the elevation of Naksaluk moraine above present sea level. Clark and Josenhans (1986) in their Late Wisconsinan ice model for Nachvak Fiord derived an ice thickness of 500-550 m in this area from the elevation of possible MI or MII moraine equivalents on the northern side of the fiord (Figs. 3-5 & 5-25). Also included in their model was Quartzite moraine which they calculated to be at *ca.* 280 m asl, but which was measured on numerous occasions during field work at *ca.* 130 m aht. They used the gradients on these moraines to reconstruct a Late Wisconsinan ice surface which included an ice shelf that extended 30-50 km onto the Labrador Shelf.

Evidence for the regionally extensive MI glacial phase in the fiord subbottom sediments may be related to the thick (80-130 m) accumulations of acoustically stratified sediment that underly unit C. The acoustically massive base of the fiord may be partially composed of till which cannot be differentiated from the bedrock basement on the air gun record. This tentative glacial - deglacial sequence could therefore be Early Wisconsin in age based on the estimated age of the MI glacial phase. Further speculation may suggest that the upper till on the the Labrador Shelf is the eastward extension of this lower glacial sequence.

Analysis of the 3.5 kHz acoustic record of Adams Lake sediments reveals three acoustic units which are tentatively interpreted as representing sedimentary facies related to a glacial - deglacial cycle (Table 7-2). Adams Lake is characterized by three basins, *X*, *Y* and *Z* (from west

Table 7-2: Summary description of Adams Lake subbottom acoustic units. Thickness is expressed in metres and is provided for basins from west to east in the lake.

Summary description of Adams Lake acoustic units				
Unit	Thickness	Acoustic properties	Depositional style	Interpretation
C	1.5/1.5/1.5	Acoustically transparent to semi-transparent	Sedimentation from suspension, modern lacustrine processes	Homogeneous fine grained sediment
B3	1.5/3/-	Acoustically stratified, conformable, upward transparency	Density currents/ fall out from suspension	Variable fine grained sediment, possibly fining upwards
B2	1/1/-	Strong, well-defined acoustic reflector, conformable	Density currents	Coarse clastic deposit
B1	>3/10/2	Acoustically stratified, basin-fill deposit	Density currents and underflows	Ice proximal to ice distal sediments
A		Acoustically massive/ impenetrable		Till/bedrock

to east), ranging from 12 to 21 m in depth, separated by distinct barriers between 3 m and 6 m below lake level. Unit A, the lowermost unit, is characterized by a massive structure and reflector-free image and may comprise drift or bedrock. The surface morphology is variable and irregular and defines the basins in which the overlying units are recorded. The most westerly barrier is likely the subaqueous extension of the Tinutyarvik moraine, and therefore it is suggested that Unit A comprises morainic drift/till in this locality.

Unit B is acoustically stratified and is divided into subunits on acoustic stratigraphical relationships within the lake basin. Subunit B1 shows an onlap relationship with Unit A and may be classified as a basin-fill deposit. Subunit B2 pinches out along the basin margins and is characterized by a strong, well-defined acoustic reflection. It generally appears conformable with the stratification of the underlying subunit and occurs only in basins X and Y. Subunit B3 is a basin fill deposit pinching out along the lower slopes of the basins. Unit B is interpreted as a proglacial deposit ranging in depositional environment from ice-proximal to ice-distal, towards the east: it also occurs only in basins X and Y. Subunit B2, interpreted as a coarse clastic deposit, is tentatively related to the draining of a moraine-dammed lake in southern Tinutyarvik valley.

Unit C, the uppermost unit in the acoustic record, appears draped over the sediment package beneath. It is predominantly an acoustically transparent unit with a uniform thickness of approximately 1.5 m. Unit C is interpreted as a postglacial lacustrine sediment.

The identification of the subaqueous extension of Tinutyarvik moraine in the acoustic stratigraphy of Adams Lake sediments provides a tentative framework for the depositional history of the three acoustic units, and their interpretation as discrete sedimentary facies related to a glacial - deglacial cycle. Tinutyarvik moraine is considered to be the limit of the Late Wisconsinan ice advance into Adams Lake valley and therefore unit A which composes Barrier *U*/Tinutyarvik moraine should be Late Wisconsin in age within Basin *X*. Barriers *V* and *W* and unit A in Basins *Y* and *Z* may be either bedrock or moraines/till which were deposited during the Adams Lake glacial phase.

The acoustically stratified subunit B1 is interpreted to be a proglacial deposit and may comprise sediments resulting from deglaciation during both the Adams Lake and Nachvak glacial phases. This subunit is greater than 10 m thick in Basin *Y*. Subunit B2 may record the significant input of coarse clastic debris dammed by Quartzite moraine. A trench cuts through the moraine and feeds a major multichannelled alluvial fan spreading towards Adams Lake. Since the alluvial fan appears to have been formed predominantly under subaerial conditions, its deposition necessarily occurred after Tinutyarvik valley was deglaciated. Therefore, subunit B2 may act as a stratigraphical marker indicating the minimum depth of postglacial sediments within the acoustic record.

In western Adams Lake valley, a marine terrace is cut into the alluvial fan deposits at approximately 8 m above lake level. The next lowest marine shoreline is at an elevation of 34.25 m aht, or 1.75 m below lake level in central Adams Lake valley. Therefore the deposition of the alluvial fan may have been close to the time when Adams lake changed from an inlet of the sea to a freshwater lake. As unit C is interpreted to be a relatively homogeneous, fine grained unit deposited under low sediment influx rates in a lacustrine environment, the marine/freshwater transition is probably recorded in the upper part of subunit B3, between 1.5 m and 3 m below the lake bottom.

Shortly before this thesis was submitted, a preliminary radiocarbon date was obtained from Isotrace Ltd. on marine organic material found near the base of a 2 m core (C2, Fig. 4-12) taken from Basin *Z* in 1985. The date is 22,090 ±170 years BP, suggesting that Late Wisconsinan ice did not cover the eastern end of the lake.

7.4. Relative sea level change and raised marine evidence

Chapter 5 was concerned with the late-glacial and postglacial emergence history of the outer coast of Nachvak Fiord. The extrapolation of this information up-fiord provides the basis for the presentation of a model which describes the pattern and extent of glacio-isostatic recovery in the region. The identification of moraine complexes in association with the stepwise decrease in marine limit elevations upfiord permits an appraisal of the emergence history in relation to the final retreat of Laurentide ice from the fiord region. A late-glacial chronology is presented which, although mainly a relative one, provides an important framework for comparison and correlation of major glacio-isostatic events in northern Labrador.

The data from which the above model is derived were recorded from the elevations of 168 shoreline segments in the study area. A consistent and systematic approach to data acquisition has allowed a quantitative assessment of the data. Four types of error are recognized and their magnitudes are estimated for this study. The combination of *datum error* and *instrument error* provides an error margin of ±0.684 m for surface elevations quoted. The addition of this figure to the *observation error* associated with the point of measurement on the landform, plus the error assigned to the correction factor which relates marine features to tidal regime gives a total error range of ±1.684 - 2.184 m for sea level altitudes computed from raised marine evidence in the study area. These results compare favourably with similar research carried out in Scotland, where superior data sets from well-preserved marine features and 1st order bench marks provide more ideal conditions than in northern Labrador.

Correlation of raised shoreline segments was achieved using a combination of trend surface analysis and the isobase trend of a well-developed shoreline at approximately 25 m aht in Adams Lake valley. Using this method, a sequence of eleven shorelines, ranging in elevation from 9 m to 73 m, was resolved from the data. Results from statistical analyses of these best-fit planes suggest that they are accurate reconstructions of former sea levels. Shoreline segment elevation points were plotted in a vertical plane (236° true azimuth) perpendicular to the mean shoreline isobase trend. Slopes on the projected shorelines range from *1.722* up-fiord to horizontal for the two lowest levels. Two inversions in the normal pattern of decreasing gradient with decreasing elevation for plotted shorelines may represent reversals in the uplift rates associated with isostatic recovery. These episodes of renewed downwarping of the earth's crust are tentatively related to glacial readvance phases west of the outer coast.

The marine limit in the study area and along the fiord coastline in general displays a stepwise westerly decrease in elevation, although lower raised shorelines are tilted up to the west. This suggests that fiord ice retreated in contact with the sea so that only those areas beyond an ice margin recorded evidence of sea level stands during deglaciation.

Shoreline K (67-73 m aht) comprises marine limits observed in central and eastern Adams Lake valley, Delabarre Bay and Valley of the Flies. The validity of this shoreline is in doubt for two reasons: **(1)** It is a composite of marine limit elevations and therefore it may not have been formed contemporaneously throughout the area; and **(2)** the low gradient of the shoreline (*1:3000*) is difficult to explain in relation to the gradients of lower late-glacial shorelines. The western limit of Shoreline K is Tinutyarvik moraine in Adams Lake valley and the fiord threshold in Nachvak Bay.

