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THE LATE FOXE DEGLACIATION OF THE BURTON BAY AREA,
SOUTHEASTERN BAFFIN ISLAND, N.W.T.

by

Cynthia Anne Squires

A Thesis submitted to the Faculty of Graduate Studies and Research through the Department of Geography in Partial Fulfillment of the requirements for the Degree of Master of Arts at the University of Windsor

Windsor, Ontario, Canada

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ABSTRACT

THE LATE FOXE DEGLACIATION OF THE BURTON BAY AREA,
SOUTHEASTERN BAFFIN ISLAND, N.W.T.

by

Cynthia Anne Squires

A detailed chronology of deglaciation is presented
for an area on the northwest shore of upper Frobisher Bay,
where the multiple ridges of the Frobisher Bay moraine
system occur. The chronology is based on the mapped
distribution of surficial deposits, evidence of relative
sea level variations, and radiocarbon dates from raised
marine deposits. Relative dating of moraine ridges was
attempted using lichenometry. A regional emergence curve
shows the present elevation of raised marine deposits
formed at any date after deglaciation and depicts restrained
rebound for sites deglaciated between 8450 and 7100 yr B.P.

Laurentide ice in upper Frobisher Bay retreated
slowly up-bay between Lewis Bay and Peterhead Inlet at an
average rate of 18 m yr^{-1}. Five main phases of moraine
formation are recognized and dated at 8450, 7800, 7510,
7340, and >7080 yr B.P. The ice margin retreated beyond
the head of the bay shortly after 7080 yr B.P. Two later
periods of shoreline formation within the Burton Bay area
may represent additional stillstands of the inland ice.

The marine limit declines continuously across the
Frobisher Bay moraine system from a high of 119 m beyond
the outer moraine ridge to 29 m at the head of Frobisher Bay, reflecting the slow ice retreat. High emergence rates (10.5 cm yr$^{-1}$) occurred between 8450 and 7800 yr B.P., during retreat from Lewis Bay to Burton Bay. Sites farther up-bay were deglaciated when lower emergence rates (3.1 cm yr$^{-1}$) prevailed. An attempt to draw emergence curves for unique sites using Andrews' (1970) general equation for the Canadian Arctic was unsuccessful because of the large amount of emergence before 7800 yr B.P. Moraines of different ages could not be differentiated by lichenometry because the largest lichens do not represent the age of moraine formation.

This local study of deglaciation from within the Frobisher Bay moraine system adds new details to the general model of late Foxe deglaciation for Frobisher Bay and southeastern Baffin Island presented by Blake (1966) and Miller (1980).
ACKNOWLEDGEMENTS

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My sincere thanks are extended to Dr. Jacobs for his guidance and patience, to Dr. Bill Mode of the University of Wisconsin for several useful suggestions, to both men for making some of their unpublished data available to me, and to Beth Brown and Kirsten Hedgecock for their field assistance and companionship while trudging through the tundra. Cartographic assistance from Mr. Ron Welch is also gratefully acknowledged. Lastly, I would like to thank those people whose friendship and encouragement during these years has meant more than they probably realize, and my parents, who would have been both proud and relieved to see this work completed.
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CHAPTER ONE

INTRODUCTION

1.1 Purpose and Scope

Baffin Island has been the focus of a considerable number of late Quaternary deglaciation studies during the last twenty years. Most of the research has centered on areas north of latitude 65°N, and as a result, the pattern of retreat of the last ice sheet is well established there. In contrast, little work has yet been published on the deglaciation of southern Baffin Island. In the Frobisher Bay area, studies of two major end moraine systems provide a framework for the chronology of deglaciation and relative sea level movements. Blake (1966) has examined the Frobisher Bay moraine system at the head of the bay as part of an extensive reconnaissance survey of southern Baffin Island, and Miller (1980) has studied the Hall moraine terminating in outer Frobisher Bay. More local fieldwork is needed, particularly in upper Frobisher Bay, to test and refine the general model of deglaciation established in those studies.