Shoreline J (47-68 m aht) occurs as the marine limit in western Adams Lake valley and Bigelow valley, but further east it exists below the Sl-K level. There is no distinct western limit to this shoreline in Adams Lake valley nor Tinutyarvik valley; however the fiord threshold marks this limit in Nachvak Bay.

Shoreline I (40-67 m aht) occurs throughout the study area and may have extended as far west as Ivitak Cove where a lake shoreline (67 m aht) recorded by Evans and Rogerson (1986) is reinterpreted to be of marine origin and to be a correlative of this shoreline. During this period of shoreline formation, the central part of the study area stood as an island off the mainland coast. The western limit of this shoreline is unknown as a subsequent readvance of Laurentide ice eroded the evidence of this former shoreline, except for the segment beyond the glacial maximum moraine in Ivitak Cove, and the segments east of Tinutyarvik Cove.

Shoreline H (37-53 m aht) records the transition of the study area from an island to a peninsula which was joined to the mainland by a narrow isthmus located between Tinutyarvik and Adams Lake valleys. The western limit of observations on this shoreline is a terrace (53 m aht), south of Ivitak Cove, initially recorded as lacustrine by Evans and Rogerson (1986), but tentatively proposed to be of marine origin in this study. As in the case of the previous shoreline, the western limit of Sl-H is unknown due to a subsequent ice readvance which removed the evidence of its former presence as far east as Tinutyarvik Cove, with the exception of southern Ivitak Cove.

Shoreline G (28-48 m aht) was formed at some point when ice had readvanced through the fiord. The maximum possible extent of this readvance [Shoal Water Cove readvance] lies between Ivitin Cove and Tinutyarvik Cove. Downwarping of the earth's crust associated with this readvance resulted in Shoreline H being tilted down towards the ice centre and subsequent submergence by the Sl-G sea level west of, and including Ivitin Cove. Therefore even though ice may not have readvanced further east than Ivitin Cove, evidence of the Sl-H sea level would not have been preserved in this cove.

Prior to the formation of Shoreline G, a lake dammed by Quartzite moraine emptied and discharged north through Tinutyarvik valley and east through Adams Lake valley. Shoreline G, west of Adams lake, is recorded as a terrace eroded into an alluvial fan which was deposited during this event.

Shoreline F (26-55 m aht) was formed during the Kogarsok still-stand, the western limit of the shoreline occurring on the distal side of Kogarsok moraine. Prior to the formation of this shoreline, Adams Lake changed from a sea inlet to a freshwater lake. The recently acquired preliminary radiocarbon date of 22,090 $\pm$170 years BP (Isotrace Ltd.) from marine organics in Core 2, Adams Lake suggests a maximum age for the Kogarsok still-stand.

Shoreline E (23-48 m aht) extends from the outer coast west as far as Townley Head which is the location of a moraine that reaches more than halfway across the fiord at this point. It is suggested that this moraine, the Townley Head moraine, was deposited contemporaneously with the formation of Shoreline E, during the Townley Head still-stand.

Shoreline D (19-41 m aht) was recorded as far west as Tasiuyak Arm, the most westerly site selected for shoreline measurements. It is probable that this shoreline extends to the head of the fiord, 2 km to the west. Shoreline D was formed during the Tessersoak readvance and may be associated with a moraine system situated south and west of Nachvak Lake. These moraines were tentatively described as 'Sheppard moraine' equivalents by Rogerson (personal communication, 1986), a term first used by Löken (1962*b*) to describe moraines associated with the Kangalaksiorvik readvance in northernmost Labrador.

Shoreline C (17-24 m aht) is the lowest of the tilted shorelines in the region and may relate to a glacial still-stand as ice receded west of the Labrador-Ungava drainage divide. Shorelines B (15.5 m aht) and A (9 m aht) are recognised as horizontal shorelines in the study area.

Tinutyarvik moraine and the Fiord threshold are proposed as the limits of the Nachvak glacial phase in the study area. The marine limits identified east of, and distal to these limits compose the Sl-K level which appears anomalous to the general pattern of glacio-isostatic recovery in the region. It is suggested that each marine limit, or combination of limits represents an individual shoreline, and consequently Sl-K does not demonstrate the true extent and geometry of a single synchronous sea level. The preservation of these shoreline segments distal to the Late Wisconsinan maximum ice extent provides for the possibility that some of the shorelines which they represent are pre-Late Wisconsinan in age. Maximum dates for these high sea level stands can be estimated from the proposed ages of the drift in which they are eroded. Segments recorded in Valley of the Flies are situated in a distal position relative to the Adams Lake glacial phase maximum, and can therefore be attributed a maximum age of Early Wisconsinan (MI drift). Remaining segments are eroded in Adams Lake drift and may therefore be Middle Wisconsinan in age or younger.

Higher sea levels than present during the Middle Wisconsinan are evident from the marine sand unit and glaciomarine diamicton recorded in southern Valley of the Flies. It is suggested that the diamicton was deposited during the maximum of the Adams Lake glacial phase, tentatively dated at 34-50 ka BP. The fossiliferous marine sand overlying the diamicton was tentatively dated at 34-37.4 ka BP, based on predicted numeric age estimates derived from amino acid ratios on shells. Dreimanis and Raukas (1975) suggested that the Middle Wisconsinan (23-65 ka BP) was characterized by eustatic sea levels 10-20 m below present sea level. This, however, must be reassessed in view of recent work which questions the notion of global eustatic changes in sea level. If northern Labrador was characterized by local cirque glaciation during the Middle Wisconsinan, then the suggested sea level stands of 60-70 m above present sea level may be explained by Clark's (1976, 1980) proposal of gravitationally elevated sea levels around glacier margins and glacio-isostatic loading of the crust. Observations of isolated raised marine sediments dating from about 40 ka BP in northern Labrador and Baffin Island (Fillon, 1985, p. 232) may be interpreted as supporting a regional pattern of higher sea levels than present during the Middle Wisconsinan.

Integration of raised marine evidence and acoustic stratigraphy of fiord subbottom sediments provides the basis for a tentative relative chronology of late-glacial events in the Nachvak Fiord region. The distribution of Sl-J segments lends support to the hypothesis that Nachvak phase ice receded from the distributary outlet in Tinutyarvik and Adams lake valleys while it remained grounded or partially grounded at the fiord threshold. Shorelines I and H may extend at least as far west as Ivitak Cove, which suggests that ice had evacuated the central and eastern parts of the fiord (that is, Ivitin and Outer Basins) by this time. A tentative model which explains the style of deposition of fiord subbottom sediments suggests that a floating ice shelf developed in the fiord which may have subsequently receded to a location west of Ivitak Cove. It has been stated previously that interpreted grounding points of this ice shelf in Outer Basin may comprise an easterly dipping plane consistent with the calculated gradients on late-glacial raised marine shorelines. Potential grounding points for the ice shelf, west of Ivitak Cove, are Ivitak sill and Kogarsok sill.

The pattern of differential relative sea level change suggests that an ice readvance occurred following the formation of Sl-H [Shoal Water Cove readvance]. The distribution of Shoreline G segments in the fiord indicates that ice may have readvanced as far as Ivitin Cove, but not farther east than Tinutyarvik Cove. In Section 5.4.2 it was suggested that a likely location for this readvance may have been Shoal Water Cove moraine, based on a previous interpretation of this moraine as representing a significant still-stand during ice recession from the Late Wisconsinan maximum (Rogerson *et al.*, 1986*a*). However, reinterpretation of the fiord acoustic stratigraphy in this study tentatively suggests a lift-off origin for the moraine during initial ice shelf buoyancy. The occurrence of minor lift-off ridges on the till surface at locations in Ivitin Basin which correspond to an area between Ivitin Cove and Tinutyarvik Cove in the fiord, suggests that these features may represent the grounding points of a fluctuating ice margin. The thicker accumulations of till in this basin relative to Outer Basin may indicate longer periods of subglacial deposition, although this may not necessarily be due to a readvance of ice into the area. The greater thickness of subunit D_2 in Outer Basin may suggest that approximately 5 m of sediment was eroded from Koktortoaluk and Ivitin Basins during a readvance phase, or alternatively, open water conditions may have developed earlier in the deeper Outer Basin. Although evidence relating to the maximum position of Shoal Water Cove readvance is uncertain, there is good evidence that such a readvance probably took place, as indicated by differential relative sea level change.