The purpose of this study is to develop a detailed sequence of deglaciation for the Burton Bay area, which lies within the southern part of the moraine system that
extends almost continuously from southwestern Hall Peninsula to the head of Cumberland Sound (Fig. 1). The system marks the last major stillstand or readvance of late Wisconsin ice before northwestward retreat and final disintegration beyond the head of Frobisher Bay (Blake, 1966). Although Blake (1966) has mapped the moraines at a scale of 1:2,000,000 and collected marine shells from the margins, the detailed series of events suggested by the pattern of multiple moraine ridges near Burton Bay has yet to be determined.

A chronology of deglaciation of the Burton Bay area is proposed here, based on the distribution of surficial deposits, evidence of relative sea level variations, and radiocarbon dates. The results from within the moraine system, together with limited evidence from the north shore of upper Frobisher Bay, are discussed in relation to the regional model of late Wisconsin ice retreat advanced by Blake (1966).

1.2 Previous Deglaciation Studies in Southeastern Baffin Island

Scores of papers have been written about various aspects of the deglaciation of Baffin Island. Several summary papers have synthesized the results that pertain to general problems of late Wisconsin ice extent and retreat throughout the island. Ives and Andrews (1963) proposed an initial, tentative model of deglaciation, and Prest (1970) later outlined the main sequence of events.
Fig. 1. Physiographic divisions in southern Baffin Island (after Blackadar, 1967 and Bostock, 1970) and distribution of the major moraine systems south of Cumberland Sound (after Blake, 1966 and Miller, 1980).
Miller and Dyke (1974) discussed the late Wisconsin limits of Laurentide ice on eastern Baffin Island, and Andrews and Ives (1978) reviewed the different interpretations of the near-continuous system of moraines along the east coast. Andrews (1980) reconstructed the form of the Laurentide ice sheet over Baffin Island 8000 to 9000 years ago and examined sea-level history throughout the last glaciation. In recent papers, the regional geologic-climatic term "Foxe Glaciation" has been used to refer to the last major glaciation on Baffin Island (Miller and others, 1977; Andrews and Ives, 1978; Andrews, 1980). Andrews (1980) has defined the Foxe Glaciation as the period between about 115,000 and 6000 yr B.P.

At the maximum of the late Foxe (late Wisconsin) glaciation, the northeastern margin of the Laurentide ice sheet lay over Baffin Island, spreading from a Foxe Basin center of dispersal (Ives and Andrews, 1963; Prest, 1970). The late Foxe advance, the Baffinland Stade (Andrews and Ives, 1978), was less extensive than earlier advances, and some areas on the east coast remained ice-free or were only locally glacierized (Miller and Dyke, 1974). Most of Hall and Cumberland peninsulas and probably the higher parts of Meta Incognita Peninsula were covered by local ice caps that remained dynamically distinct from the Laurentide ice sheet (Mercer, 1956; Miller and Dyke, 1974; Andrews, 1980).

The collapse of the Laurentide ice sheet may have begun 10,000 to 12,000 years ago (Andrews and Peltier,
1976), and by 8000 yr B.P. the sea had entered Hudson Bay via Hudson Strait (Blake, 1966). During the Cockburn Substage (8000 to 9000 yr B.P.), the period immediately before the disintegration of the Laurentide ice sheet (Andrews and Ives, 1978), a major system of end moraines was deposited along the northern ice margin. First identified on Baffin Island by Ives and Andrews (1963), the moraines of Cockburn age are traceable from Frobisher Bay to the northeastern and northern coasts of Baffin Island and into Melville Peninsula and central Keewatin (Falconer and others, 1965; Andrews and Ives, 1978). These moraines delimit the maximum late Foxe advance throughout much of Baffin Island (Miller and Dyke, 1974; Andrews and Ives, 1978), although some segments have been interpreted as recessional moraines marking a major stillstand or readvance phase during deglaciation (Blake, 1966; Hodgson and Haselton, 1974).