The Shoal Water Cove readvance is tentatively correlated with the Noodleook readvance phase in northernmost Labrador (Löken, 1962*b*). Although Löken (1962*b*) also suggested a Two Loon readvance phase during recession from the Noodleook phase, it appears more likely that the latter is a correlative to the Shoal Water Cove readvance, based on the following considerations. First, Löken described the Two Loon readvance as minor compared to the Noodleook and Kangalaksiorvik readvance phases. The magnitude of shoreline deformation resulting from the Shoal Water Cove readvance suggests that it represented a major readvance of ice. Second, Löken expressed some doubt concerning the validity of the shoreline (Sl-2) related to the Two Loon readvance moraines [based on only four observations]. Shoreline G which is related to the Shoal Water Cove readvance is a distinct level with well-defined beach ridges and marine terraces. Third, there is a possibility that either the Kogarsok or Townley Head still-stands represent minor readvances, and therefore either could be a correlative to the Two Loon readvance. Last, the Two Loon readvance may only be a small local event restricted to northernmost Labrador.

Shorelines F and E were formed while ice stood still at Kogarsok and Townley Head moraines, respectively, during recession from the Shoal Water Cove readvance maximum.

The Tessersoak readvance is recognized in the record of relative sea level change as a period of marked downwarping of the earth's crust or renewed ice-water gravitational attraction. Shoreline D which records this renewed sea level deformation was documented along the fiord coastline, almost to the fiord head, and suggests that deglaciation of the fiord was complete by this time. Shells recovered from a river-eroded section in Adams Lake valley, approximately 4 m below Shoreline D, were radiocarbon dated at 9170 $\pm$100 years BP (GSC-4161), and therefore this is a possible minimum date for an ice-free fiord environment. Löken (1962*b*) dated his 'Sl-3' shoreline (25-37m aht) which was formed during the Kangalaksiorvik readvance at 8700 $\pm$470 years BP (L-642). There is a strong possibility that these two readvance phases are related and that a readvance of Laurentide ice characterized the early Holocene period in northern Labrador.

The low gradient on Shoreline C suggests that glacio-isostatic recovery was almost complete during its formation. Horizontal Shoreline B at 15.5 m aht is probably a correlative to the 'Sl-4' shoreline of Löken (1962*b*) which he recorded at 15.5 m aht in northernmost Labrador, and which he related to the maximum level of the Tapes transgression in Scandinavia, dated at 4700 years BP. Shoreline A may be a minor transgression during regression from the Sl-B sea level. Other raised shoreline segments below 15.5 m aht were recorded in the study area and may relate to similar transgressive events.

The directions of dip on the nine tilted shorelines in the study area are used to tentatively locate ice centres which affected the isostatic recovery of the Nachvak Fiord region during specific periods of shoreline formation. Tentative correlations are made between Shorelines G and D which are associated with the Shoal Water Cove and Tessersoak readvances and the 'Sl-1' and 'Sl-3' shorelines which Löken (1962*b*) related to the Noodleook and Kangalaksiorvik readvances in northernmost Labrador, respectively. Implicit in this tentative correlation is the assumption of synchroneity of shoreline development, and therefore it is suggested that the intersections of the orthogonals to the isobases of Shorelines Sl-G and Sl-1 and Shorelines Sl-D and Sl-3 mark the approximate locations of ice loading centres which existed during the formation of these shorelines. With respect to the former, the position of the intersection point is located approximately 60 km east of George River and 120 km west of the head of Saglek Fiord. Support for such a loading centre may be drawn from the observation of large ice-dammed lakes west of the Labrador-Ungava drainage divide during the late-glacial period (Ives, 1960). The orthogonals to Shoreline D isobases and Sl-3 isobases intersect an area previously proposed as a late-glacial ice centre, southwest of Schefferville (Barnett and Peterson, 1964).

An emergence rate of 3.5 m/ka is calculated for the last 9170 years based on a radiocarbon date obtained from shells stratigraphically related to Shoreline D.

7.5. Regional interpretation

In order to regionally synthesize the results of this study, it is important to initially evaluate them in the context of northern Labrador weathering zones which although controversial in concept, represent regional stratigraphic units which could form the framework for a regional Quaternary history.

The Torngat weathering zone which is regarded as a non-glaciated zone is unlikely to exist in the outer Nachvak Fiord area. Although observations on the high plateaus were reconnaissance in nature, several erratics from farther west in the fiord were positively identified in the felsenmeer spreads at sites above 750 m asl. These observations dispute the findings of Coleman (1921) who suggested that no erratics existed in the area above 650 m asl. It is likely that the area above the highest moraine system (MI level, 180 m aht) represents an equivalent of the Komaktorvik weathering zone (Ives, 1974, 1975, 1978). This zone may be extrapolated upfiord to the Selamiut Range where Evans and Rogerson (1986) recorded the elevation of the highest moraines at approximately 650 m asl (Early Ivitak subphase), and which are proposed to be correlatives of the highest moraines in outer Nachvak Fiord.

Most of the major valleys in the study area lie within the Koroksoak weathering zone, the exceptions being Tinutyarvik valley, western Adams Lake valley and northern Naksaluk valley, below the Nachvak glacial level. The latter may be Saglek weathering zone equivalents . Only relative dates for these zones can be determined on the basis of this study.

The Komaktorvik weathering zone, lying above the limit of the MI glacial phase may be older than the Wisconsinan. Support for this conclusion may be derived from the results of a preliminary analysis on the distribution of disjunct bryophytes in the central Nachvak Fiord area (T. Hedderson, personal communication, 1987). Initial results suggest that distributions are restricted to elevations above approximately 650 m asl. As the species analysed are slow to evolve, their ability to disperse outwards from refugia on the mountain summits is necessarily retarded, and therefore their present distribution may reflect approximate ice-free areas that existed throughout the entire Wisconsinan period.

The Koroksoak weathering zone is older than the Late Wisconsin, lying distal to the Nachvak glacial limit in the study area. The maximum age of this zone is considered to be equivalent to the age of the MI glacial phase which was estimated to be Early Wisconsinan and may be a correlative to the Kogalu Member on Baffin Island. This ice-proximal deposit is estimated to be $\geq$ 54 ka BP, but younger than approximately 110 ka BP (Miller, 1985). Fillon *et al.* (1981*b*) related high levels of sand input into the Labrador Sea at approximately 75 ka BP to glacial events on adjacent land masses. Andrews *et al.* (1983) recognized a major deglacial event in Hudson Strait and Hudson Bay also at 75 ka BP. The apparent consistency in the timing of this Early Wisconsinan event provides a basis for suggesting that the MI glacial phase and the Koroksoak weathering zone in the Nachvak Fiord region date from $\geq$ 75 ka BP (Table 7-3).

Table 7-3: Summary of glacial and sea level chronology for northern Labrador.

Glacial and sea level chronology for northern Labrador			
This Study	Clark (1984)	Löken (1962*b*)	Evans & Rogerson (1986)
Various sea levels below 15.5 m		Various sea levels below 15.5 m	
Tapes Transgression? SL-B 15.5 m 4.7 ka BP		Tapes Transgression? Sl-4 15.5 m 4.7 ka BP	
Tessersoak readvance Sl-D 19-41 m 9 ka BP	Two Loon and Coleman drifts SL 56 m *ca.* 9 ka BP	Kangalaksiorvik readvance SL-3 28-34 m 9 ka BP	Superguksoak I glacial phase SL 33m 5-12 ka BP
Townley Head still-stand Sl-E 23-48 m		Two Loon readvance? SL-2 31-45 m	
Kogarsok still-stand Sl-F 26-55 m <22 ka BP		Two Loon readvance? SL-2 31-45 m	
Shoal Water Cove readvance Sl-G 28-48 m		Noodleook readvance SL-1 41-56 m	
Glacial still-stands? Sl-H 37-53 m Sl-I 40-67 m Sl-J 47-68 m			
Nachvak glacial maximum SL $\geq$ 67 m ? *ca.* 20-24 ka BP			Nachvak Glaciation *ca.* 17-24 ka BP
Adams Lake glacial phase SL $\leq$ 73 m ? 29-50 ka BP	Iron Strand drift 33-50 ka BP		Late Ivitak subphase *ca.* 40 ka BP
MI glacial phase $\geq$ 75 ka BP			Early Ivitak subphase >70 ka BP

Clark (1984) also concluded that the Koroksoak weathering zone, as characterized by Iron Strand drift in northernmost Labrador, was Early Wisconsinan in age. However, he expressed concern regarding the regional context of this drift which he tentatively dated at 33-50 ka BP. Although the absolute dating of deposits older than the Late Wisconsin is extremely tentative, the similarity between estimated dates for the Iron Strand drift and the Adams Lake glacial phase (29-50 ka BP) may suggest that these two glacial events were coeval, and that the Iron Strand glaciation, as mentioned previously, may have resulted solely from local glacier activity during the Middle Wisconsin (Table 7-3). Comparable to these events is the Late Ivitak subphase in the Selamiut Range ($\geq$ 40 ka BP; Evans and Rogerson, 1986). Dreimanis and Raukas (1975), in their discussion of the Middle Wisconsin, provided evidence for a global cooling period during the time period 30-50 ka BP, and cited examples of glacial advances in North America and Eurasia.