In southern Baffin Island, moraines of Cockburn age occur at the heads of Cumberland Sound and Frobisher Bay. The Ranger moraine at the head of Cumberland Sound probably marks the maximum extent of late Foxe Laurentide ice there, and retreat from that position began about 8670 yr B.P. (Dyke, 1979). The moraine crossing upper Frobisher Bay began forming during the Cockburn Substage, but the Hall moraine, which terminates 170 km farther down-bay (Fig. 1), represents the maximum late Foxe advance (Miller, 1978; 1980). The coastal section of the Hall moraine delimits the margin of an outlet glacier that reached outer Frobisher
Bay. Moraines trending across central Hall Peninsula 50 km beyond the parallel section of the Frobisher Bay moraine mark the limit of continental ice on the peninsula. The outlet glacier retreated from the Hall moraine by about 10,760 yr B.P., briefly readvanced about 10,000 yr B.P., then receded steadily to the Frobisher Bay moraine (Miller, 1980).

The multiple ridges of the Frobisher Bay moraine system mark a long period of ice stillstands or readvances before final northwestward retreat toward Amadjuak Lake. Radiocarbon dates indicate that the eastern, outermost ridge was forming at about 8230 yr B.P. and that the ice had receded from the inner moraine by at least 6750 yr B.P. (Blake, 1966). The major moraine on Foxe Peninsula has not been dated, although Blake (1966) provisionally correlated it with the Frobisher Bay moraine. In other studies, the ice margin during the Cockburn Substage is tentatively projected across eastern Hudson Strait from the Frobisher Bay moraine to the Sheppard moraine of northern Labrador (Falconer and others, 1965; Miller and Dyke, 1974; Andrews and Ives, 1978; Andrews, 1980).

The sea entered Foxe Basin by 7500 to 7000 yr B.P. (Prest, 1970), and as deglaciation progressed, the center of ice dispersal shifted onto Baffin Island (Ives and Andrews, 1963; Prest, 1970). An absence of Paleozoic limestone erratics from the Foxe Lowland on the granitic terrain east of Amadjuak Lake (Fig. 1) was seen by Blake
(1966) to indicate a possible center of ice dispersal for southern Baffin Island over Amadjuak Lake. Andrews and Miller (1979), noting the lack of limestone in till near upper Frobisher Bay, similarly concluded that an ice divide had lain inland from Foxe Basin or that a separate ice dispersal center had existed over the lake.

The western ice margin receded inland from Foxe Basin onto the coast near Nettilling and Amadjuak lakes by about 6700 to 6900 yr B.P. (Blake, 1966; Prest, 1970). The head of Frobisher Bay was also ice-free at this time. Retreat from Cumberland Sound occurred much later, and several recessional moraines formed behind the Ranger moraine (Dyke, 1979). Laurentide ice from central Baffin Island continued to calve into Cumberland Sound until 5700 yr B.P. The last remnant of Laurentide ice in the south, located east of Amadjuak Lake, may have persisted beyond 4500 yr B.P. (Blake, 1966).

No fieldwork has yet been reported from within the Frobisher Bay moraine system. The pattern of end moraines mapped from aerial photographs (Falconer and others, 1965; Blake, 1966; Miller, 1980) and controls provided by Blake's (1966) radiocarbon dates and marine limit observations from the inner and outer margins of the system form the basis of the regional deglaciation model for upper Frobisher Bay. This local study of the Burton Bay area is intended to refine the regional model.
1.3 The Study Area

Burton Bay lies on the southwest coast of Hall Peninsula, approximately 20 km southeast of the town of Frobisher Bay. Hall Peninsula is part of the Hall Upland physiographic province (Fig. 1), which consists of Precambrian crystalline bedrock (Blackadar, 1967). Much of the interior of the peninsula is a gently rolling upland, mostly greater than 600 m in elevation, crossed by several broad, structurally controlled river valleys (Bird, 1967). Along Frobisher Bay, elevations decline from more than 600 m near the mouth of the bay to less than 60 m at the head. Paleozoic carbonates and shales of the Foxe Lowland lie east of Amadjuak Lake and in a narrow zone northwest of upper Frobisher Bay (Blackadar, 1967). Paleozoic limestones crop out as small erosional remnants at the head of the bay and are preserved below sea level in the outer reaches of Frobisher Bay and Cumberland Sound (Miller and others, 1980).

In this study the Burton Bay area is defined as the region from the Sylvia Grinnell River to Lewis Bay, extending inland approximately 22 km from the head of Burton Bay to include the multiple ridges of the Frobisher Bay moraine system (Fig. 3). The Tatsiujarjualaq River (local name) flows across the Precambrian bedrock and Quaternary drift of interior Hall Peninsula into the moraine system, meandering through a broad, sediment-filled valley for its final 3 km before reaching the tidal flats of Burton Bay. Blackadar (1967) mapped the bedrock within
the study area as hypersthene granite and quartz-feldspar gneiss.

Except for small-scale mapping of the moraine ridges, the spatial distribution of the various surface features in the Burton Bay area has not been investigated. On the small-scale maps (Blake, 1966; Miller, 1980), the moraine ridges are drawn as mostly continuous features trending roughly parallel to Burton Bay. Seen on a larger scale, the pattern is more complex and the correlation of moraine ridges is not always clear. In some areas, the till appears as prominent moraine ridges; in others, the till is massive with distinct linear components. Many segments are separated by wide valleys that lie below the local marine limit. A closer study of the surficial deposits and former drainage patterns, together with recognizable marine limits and radiocarbon-dated raised marine deltas in the coastal areas of the moraine system, should permit the correlation of glacial and sea-level events within the study area and contribute to an understanding of the final retreat of Laurentide ice in upper Frobisher Bay.
CHAPTER TWO
THEORETICAL RELATIONSHIPS.

2.1 Introduction

In this study the chronology of deglaciation is developed primarily from the positions of surficial deposits, particularly end moraines, and dates on associated sea levels. Both the spatial distribution of end moraines and the spatial variations in marine limits reflect fluctuations of the ice margin. However, the link between glacial and sea-level events is not direct, but dependent on theoretical models of interrelationships between ice extent, glacio-isostatic depression, and global sea-level changes (Andrews and others, 1981). A model of relationships between surficial deposits and relative sea-level variations as they pertain to establishing a chronology of ice retreat is presented for upper Frobisher Bay.

2.2 Spatial Distribution of Surficial Deposits

The distribution of glacial deposits is influenced by such factors as the type of ice-marginal environment, ice temperature, dynamics at the base and snout, and the amount of debris available (Andrews, 1975). A detailed study of the form, structure, and composition of the
surficial deposits would provide evidence on conditions and dynamics at the ice margin, but these investigations are beyond the scope of this study. Discussions of glacial and glaciofluvial landforms and the processes of formation are available in the literature (e.g. Flint, 1974; Embleton and King, 1975; Sugden and John, 1976). The emphasis here is on the distribution of deposits.

Sugden and John (1976) present a model of the location of depositional landforms with respect to zones of deposition beneath an ice sheet. They are concerned with landform associations at a broad scale, and so do not indicate local variations within a zone of end moraines. Price (1973) depicts the spatial distribution of features that might develop in the marginal zone of a retreating ice sheet. This model is useful at a local scale, although the characteristic patterns at successive ice stillstands are not emphasized.

End moraines mark the position at which the margin of an actively flowing ice sheet lay for a significant period of time during a halt or readvance in deglaciation. They often represent a marginal thickening of a till sheet, which may form ground moraine up-ice (Flint, 1971). Gaps are frequently present in the moraine ridge, formed by contemporaneous meltwater streams, or marking sites where there was little debris for deposition or where the ice margin was locally stagnant (Flint, 1971). A long, continuous moraine ridge provides the best evidence of an ice-front
position.

Meltwater flowing at or beyond the ice margin results in ice-contact or proglacial features composed of stratified drift. In terms of a deglaciation chronology, marine deltas that develop in contact with the ice margin or from the discharge of a meltwater stream into the sea are important features, linking the ice-front position and sea level at the time of deposition. Glaciofluvial erosion and deposition continue during ice retreat from the end moraine, and sections of the moraine ridge will be altered or destroyed unless drainage is directed away.

As deglaciation proceeds, a similar pattern of glacial and glaciofluvial deposits will form at the ice margin during subsequent stillstands. If ice retreat continues with no significant readvance over older deposits, a landscape of moraines and associated glaciofluvial deposits will result. The moraine ridges will be successively younger from the outer margin toward the center of ice retreat. A glacial readvance beyond an older moraine ridge would likely destroy the morphological features of the previous stillstand. Evidence of the readvance could appear in the drift stratigraphy as till overlying stratified drift, but care must be taken in interpreting such evidence (Price, 1973). The resulting pattern of deposits after ice retreat would be similar to that described above, with younger moraines located farther from the outermost former ice margin. If the ice front readvanced over only part of
an older moraine ridge, younger ridges could extend beyond older ones. The relative ages of the ridges could be determined by cross-cutting relationships.