With the possible exceptions of two high moraine segments in Kangalaksiorvik Fiord and Ryans Bay, the general absence of Early Wisconsinan glacial evidence (MI moraine equivalents) in northernmost Labrador is problematic, but may be explained in the context of Late Wisconsinan ice limits in that area. Clark and Josenhans (1986) have proposed an extensive Late Wisconsinan glaciation in northern Labrador with ice extending east onto the Labrador Shelf. Although this does not appear to be the case in Nachvak Fiord, there is greater potential for such an occurrence north and south of the study area (Clark and Josenhans, 1986). The Torngat Mountains are highest in the Nachvak Fiord region and therefore a considerable thickness of ice is required to breach the watershed and flow through the fiord. However, the mountain relief decreases substantially north and south of the fiord and it is proposed that Laurentide ice moving east from Labrador/Ungava ice dispersal centres would diverge around the highest part of the Torngat Range, and consequently larger volumes of ice would extend through Kangalaksiorvik and Saglek Fiords. This hypothesis may explain the apparent higher elevations of the Saglek weathering zone in both the above fiords [200-800 m aht in Kangalaksiorvik Fiord (Clark, 1984), 200-700 m aht in Saglek Fiord (Clark and Josenhans, 1986)]. The elevation of the Koroksoak weathering zone in Nachvak Fiord is 115-650 m aht, and therefore it is possible that the Late Wisconsinan glaciation was altitudinally more extensive than the Early Wisconsinan glaciation in areas north and south of the highest Torngat Mountain region.

Middle Wisconsinan glacial deposits in northernmost Labrador are generally distributed in coastal areas, between the major through-troughs which hosted outlet glaciers from the Laurentide ice sheet. The style of regional glaciation was therefore responsible for the preservation of these deposits in such locations during the Late Wisconsinan ice advance. If a similar glacial style can be evoked for the Early Wisconsinan glaciation of northern Labrador, then it may be assumed that these locations were ice- free during this period, and that Early Wisconsinan deposits were never present in these locations.

The above discussion provides a tentative solution to the apparent incompatibility of the Late Wisconsinan ice models proposed for northern Labrador by Clark and Josenhans (1986) and Evans and Rogerson (1986). However, this argument assumes that data used as evidence to describe an extensive Late Wisconsinan glaciation on the Labrador Shelf is unequivocal, although discussion of such evidence throughout this thesis suggests that further research is required in order to make this assumption.

The deglacial chronology proposed for Nachvak Fiord suggests that a stepwise recession of Laurentide ice characterized the late-glacial period. Similar ice-margin fluctuations have been proposed by Löken (1962*b*) for northernmost Labrador, by Andrews (1963) for the northern Nain-Okak area, and by Johnson (1985) for the Labrador coast in general. Comparisons with northernmost Labrador suggest that two major ice readvances occurred at 9 ka BP [Tessersoak/Kangalaksiorvik] and at >9 ka BP [Shoal Water Cove/Noodleook] (Table 7-3). In the Nachvak Fiord area, the latter readvance approached within several kilometres of the Late Wisconsinan ice maximum whereas in Eclipse Channel, northernmost Labrador, the shoreline associated with this readvance is the highest recorded marine evidence. Further south in Kangalaksiorvik Fiord, Clark (1984) suggests the marine limit at 56 m asl is 8700 years old, and was formed on the initiation of Late Wisconsinan ice retreat from the fiord. Clark's hypothesis may suggest that Late Wisconsinan ice was much thicker in Kangalaksiorvik Fiord [he suggests a much higher sea level at 9 ka BP] and that it persisted at the mouth of the fiord much later than in Eclipse Channel and Nachvak Fiord [Late Wisconsinan ice is suggested to have been at the Sheppard moraines (and their equivalents), well-inland, but east of the Labrador/Ungava drainage divide at this time].

Evans and Rogerson (1986) associated the 33 m asl marine limit in Ivitak Cove with the Superguksoak I glacial phase in the Selamiut Range. This marine limit was formed during the Tessersoak Readvance and therefore the Superguksoak phase may represent local glacier activity during the same period (Table 7-3). Evans and Rogerson (1986) soils-dated this phase at 5-12 ka BP but suggested it may be a correlative to the Kangalaksiorvik readvance at 9 ka BP. This study lends support to their tentative correlation as the Tessersoak readvance is also dated at 9 ka BP. This may indicate that a climatic factor was responsible for the readvance of Laurentide ice at this time.

7.6. Future research

The fiord coastline of northern Labrador has attracted many researchers seeking to unravel the complex movements of Laurentide ice in the eastern Canadian Arctic. These spectacular through-troughs undoubtedly held ice streaming towards the Labrador Sea during the Wisconsin, however, succeeding ice advances have eroded the traces of previous glaciation leaving the observer with an incomplete picture of prior glacial events. The results of this study indicate that distributary outlets for Laurentide ice, situated adjacent to the fiords, are potential areas for the identification of glacial and marine deposits older than the last ice advance. Although ice-free areas that existed during the Late Wisconsin have been mapped along the coastline of northern Labrador (*e.g.* Iron Strand), they do not provide an opportunity to examine Late Wisconsinan maximum deposits in an ice-marginal environment.

Drift pebble lithology and drift geochemistry were used in this study to distinguish Late Wisconsinan regional ice deposits from Middle Wisconsinan local ice deposits. The key factor in these analyses was the location of the sedimentary Ramah Group along the outer coast of the fiord. The recognition of distinctly deformed and mylonitized gneisses from west of the Ramah Group in drift sampled on, and east of the Ramah Group provided unequivocal evidence for the former presence of regional fiord ice. The Ramah Group forms a linear north-trending fold belt between Nachvak Fiord and Hebron Fiord, 100 km to the south, and therefore these same lithologies can be used to map the limit and flow direction of regional ice throughout this coastal area. It is also suggested that the geochemical properties of regional and local drift sheets can be discerned on the basis of their metal concentrations, and consequently drift sediment samples only may be required to detect the former presence of regional ice. However, to employ only one characteristic to identify regional deposits is to underestimate the complex interactions of local and regional ice in the fiord landscapes. Clark and Josenhans (1986) used morphological criteria to discern Late Wisconsinan maximum moraines from older moraines but, as illustrated in this study, their criteria are not necessarily exclusive to either set of moraines (Naksaluk moraine *v.* Adams Lake moraine).

The model of relative sea level change proposed for Nachvak Fiord requires control elevations on raised marine shorelines situated north and south of the study area. Although there are many reports on shoreline elevations in the literature, it is imperative for the successful expansion of this model that shoreline sequences are used, instead of isolated marine limits. This requirement is stressed in view of the limited number of radiocarbon dated shorelines in northern Labrador, and the disparate character of deglaciation along the coastline. The adaption of techniques employed to correlate fragmented shorelines in the study area may prove successful in other areas of northern Labrador, but it is emphasized that an important pre-requisite of this approach is a reliable and accurate data set from which shorelines can be resolved.

Onshore - offshore correlation of sedimentary facies cannot be traced directly through acoustic reflection techniques because of the shallow inner shelf which contains only a thin veneer of sediments reworked by wave and iceberg scouring. Instead, this correlation may be achieved through identification of pollen zones in core sediment stratigraphy, and the use of lithology to characterize particular sediment units (Josenhans *et al.*, 1986). However, both these techniques are limited by core depth, and considering the potential errors in extrapolating data down-section, it may be more relevant to use such information to test existing land-based hypotheses, rather than presenting ice models based primarily on sampled and interpreted marine data.

The mapping of fiord and lacustrine acoustic units onto the land should be an integral part in the reconstruction of glaciomarine and glaciolacustrine environments. The identification of specific stratigraphical markers related to terrestrial events, as documentated in Adams Lake, provides essential control for acoustic interpretation. In addition, palynology and radiocarbon dating of core samples provide a tentative framework for late-glacial and postglacial events.

Aminostratigraphical analysis is at an early stage in northern Labrador and requires a greater age-range of samples to ensure accurate correlation between sites. The apparent discrepancies between alle/Ile ratios reported for Nachvak Fiord and Iron Strand require further investigation, if it is accepted that the sedimentary units in question are of a similar age.