A series of closely spaced end moraines, which require a stillstand or readvance of an active ice margin for formation, would suggest that no regional ice stagnation, with cessation of flow from the accumulation area, had occurred. However, local ice stagnation is possible in the lee of large relief features when thin, active ice becomes unable to surmount a barrier and the marginal section is cut off (Price, 1973; Embleton and King, 1975). A complex series of mounds and hollows of ablation till could result after the ice finally disappeared by down-wasting. Decaying stagnant ice produces meltwater which can form ice-contact glaciofluvial deposits such as kames and kame terraces, and also proglacial features of stratified drift. Landforms developed by the wasting of stagnant ice would occur behind an end moraine where suitable topographic conditions existed. Although the deposits represent a broad ice-marginal position intermediate in time between the margins represented by moraines on either side, the deposits themselves could be younger, depending on when the ice melted away.

2.3 Variations in Marine Limits

During the deglaciation of coastal areas, the sea transgressed onto the glacio-isostatically depressed land. Although eustatic sea level was probably 20 to 30 m or
more below present sea level at 8000 to 9000 yr B.P. (Andrews, 1970a), within the Canadian Arctic periphery of the Laurentide ice sheet the initial rapid rate of post-glacial uplift exceeded the rate of eustatic sea level rise, so that shoreline features associated with the sea level at the time of deglaciation mark the maximum elevation reached by the sea (Andrews, 1966; 1970a). The marine limit, which is this maximum elevation measured with respect to present sea level, thus indicates relative sea level at deglaciation at a site that has been ice-covered.

Spatial variations in marine limit elevations reflect local fluctuations in ice recession (Andrews, 1970a; Miller, 1980). Instantaneous disappearance of an ice sheet would produce marine limits that incline upward toward the former ice center, where total glacio-isostatic depression was greatest. Rapid and continuous ice retreat would also produce this pattern. Slow retreat can result in marine limits that decline toward the center of an ice sheet. During slow retreat, relative sea level will fall because the ice-free areas are rebounding faster than eustatic sea level is rising. Ice-covered areas also experience uplift in the form of restrained rebound because the load of the thinning ice sheet is less than the load during the maximum thickness. The presence of ice prevents a shoreline from forming, however. By the time the ice retreats from a site and the marine limit forms at the
existing sea level, the marine limit at a nearby site that was deglaciated much earlier will already be elevated above sea level. An abrupt decline in marine limit elevations between the distal and proximal sides of an end moraine thus indicates a long stillstand of the ice front which enabled a substantial amount of rebound to occur (Blake, 1966; Andrews, 1970a; 1975; Miller, 1980).

Local variations in marine limit elevations could also be caused by a glacial readvance or by an isolated body of ice that remained in a coastal valley after the ice front retreated. In the first case, advancing ice could erase earlier marine limits. During deglaciation, a new sequence of marine limits would form at lower levels. Late-lying ice could cause anomalous low marine limits in the areas where deglaciation was delayed (Ives, 1964).

Whereas the marine limit is a metachronous feature, formed at the instant of deglaciation and thus becoming younger in the direction of ice retreat, a shoreline related to a water plane existing at any one time is, by definition, a synchronous feature. Elevations of this strandline may vary in response to tilting or warping caused by spatial variations in post-glacial rebound, but morphological features related to the particular relative sea level should be of similar age along the coast (Løken, 1962; Andrews, 1970a). In general, strandlines incline upward toward the center of the former ice mass, where the ice load was greatest.
2.4 Model of Relationships

In the Frobisher Bay area the ice retreated along the trend of the bay, so fluctuations in ice recession are depicted by variations in the elevation of marine limit features along the coast. Because the marine limit is synchronous with deglaciation (Andrews, 1966; 1970a), radiocarbon dates on shells found in raised features deposited at the marine limit can be used to date the deglaciation of the site. If the raised marine deposits can be associated with ice-marginal or proglacial streams that issued beyond a particular moraine, then the dated marine shells provide chronological control for the period of moraine formation and for the site of a well-defined ice front.