Lastly, to reiterate the remarks of Johnson (1985), it is suggested that further localized studies should be carried out along the northern Labrador coast to build upon the present framework of local and regional glacial chronologies and to critically examine proposed models of glaciation in northern Labrador.

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Appendix A

Drift Pebble Lithology Data

Table A-1: Drift pebble lithology data showing sample site number, lithological composition (%) by pebble class, number of pebbles in each sample, and cluster membership.

Drift Pebble Lithology Results (%)							
Site	Class A	Class B	Class C	Class D	Class E	$n=$	Cluster
L01	00.0	00.1	66.0	23.0	10.5	209	P
L02	00.0	38.6	25.1	21.9	14.4	215	N
L03	00.0	51.8	26.1	16.1	06.0	199	N
L04	00.0	08.3	72.0	17.6	02.1	193	P
L05	00.0	10.4	74.5	13.0	02.2	184	P
L06	00.0	04.9	72.1	13.5	09.6	208	P
L07	00.0	04.0	67.6	09.3	19.2	182	P
L08	00.0	00.0	78.7	15.9	05.5	183	P
L09	00.0	04.2	50.5	33.7	11.7	214	P
L10	00.0	01.5	58.5	11.5	28.5	200	P
L11	00.0	00.0	45.2	20.1	34.7	199	P
L12	00.0	05.8	53.1	37.5	03.7	192	P
L13	00.0	02.7	80.5	14.2	02.7	226	P
L14	00.0	02.8	77.0	06.9	13.4	217	P
L15	00.0	00.0	75.3	23.7	01.1	186	P
L16	00.0	03.2	84.0	07.5	05.3	188	P
L17	00.0	00.0	91.7	08.3	00.0	216	P
L18	00.0	11.4	34.2	07.4	47.0	149	P
L19	00.0	100	00.0	00.0	00.0	235	M
L20	00.0	90.5	06.2	03.3	00.0	211	M
L21	00.0	35.7	55.8	08.5	00.0	199	P
L22	00.0	77.3	12.7	05.5	04.6	220	N
L23	00.0	96.4	03.2	00.5	00.0	216	M
L24	00.0	77.0	00.5	22.2	00.5	221	M
L25	00.0	86.8	00.0	06.9	06.4	203	M
L26	00.0	99.1	00.9	00.0	00.0	107	M
L27	00.0	94.7	01.4	03.9	00.0	208	M
L28	00.0	24.6	34.2	09.1	32.2	199	P
L29	00.0	69.8	08.9	00.0	21.3	202	N
L30	00.0	24.4	49.5	21.7	04.4	184	P
L31	00.0	58.9	27.5	00.5	13.2	204	N
L32	00.0	60.5	16.0	03.9	14.6	206	N
L33	00.0	48.1	30.5	21.4	00.0	187	N
L34	11.3	52.0	13.2	11.8	11.8	212	N

Site	Class A	Class B	Class C	Class D	Class E	$n=$	Cluster
L35	00.0	68.2	27.0	01.0	03.9	204	N
L36	00.0	96.3	03.3	00.5	00.0	213	N
L37	00.0	95.5	00.9	00.0	03.6	221	M
L38	89.0	08.0	00.0	03.0	00.0	100	O
L39	72.5	05.1	00.0	18.8	03.6	138	O
L40	69.7	11.4	04.7	14.2	00.0	211	O
L41	47.9	22.0	04.8	25.4	00.0	209	O
L42	30.9	25.5	04.9	38.6	00.0	223	O
L43	19.7	34.7	23.0	04.2	18.3	213	N
L44	36.2	48.2	00.0	15.6	00.0	199	N
L45	47.2	10.7	00.0	38.1	04.1	197	O
L46	32.4	64.7	00.0	02.9	00.0	204	N
L47	41.5	43.4	00.0	15.1	00.0	205	N
L48	40.9	46.6	03.4	09.1	00.0	208	N
L49	38.2	50.5	00.0	11.3	00.0	212	N
L50	31.3	54.8	06.8	07.3	00.0	192	N
L51	17.4	71.5	04.6	00.5	06.1	196	N
L52	15.8	48.4	12.6	12.1	11.1	190	N
L53	38.1	42.5	03.9	02.4	13.2	205	N
L54	37.6	55.3	00.0	03.6	03.6	197	N
L55	26.2	57.8	08.7	06.0	01.4	218	N
L56	40.2	40.2	01.9	16.4	01.4	214	N
L57	00.0	99.2	00.8	00.0	00.0	131	M
L58	00.0	99.2	00.8	00.0	00.0	127	M
L59	00.0	00.0	75.0	25.0	00.0	200	P

Appendix B
Drift Geochemistry Data

Table B-1: Drift geochemistry data showing site number, cluster membership and sample composition.

Geochemistry Results 1985 <ppm>[34]										
Site	Cluster	Cr	Mn	Fe	Co	Ni	Cu	Zn	Pb	Ufl
G01	P	38	150	2.4	10	28	35	30	06	0.4
G02	N2	67	520	3.9	33	66	78	76	10	0.5
G03	N2	40	280	3.2	15	34	49	62	04	0.5
G06	P	32	200	2.2	11	27	38	40	03	0.2
G07	P	37	150	2.5	10	29	42	45	04	0.3
G09	P	44	210	3.0	12	39	50	60	03	0.4
G10	P	42	210	2.4	15	47	58	40	04	0.2
G11	P	40	220	2.3	14	39	43	36	04	0.2
G13	P	48	148	2.1	12	47	42	36	09	0.2
G14	P	35	158	2.1	12	38	44	32	02	0.3
G15	P	30	125	1.7	09	28	32	23	03	0.2
G16	P	57	220	3.2	22	60	101	52	08	0.5
G17	P	38	130	2.2	09	40	35	31	03	0.5
G18	P	50	620	4.1	18	49	105	75	09	0.2
G19	M	32	580	5.0	29	57	89	75	19	0.6
G20	M	36	960	5.8	32	63	100	95	21	0.9
G22	N2	26	460	4.0	22	44	73	60	16	0.6
G23	M	32	650	4.7	23	50	63	110	14	0.8
G25	M	32	1200	6.2	29	51	103	122	17	0.9
G26	M	55	370	12.0	18	77	195	260	30	1.4
G27	M	32	700	5.6	21	52	51	82	09	0.6
G28	P	42	900	4.7	28	64	102	120	10	0.2
G29	N2	55	3400	7.0	26	64	139	178	17	1.3
G30	P	41	290	2.8	16	61	74	44	06	0.1
G31	N2	65	960	6.8	39	86	161	220	20	1.2
G32	N2	81	1300	7.3	36	99	154	230	20	0.9
GAA	N2	86	920	13.0	40	123	341	400	35	1.7
G33	N2	63	840	8.5	40	89	304	220	23	1.4
G34	N2	75	1800	8.9	43	100	187	300	24	1.3
G35	N2	74	880	8.0	24	88	197	240	23	1.0

[34] Fe units = %

Site	Cluster	Cr	Mn	Fe	Co	Ni	Cu	Zn	Pb	Ufl
G36	M	25	520	4.0	21	70	100	135	08	1.3
G37	M	54	310	13.0	14	79	144	135	26	3.9
G40	O	35	460	3.3	19	56	156	57	11	0.6
G41	O	31	240	2.5	12	27	51	40	07	0.4
GBB	O	45	340	4.6	13	21	42	60	12	0.6
GCC	O	47	270	3.8	16	19	48	70	07	0.4
G42	O	44	450	4.0	26	57	136	95	14	0.6
G43	N1	115	740	5.8	33	106	149	135	19	0.5
G47	N1	53	310	3.7	12	35	35	80	07	0.4
G48	N1	40	280	2.8	12	32	35	55	07	0.6
G49	N1	32	300	3.0	16	36	46	55	08	0.6
G50	N1	47	480	4.4	22	34	61	92	12	0.2
G51	N1	28	400	3.4	19	32	50	68	08	0.1
GDD	N1	55	500	5.2	31	62	106	126	18	0.3
G52	N1	44	390	4.2	21	50	75	94	12	0.2
G53	N1	22	240	2.6	13	22	34	43	09	0.1
G54	N1	20	240	2.6	12	20	33	44	06	0.1
G55	N1	23	280	3.1	22	23	71	55	14	0.6
Geochemistry Results 1984										
G04	P	91	525	6.3	32	94	144	175	14	2.7
G05	[illegible]	94	460	5.6	25	100	160	156	12	0.9
G08	[illegible]	90	400	5.1	36	144	174	152	23	0.8
G11	[illegible]	96	440	4.6	30	136	205	114	05	0.3
G20		110	2200	6.2	80	180	235	144	55	1.4
G34		98	1000	8.3	54	205	300	425	38	1.4
G44		90	610	6.1	26	100	150	154	20	0.9
G45		80	1100	7.2	44	98	215	190	30	1.3
G46		88	620	4.8	25	83	118	136	18	1.2
G59		140	560	4.3	49	300	310	100	05	0.7
Geochemistry Results 1983[35]										
IVI	See Text	103	988	4.7	35	115	113	75	08	1.6
NAC	See Text	142	1394	5.9	54	245	307	172	10	1.3
SUP	See Text	155	1130	6.2	46	193	333	128	09	1.4

[35] Evans, 1984

Table B-2: Concentrations of Mo and Ag in 1985 and 1984 drift samples

Sample concentrations of Mo and Ag <1985>								
Sample	Mo	Ag	Sample	Mo	Ag	Sample	Mo	Ag
G01	<1	0.1	G02	<1	0.1	G03	<1	<0.1
G06	<1	<0.1	G07	<1	<0.1	G09	<1	0.1
G10	<1	<0.1	G11	<1	<0.1	G13	<1	<0.1
G14	<1	<0.1	G15	<1	<0.1	G16	<1	<0.1
G17	<1	0.1	G18	2	0.1	G19	<1	<0.1
G20	<1	<0.1	G22	<1	0.2	G23	1	0.2
G25	3	0.1	G26	46	0.2	G27	<1	<0.1
G28	5	0.1	G29	5	<0.1	G30	<1	0.2
G31	10	<0.1	G32	5	<0.1	GAA	18	0.2
G33	7	0.1	G34	5	0.1	G35	13	0.1
G36	1	0.2	G37	16	0.1	G40	<1	0.1
G41	<1	0.1	GBB	2	0.1	GCC	1	0.1
G42	4	0.1	G43	3	<0.1	G47	5	0.1
G48	5	<0.1	G49	<1	0.1	G50	<1	0.2
G51	<1	0.1	GDD	4	0.1	G52	2	0.1
G53	<1	<0.1	G54	<1	<0.1	G55	3	0.1
Sample concentrations of Mo and Ag <1984>								
G04	6	0.2	G05	2	0.2	G08	3	<0.1
G11	4	0.1	G20	5	0.5	G34	11	1
G44	3	0.4	G45	6	0.2	G46	5	0.2
G59	3	0.3						

Table B-3: List of discriminant scores calculated for each sample in the 1985 data set.

Discriminant scores calculated for each sample				
Sample	Function 1	Function 2	Function 3	Function 4
G01	-1.9520	-0.6798	-0.6628	-0.1543
G02	-0.1058	+0.8148	+0.3871	-0.7329
G03	-1.2867	-0.1881	+0.5068	+0.0799
G06	-1.7906	-0.6097	+0.2257	+0.0950
G07	-2.0420	-0.5497	+0.1049	+0.0216
G09	-1.7471	-0.6062	+1.0830	-0.0949
G10	-1.3537	-1.0858	1.1737	-0.0312
G11	-1.4419	-0.8631	+0.5673	-0.2064
G13	-1.6807	-1.0940	-0.3948	-1.2095
G14	-1.4783	-1.2509	+1.1579	+0.0818
G15	-1.8584	-1.1528	+0.1917	-0.0391
G16	-1.7257	-0.2244	+0.8780	+0.5796
G17	-1.4757	-1.6382	+1.0120	-0.4030
G18	-1.6740	+0.0406	+0.3861	+0.9883
G19	+2.8547	-0.7524	-1.5486	-0.1100
G20	+3.4557	-0.5018	-1.5386	-0.1977
G22	+1.5655	-0.8599	-1.6671	-0.0387
G23	+2.1189	-0.4262	-0.5141	-1.0700
G25	+2.7150	+0.2104	-0.9096	+0.4061
G26	+3.3303	+0.0596	-0.6536	+0.7461
G27	+2.9815	-1.7204	+1.0081	-0.2188
G28	+1.3130	+0.0308	+1.4400	-0.1029
G29	+0.7503	+1.1822	+0.1648	+0.1656
G30	-0.3223	-1.9604	+1.7939	-0.0781
G31	+1.6699	+2.2574	+0.6277	-0.9646
G32	+1.4574	+1.7687	+1.3797	-1.8644
GAA	+1.9275	+3.7474	+1.5366	+1.3207
G33	-0.6030	+2.9429	+0.9855	+4.1604
G34	+2.8124	+3.0860	+1.1153	-1.6437
G35	-0.2960	+1.5775	+0.6247	-0.2840
G36	+2.0647	-1.5508	+2.9417	-0.6341
G37	+5.6158	-3.5020	+0.0101	+1.9598

Sample	Function	Function 2	Function 3	Function 4
G40	-1.7218	-0.6654	+0.6535	+2.4433
G41	-1.6191	-0.4683	-0.7563	+0.2865
GBB	-1.2363	+0.4476	-2.4591	+0.0471
GCC	-2.2839	+1.3647	-1.4315	+0.3215
G42	-0.7705	+0.6201	-0.2523	+1.0238
G43	-2.0151	+1.5098	+0.6531	-1.5013
G47	-1.7657	+0.1858	-0.3395	-1.0269
G48	-1.4259	-0.3853	-0.5453	-0.6826
G49	-0.2545	-0.6684	-0.3664	-0.3333
G50	-0.5638	+1.1788	-1.5970	-0.3476
G51	+0.209	-0.1711	-0.4825	-0.0465
GDD	+0.828	+1.1693	-1.1276	-0.6939
G52	+0.161	-0.0239	-0.3157	-0.5451
G53	-0.463	-0.5758	-1.5190	-0.2000
G54	-0.5976	-0.6630	-0.7708	+0.1227
G55	-0.280	+0.7053	-2.7565	+0.6055

Table B-4: Comparison of predicted and probable <1985> sample classification as derived from the discriminant analysis classification procedure.

Predicted and probable sample classification[36,37] <1985>						
	Pred.	1st Prob.		2nd Prob.		
Sample	Group	Group	$\{P(G/X)\}$	Group	$\{P(G/X)\}$	$\{P(X/G)\}$
G01	P ***	O	0.3734	P	0.3710	0.7254
G02	N2	N2	0.4305	N1	0.4106	0.7231
G03	P ***	P	0.5941	O	0.1986	0.9773
G06	P	P	0.6409	O	0.2244	0.9813
G07	P	P	0.6026	O	0.2631	0.9359
G09	P	P	0.8684	O	0.0718	0.9831
G10	P	P	0.9161	N1	0.0427	0.9865
G11	P	P	0.7916	N1	0.1132	0.9998
G13	P	P	0.5787	N1	0.3249	0.6268
G14	P	P	0.9286	O	0.0379	0.9767
G15	P	P	0.7654	O	0.1399	0.9699
G16	P	P	0.7054	O	0.2147	0.9190
G17	P	P	0.9460	N1	0.0329	0.9210
G18	P ***	O	0.4842	P	0.4050	0.8048
G19	M	M	0.9938	N1	0.0040	0.7066
G20	M	M	0.9983	N2	0.0010	0.6642
G22	N2 ***	M	0.6662	N1	0.2899	0.3047
G23	M	M	0.8735	N1	0.0658	0.5699
G25	M	M	0.9351	N2	0.0543	0.6686
G26	M	M	0.9888	N2	0.0105	0.7606
G27	M	M	0.9990	N2	0.0008	0.7426
G28	P ***	N2	0.7160	M	0.1531	0.4308
G29	N2	N2	0.8251	N1	0.1345	0.9835
G30	P		0.9737	N1	0.0172	0.4477
G31	N2	N2	0.9892	N1	0.0087	0.7103
G32	N2	N2	0.9848	N1	0.0120	0.3220
GAA	N2	N2	0.9999	N1	0.0001	0.0779##
G33	N2	N2	0.6835	O	0.3115	0.0003##

[36] *** denotes wrongly predicted group.

[37] ## denotes anomalous sample.