The spatial relationships between moraine ridges and marine limit features for upper Frobisher Bay are shown diagrammatically in the simple model of Figure 2. It is assumed that northwesterly ice retreat occurred over a long period of time with several halts or readvances, as proposed by Blake (1966). The model presumes that no readvance extended beyond an end moraine that had been deposited earlier.

The parallel moraine ridges, together with other associated glacial and glaciofluvial deposits that represent a distinct position of the former ice margin, decrease in age in the direction of ice retreat, toward the head of Frobisher Bay. Marine limit elevations
Fig. 2. Generalized relationships between end moraines and marine limit features in upper Frobisher Bay, assuming slow ice retreat toward the northwest.

decline toward the head of the bay, reflecting the slow retreat of the ice during the period of moraine formation. This general pattern is supported by Blake's (1966) observation of a decline of 90 m in the marine limits between the outer and inner margins of the moraine system.

In the Burton Bay area, meltwater streams would have drained toward Frobisher Bay, roughly parallel to the ice front. When the ice receded, meltwater occupied other
channels closer to the ice margin. Earlier drainage routes were abandoned or experienced a great reduction in discharge. As a result, meltwater erosion of older deposits in the proglacial area would have been slight. Another consequence, important for developing a chronology of deglaciation, is that marine limit features such as raised beaches and deltas can be uniquely associated with particular ice-front positions. Shells from a raised marine deposit directly linked to a moraine ridge and dating the marine limit provide a maximum (i.e. oldest) date for the beginning of moraine formation. Even if a marine limit feature cannot be directly related to a moraine ridge, it still indicates the minimum date for deglaciation of the site and hence for the retreat from the moraine located down-bay from it. The elevation of marine limit features can also be used to correlate moraine ridges on opposite sides of Frobisher Bay. Sites with similar elevations, which reflect similar sea levels at deposition, would have been deglaciated at the same time, assuming that the rates of post-glacial uplift were equal on both sides of the bay.

The age relationships of the moraine ridges can be determined independently of raised marine features by using various relative dating methods such as boulder weathering, lichenometry, and the degree of soil formation. These techniques must be able to detect variations within the 1000 to 1500 year span of moraine formation at approx-
imately 7000 to 8000 yr B.P. Relative dating methods provide a means to correlate discontinuous moraine ridges and thus to identify synchronous ice-front positions. Direct radiocarbon dating of ridges is rarely feasible because of the scarcity of suitable organic materials.

The actual pattern of ridges in the Frobisher Bay moraine system is more complicated than that shown in Figure 2 (Blake, 1966; Miller, 1980). Variations of the marine limit within the system are possibly more complex also. The model illustrates the relationships between moraines and sea level for a slow deglaciation marked with several halts or slight readvances. In the remainder of this study, the details of the distribution of moraine ridges and other surficial deposits within the Burton Bay area will be examined to evaluate this model of slow ice retreat.
CHAPTER THREE

METHODOLOGY AND RESULTS

3.1 Introduction

The chronology of deglaciation will be developed from an examination of the surficial deposits, radiocarbon dates, and variations in the marine limit. In this section, the procedures are described and results given for the mapping of the surficial deposits, relative dating of deposits, determination of the marine limits, and the construction of an emergence curve. The distribution of surficial deposits, shown by a morphogenetic map, is used to infer successive positions of the ice front and former locations of stagnant ice. Contemporaneous deposits can often be identified using the relative dating technique of lichenometry. Variations in the elevation of the marine limit illustrate the relative rate of ice retreat toward the head of Frobisher Bay, and radiocarbon-dated shells from marine limit features indicate dates of deglaciation. An emergence curve, showing the rate of post-glacial land emergence, is drawn using the elevations and dates of marine limits and lower sea level stands. In the discussion following this chapter, the results are synthesized to produce a chronology of deglaciation for the Burton Bay area.
3.2 Morphogenetic Mapping of Surficial Deposits

3.2.1 Method

The surficial deposits within the Burton Bay area are varied. They include not only moraine ridges, other till deposits, and glaciofluvial features, but also marine and glaciomarine deposits associated with a coastal location. Morphogenetic mapping from vertical aerial photographs, in the manner of Fulton and others (1975), provides a systematic means of classifying surface features within a large area. Terrain units are designated by genetic categories, which are subdivided in terms of morphology. Ground observations can be incorporated into the mapping scheme to identify the nature and morphology of deposits at a local scale and to verify the air photo interpretation. The resulting map depicts the continuous distribution of deposits rather than the discrete location of selected landforms such as deltas and moraine ridges.