	Pred.	1^{st} Prob.		2^{nd} Prob.		
Sample	Group	Group	$\{P(G/X)\}$	Group	$\{P(G/X)\}$	$\{P(X/G)\}$
G34	N2	N2	0.9976	M	0.0020	0.0546##
G35	N2	N2	0.7310	N1	0.1784	0.8654
G36	M	M	0.9214	N2	0.0399	0.0210##
G37	M	M	1.0000	N2	0.0000	0.0034##
G40	O	O	0.5557	P	0.4237	0.2168##
G41	O	O	0.4919	N1	0.2637	0.9335
GBB	O	O	0.5022	N1	0.4867	0.5065
GCC	O	O	0.7893	N1	0.1910	0.6650
G42	O	O	0.5603	N1	0.2583	0.8947
G46	N1	N1	0.4562	P	0.2963	0.1262##
G47	N1	N1	0.4926	P	0.2884	0.7244
G48	N1	N1	0.4651	P	0.3156	0.8701
G49	N1	N1	0.5564	P	0.2911	0.8969
G50	N1	N1	0.7021	O	0.2374	0.8203
G51	N1	N1	0.6303	O	0.1344	0.9100
GDD	N1	N1	0.5469	N2	0.4102	0.5546
G52	N1	N1	0.6652	P	0.1260	0.9306
G53	N1	N1	0.6885	O	0.2308	0.8879
G54	N1	N1	0.4920	O	0.2761	0.9010
G55	N1	N1	0.5378	O	0.4507	0.2788

Table B-5: List of discriminant scores calculated for each sample in the 1984 data set.

Discriminant score	
Sample	Function 1
G04	-0.6185
G05	-0.5064
G08	+0.3515
G11	-1.6958
G20	+2.2036
G34	+1.3604
G44	+0.4929
G45	+0.8438
G46	+0.2833
G59	-2.7146
EVS	-1.5691
EVS	-2.3247
EVS	-2.0273

Table B-6: Comparison of predicted and probable <1984> sample classification as derived from the discriminant analysis classification procedure.

Predicted and probable sample classification[38,39] <1984>						
	Pred.	1^{st} Prob.		2^{nd} Prob.		
Sample	Group	Group	$\{P(G/X)\}$	Group	$\{P(G/X)\}$	$\{P(X/G)\}$
G04	P	P	0.7829	Z	0.2171	0.67582
G05	P	P	0.7408	Z	0.2592	0.5958
G08	P ***	Z	0.6745	P	0.3255	0.4931
G11	P	P	0.9711	Z	0.0289	0.5099
G20	Z	Z	0.9897	P	0.0103	0.2433##
G34	Z	Z	0.9438	P	0.0562	0.7462
G44	Z	Z	0.7354	P	0.2646	0.5865
G45	Z	Z	0.8519	P	0.1481	0.8469
G46	Z	Z	0.6428	P	0.3572	0.4511
G59	P	P	0.9964	Z	0.0036	0.0934##
EVS	UNGRPD	P	0.9628	Z	0.0372	0.5945
EVS	UNGRPD	P	0.9920	Z	0.0080	0.1978##
EVS	UNGRPD	P	0.9853	Z	0.0147	0.3219

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38 *** denotes wrongly predicted group.

39 ## denotes anomalous sample.

Appendix C
Shoreline Data

Table C-1: Location, elevation and residual value for each shoreline segment observed in the study area.

Shoreline segment information					
Segment	Easting	Northing	Elevation	Residual	Traverse
A01	479522	6542773	08.992	+0.398	06
A02	479995	6542032	07.911	-0.702	09
A03	480188	6543305	09.176	+0.636	05
A04	486306	6546324	08.286	+0.149	20
A05	474928	6546899	09.178	+0.609	12
A06	474453	6547064	08.374	-0.206	13
A07	469560	6542174	09.359	+0.330	17
A08	469390	6543015	08.675	-0.317	16
A09	486923	6545503	08.127	-0.027	21
A10	480950	6543411	08.094	-0.410	32
A11	476228	6548004	08.000	-0.460	ALT(x)
B01	479524	6543021	15.401	-0.112	02
B02	481135	6543492	15.615	+0.078	04
B03	482156	6543472	15.388	-0.172	03
B04	478935	6541760	15.810	+0.279	ALT(11)
B05	483777	6544040	15.693	+0.111	20
B06	485307	6545046	15.802	+0.211	20
B07	485976	6545902	15.179	-0.406	24
B08	474296	6547049	15.228	-0.070	13
B09	475103	6546976	15.648	+0.330	12
B10	469444	6543015	15.041	-0.249	16
C01	485939	6545998	17.190	+0.090	20
C02	486124	6546300	17.181	+0.177	20
C03	484913	6543621	17.244	-0.553	19
C04	486923	6545508	17.000	-0.095	ALT(21)
C05	476265	6547942	18.000	+0.162	ALT(x)
C06	474032	6546948	18.249	-0.104	13
C07	475028	6546955	18.178	-0.050	12
C08	479990	6541887	19.343	+0.519	09
C09	469750	6542050	19.943	-0.116	16
C10	469752	6542945	19.815	-0.030	16
D01	477784	6541965	26.300	+0.345	06

Segment	Easting	Northing	Elevation	Residual	Traverse
D02	477870	6542266	26.233	+0.677	06
D03	478398	6542520	25.113	+0.151	06
D04	478562	6542600	24.260	-0.517	08
D05	478774	6542694	23.807	-0.741	06
D06	479007	6542810	23.417	-0.865	07
D07	479148	6542872	23.521	-0.609	06
D08	479485	6543090	23.193	-0.493	02
D09	480158	6543478	22.866	+0.011	05
D10	481129	6543657	21.724	-0.377	04
D11	482202	6543700	21.572	+0.125	03
D12	478936	6541562	26.920	+1.142	11
D13	479459	6541690	24.418	-0.917	10
D14	479991	6541839	24.202	-0.660	09
D15	470348	6541590	31.973	+1.401	17
D16	470589	6541835	30.836	+0.686	16
D17	470221	6542315	29.002	-0.795	16
D18	470435	6545697	25.244	-0.485	18
D19	482708	6546151	19.000	+0.698	ALT(y)
D20	484943	6543626	20.599	+0.606	19
D21	483753	6544038	19.867	-0.314	20
D22	484179	6544509	20.078	+0.686	20
D23	484337	6544718	19.579	+0.520	20
D24	484485	6544875	18.519	-0.274	20
E01	477533	6541906	30.024	+0.257	06
E02	478528	6542630	29.069	+0.246	08
E03	479015	6542949	28.772	+0.392	07
E04	479427	6543190	28.368	+0.348	02
E05	482198	6543732	27.031	+0.845	03
E06	479463	6541593	28.360	-0.502	10
E07	479978	6541790	29.428	+0.959	09
E08	483720	6544054	25.863	+0.697	20
E09	484940	6543619	24.482	-0.239	19
E10	485163	6543933	23.597	-0.831	21
E11	484461	6544889	23.208	-1.095	20
E12	485220	6545298	22.812	-0.848	20

Segment	Easting	Northing	Elevation	Residual	Traverse
E13	485744	6545836	22.901	-0.177	31
E14	486017	6546248	22.829	+0.125	20
E15	482709	6546143	25.000	+0.399	ALT(y)
E16	476271	6547906	26.500	-0.732	ALT(x)
E17	470448	6545699	31.659	-0.004	18
E18	469755	6542969	32.707	-0.815	16
E19	470510	6542216	33.775	+0.267	16
E20	469572	6542004	32.887	-1.257	17
E21	470140	6541656	34.160	+0.144	17
E22	470307	6541517	33.677	-0.321	17
E23	474150	6545550	31.825	+2.142	13
F01	478978	6543027	34.662	+0.986	07
F02	480140	6543623	33.428	+1.063	05
F03	476388	6547776	33.500	+0.177	ALT(x)
F04	470531	6545671	38.876	-0.575	18
F05	469772	6542962	42.708	+1.150	ALT(16)
F06	470870	6541764	40.412	-0.857	16
F07	470587	6541183	41.167	-0.656	17
F08	483720	6544064	28.865	-0.210	20
F09	485209	6545391	26.472	-0.619	20
F10	485183	6543791	28.355	+0.380	21
F11	477452	6541894	34.250	-1.338	06
F12	478494	6542744	33.569	-0.673	08
G01	474235	6545490	37.721	-2.171	13
G02	475822	6541873	38.373	-1.402	13
G03	476433	6541859	37.523	-1.598	12
G04	476280	6547788	36.500	-0.170	ALT(x)
G05	475497	6541297	40.848	+0.468	09
G06	472605	6541015	44.237	+0.610	15
G07	472873	6540719	43.071	-0.397	14
G08	469564	6541930	47.602	+1.096	17
G09	470956	6541673	45.881	+0.764	16
G10	470527	6545679	44.891	+1.079	18
G11	482213	6543801	32.597	+0.574	03
G12	485711	6545850	28.251	+0.909	30