The surficial deposits in the study area were mapped from 1:60,000 scale vertical aerial photographs using the landform classification system of the Canada Soil Survey Committee (1978), which is similar to the classification system developed by Fulton and others (1975) that has been applied in northeastern Baffin Island by Hodgson and Haselton (1974). Map units are depicted with alphabetic symbols that represent a genetic category, a morphological modifier, and where appropriate, a process modifier (Table
Areas too small to be delineated individually are mapped as mixed units, with the main component shown first and separated from the subordinate unit by a slash (/). Narrow moraine crests, beach ridges, and glaciofluvial ridges are indicated with linear symbols.

The mapped area extends northward from the shores of Frobisher Bay for 12 to 30 km inland, and from the Sylvia Grinnell River eastward to Lewis Bay (Fig. 3). Surficial deposits were mapped at the 1:60,000 scale and transferred to a 1:50,000 base map for reproduction.

Some areas were visited in the field to verify a preliminary map covering Laird Peninsula, the lower Tatsiujjarjualaq valley, and the central area farther north. Traverses were made across southeastern Laird Peninsula, throughout the valley, and to the central area near the Tatsiujjarjualaq River. The form and material of the deposits were noted and some sediment samples were collected. This information and field data from other areas (e.g. Pichit Peninsula and north of the settlement of Frobisher Bay) were incorporated into the final map of the larger area.

The greatest recurring uncertainties in the air photo mapping involved distinguishing fluvial and glacio- fluvial deposits that occur in former meltwater channels and identifying the dominant component where both moraine veneer (Mv) and bedrock (B) exist. If not associated with a nearby ice margin and thus designated glaciofluvial,
<table>
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<th>Process Modifier</th>
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<td>e eroded</td>
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<tr>
<td>C colluvial</td>
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<td></td>
</tr>
<tr>
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<td></td>
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</tr>
<tr>
<td>pG glaciofluvial</td>
<td></td>
<td></td>
</tr>
<tr>
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<td></td>
</tr>
<tr>
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<td></td>
</tr>
<tr>
<td>M morainal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>W marine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>wG glaciomarine</td>
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**Source:** Modified from Canada Soil Survey Committee (1969).
deposits close to present stream level were classified as fluvial and higher deposits in former meltwater channels were mapped as glaciofluvial. Field identification was useful for classifying B/Mv and Mv/B units, but since the boundary between the two units is gradational, it was difficult to be consistent in mapping beyond the traversed areas. Thick till areas (Mb), moraine ridges (Mr), and most marine deposits (e.g. Wpt, Wp) were easily recognized from photos, but knowledge of the marine limit helped in identifying marine deposits.

The distribution of surficial deposits as shown on Figure 5 is discussed below. Interpretation of selected features was aided by the results of a sediment analysis (Table 2). Particle size analysis was done by sieving and hydrometric methods. Sand, silt and clay percentages and Folk and Ward (1957) graphic measures have been calculated. Sediment sample sites are located on Figure 4.

3.2.2 Morainal Deposits

Morainal deposits cover a large part of the study area between the inner and outer ridges of the Frobisher Bay moraine system. Major till deposits occur as end moraines (Mr), moraine blanket (Mb), hummocky moraine (Mh), and moraine veneer (Mv). Several individual ridges are broad enough to be mapped as areal units at the scale of 1:50,000. Ridges too narrow to be mapped at this scale are shown with a linear symbol. In some areas, such as on Laird Peninsula, the Mr unit refers to a group of closely
<table>
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<tr>
<th>Site (Fig. 4)</th>
<th>Morphogenetic class\textsuperscript{a}</th>
<th>&gt; 2 mm (&lt; of bulk sample)</th>
<th>Sand (% of &lt; 2 mm fraction)</th>
<th>Silt (% of &lt; 2 mm fraction)</th>
<th>Clay (% of &lt; 2 mm fraction)</th>
<th>Folk and Ward Graphic Measures\textsuperscript{b}</th>
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\textsuperscript{a}As mapped on Figure 3.