Segment	Easting	Northing	Elevation	Residual	Traverse
G13	484985	6543712	29.308	+0.238	19
H01	478471	6542786	43.281	+0.014	08
H02	479739	6543540	42.428	+0.336	33
H03	480631	6543820	40.915	-0.472	33
H04	482224	6543838	39.582	-0.781	03
H05	479984	6541696	43.824	+0.991	09
H06	478933	6541353	43.483	-0.186	11
H07	483681	6544122	39.093	-0.204	20
H08	485191	6543735	38.070	-0.452	19
H09	485895	6546221	37.075	+0.210	20
H10	475836	6541914	46.678	+1.308	13
H11	485839	6546034	36.843	-0.148	27
H12	482503	6546126	39.000	-0.073	ALT(y)
H13	476434	6547761	43.000	+0.854	ALT(x)
H14	470563	6545673	46.604	-0.299	18
H15	470412	6542421	47.436	-1.145	ALT(16)
H16	471390	6541304	48.497	-0.004	16
H17	470880	6540987	49.032	+0.052	17
I01	478438	6542827	48.780	-0.280	08
I02	478920	6543128	47.819	-0.607	07
I03	479514	6541442	47.637	-0.430	10
I04	479982	6541644	48.436	+0.967	09
I05	476442	6542009	52.601	+1.002	12
I06	475828	6541980	53.467	+1.133	13
I07	470611	6545662	57.772	-0.001	18
I08	472149	6540974	57.020	+0.110	16
I09	474250	6545431	53.631	+0.133	13
I10	485222	6545648	40.931	+0.508	20
I11	484376	6544952	40.520	-1.051	20
I12	472875	6540690	54.622	-1.484	14
J01	478427	6542866	58.633	+0.272	08
J02	480111	6543781	55.521	-0.250	05
J03	478921	6541187	57.414	-0.434	11
J04	479551	6541308	57.687	+0.781	10
J05	475833	6542029	60.377	-1.903	13

Segment	Easting	Northing	Elevation	Residual	Traverse
J06	485248	6545689	47.311	-0.669	20
J07	485561	6545987	47.336	-0.146	20
J08	485787	6546208	47.740	+0.618	20
J09	472810	6540669	68.000	+1.106	ALT(14)
J10	473426	6540842	66.000	+0.033	ALT(14)
J11	482203	6546059	53.000	+0.592	ALT(y)
K01	479256	6543537	70.585	+0.092	02
K02	480076	6543854	69.828	-0.216	05
K03	481116	6544142	68.110	-1.360	04
K04	478909	6541012	70.563	-0.024	11
K05	479571	6541224	69.660	-0.563	10
K06	476402	6542329	73.110	+1.058	12
K07	475827	6542130	71.387	-0.980	13
K08	485141	6545761	67.733	+0.464	22
K09	485342	6546036	67.587	+0.420	23
K10	476784	6541152	72.900	+1.111	ALT(9)
X01	480150	6542749	12.521	—	06
X02	479946	6541988	13.284	—	09
X03	469593	6542239	13.145	—	17
X04	484871	6543593	13.430	—	19
X05	483810	6543993	13.142	—	20
X06	476212	6548059	14.000	—	ALT(x)
X07	482814	6548155	13.500	—	ALT(y)
X08	470607	6541205	38.196	—	17
X09	470416	6542387	37.436	—	16
X10	470897	6541772	37.070	—	16
X11	469505	6541875	54.925	—	17
Y01	482249	6543917	52.042	—	03
Y02	476400	6542380	75.567	—	12
Y03	475830	6542250	75.166	—	13
Y04	485205	6543750	45.636	—	19

Table C-2: Comparison of predicted shoreline elevations with observed elevations at selected sites along the fiord coastline. [A] - refers to altimeter measurement

Comparison of predicted and observed shoreline elevations west of the study area							
Gloom Cove							
	Sl-C	Sl-D	Sl-E	Sl-F	Sl-G	Sl-H	Sl-I
Pred.		**29.[illegible]47 m**					
Obs.		**31.[illegible] m[A]**					
Ivitin Cove							
Pred.			37.338 m				
Obs.			36.5 m[A]				
Ivitak Cove							
Pred.		34.[illegible]78 m				54.970 m	70.980 m
Obs.		33.[illegible] m[A]				53.0 m[A]	67.0 m[A]
Eskimo Cove							
Pred.	21.440 m		41.245 m	53.453 m			
Obs.	21.116 m		39.789 m	49.425 m			
Green Point							
Pred.	21.685 m			56.254 m			
Obs.	18.0 m[A]			52.0 m[A]			
Bay Cove							
Pred.	21.037 m		41.379 m	54.954 m			
Obs.	20.0 m[A]		42.0 m[A]	55.0 m[A]			
Tallek Arm site #1							
Pred.				58.009 m			
Obs.				>48 m[A]			
Tallek Arm site #2							
Pred.			48.308 m	62.025 m			
Obs.			48.0 m[A]	55.0 m[A]			
Kogarsok Brook Cove							
Pred.	21.176 m	34.[illegible]6 m	42.253 m				
Obs.	21.064 m	34.[illegible]4 m	42.537 m				
West Cove							
Pred.	21.040 m	34.[illegible]6 m	42.509 m				
Obs.	21.137 m	33.[illegible]7 m	41.835 m				

	Sl-C	Sl-D	Sl-E	Sl-F	Sl-G	Sl-H	Sl-I
Townley Head							
Pred.	21.677 m						
Obs.	19.781 m						
Moraine Point							
Pred.	21.445 m		44.047 m				
Obs.	24.0 m[A]		46.0 m[A]				
Tasiuyak Arm site							
Pred.	22.317 m	40.231 m					
Obs.	20.0 m[A]	41.0 m[A]					

Appendix D
Soil profile descriptions

Table D-1: Description of soil profiles observed on moraine crests in study area

Soil pit	Depth (cm)	Horizon	Colour
Adams Lake moraine			
(1)	0 - 2	A	10 YR 4/2
	2 - 31	B	10 YR 3/3
	31 - $\geq$ 68	Cox	10 YR 4/6
(2)	0 - 12	O/A	5 YR 2.5/2
	12 - 41	B	10 YR 3/3
	41 - $\geq$ 61	Cox	10 YR 4/6
Quartzite moraine			
(1)	0 - 1	A	
	1 - 24	B	10 YR 5/3
	24 - 36	Cox	10 YR 5/2
	36 - $\geq$ 55	Cn	2.5 Y 5/2
(2)	0 - 11	A	10 YR 5/6
	11 - 31	B	10 YR 5/2
	31 - 57	Cox	10 YR 6/3
	37 - $\geq$ 51	Cn	10 YR 6/2
(3)	0 - 3	A	
	3 - 27	B	10 YR 3/2
	27 - 57	Cox	10 YR 5/3
	57 - $\geq$ 90	Cn	10 YR 5/2
(4)	0 - 17	A	5 YR 2.5/2
	17 - 37	B	10 YR 5/2
	37 - 65	Cox	10 YR 4/6
	65 - $\geq$ 97	Cn	10 YR 5/2
(5)	0 - 7	A	
	7 - 30	B	7.5 YR 4/4
	30 - 57	Cox	10 YR 5/3
	57 - $\geq$ 72	Cn	10 YR 5/2
Kammarsuit Valley West moraine			
(1)	0 - 13	A	10 YR 2/2
	13 - 42	B	5 YR 3/2
	42 - $\geq$ 49	Cox	10 YR 5/6

Soil pit	Depth (cm)	Horizon	Colour
Kammarsuit Valley East moraine			
(1)	0 - 14	A	5 YR 2.5/2
	14 - 48	B	10 YR 3/3
	48 - $\geq$ 81	Cox	10 YR 4/6
Naksaluk moraine			
(1)	0 - 18	B	2.5 YR 2.5/2
	18 - $\geq$ 48	Cox	10 YR 4/6
(2)	0 - 6	O/A	10 YR 2/1
	6 - 21	B	10 YR 4/4
	21 - $\geq$ 48	Cox	10 YR 5/6
(3)	0 - 2	A	
	2 - 23	B	
	23 - $\geq$ 67	Cox	
(4)	0 - 3	O/A	
	3 - 30	B	
	30 - $\geq$ 59	Cox	
(5)	0 - 2	A	
	2 - 23	B	
	23 - $\geq$ 63	Cox	
(6)	0 - 3	O/A	
	3 - 23	B	
	23 - $\geq$ 68	Cox	
(7)	0 - 8	A_0	10 YR 2/1
	8 - 28	B	10 YR 2/2
	28 - 47	Cox	10 YR 5/3
	47 - $\geq$ 56	Cn	2.5 Y 4/2
(8)	0 - 5	O_1	5 YR 2.5/2
	5 - 12	A	10 YR 2/2
	12 - 37	B	7.5 YR 3/4
	37 - 52	Cox	10 YR 4/6
	52 - $\geq$ 60	Cn	10 YR 4/4