\textsuperscript{b}Folk and Ward (1957): \( \bar{\phi} \) = mean grain size; \( \sigma \phi \) = standard deviation (sorting); \( \alpha \phi \) = skewness.
Fig. 4. Distribution of moraine ridges, location of sediment samples and features discussed in Sect. 3.4, and area covered by Fig. 21.
spaced till ridges.

In general, the moraine ridges within the mapped area are roughly parallel, trending north-northwesterly. In detail, correlation of the ridges is not straightforward because moraine ridges are rarely continuous for more than 5 km and departures from the general trend occur.

The discontinuous outermost moraine ridge appears to mark a synchronous ice margin. Along much of its length it forms a well-defined boundary between bedrock to the east and till deposits to the west. This narrow ridge occurs at elevations of 230 to 260 m in the north and gradually declines to an elevation of less than 100 m at the head of Lewis Bay, where it has been washed by a higher relative sea level. On Pichit Peninsula the moraine ridge is slightly wider. It occurs at elevations of 60 to 120 m and parts have been washed (Fig. 5).

The outer moraine ridge bifurcates about 5 km north of the head of Lewis Bay. The outermost section continues southward onto Pichit Peninsula, as described above, and the western branch trends toward northwestern Forter Inlet. The western ridge is not as well defined, consisting of short, narrow segments joined in places by zones of thick till. Small washed moraines dam the northwest and south-east ends of Easy Lake (unofficial name, Jacobs and Mode, personal communication), which lies in the lowland between Forter Inlet and an arm of Burton Bay ('a' on Fig. 4). At the head of Forter Inlet and in the lowland north of Lewis
Fig. 5. Distal (east) side of outer moraine ridge on Pichit Peninsula, lying about 90 to 100 m asl here.

Bay the moraine segments have been washed by the sea. Thick till deposits lie west of the outer moraine ridge in many areas, generally at elevations of 90 to 150 m in the south and above 150 m in the central region.

In the northern part of the mapped area, south of the large lake there (b), moraine ridges up to 500 m wide indicate several distinct ice-front positions. Three nearly parallel ridges are apparent, as well as another ridge that arcs around the southern margin of the lake. The broad ridges are 25 to 40 m high and their bases generally lie at 200 to 260 m. The innermost ridge is higher, lying at 300 m in the north. Cross-cutting
relationships observed on the air photos suggest that after the two eastern ridges (c and d) were deposited a small lobe advanced from the north, destroying part of the central ridge and forming the ridge that arcs around the lake (e). This ridge continues northward beyond the map area (partly shown on Fig. 4), but southern extensions have been overrun by a later readvance that formed the innermost parallel ridge (f). This western, inner ridge cuts across the curved moraine and assumes a similar curved form as it continues north of the lake. The only segment of these moraines examined in the field was the southern part of the eastern ridge (site 1, Fig. 4). The surface of the broad, flat crest here is littered with large boulders and gravel patches. The till consists of a silty sand matrix with more than 30 percent of the total sample of pebble size (site 1, Table 2).

Zones of thick till cover (Mb) and hummocky moraine (Mb) occur in the northern area between the Tatsiujarjualaq River and the western, innermost ridge. In the hummocky moraine areas the deposits form rounded hills 20 to 30 m high. Several hills occur as circular or near-circular rings up to 500 m in diameter with small central depressions 50 to 100 m across. These deposits are best developed on the distal side of the broad moraine ridges (g) and are mostly, but not exclusively, on land above 240 m. The zones of hummocky topography suggest stagnant ice deposits
that formed from the melting of ice beneath a cover of supraglacial debris.

Farther south, beyond the lake that forms part of the Tatsiujarjualaq River system (h), the pattern of moraine ridges is less clear. The surficial deposits are primarily moraine blanket (Mb), hummocky moraine (Mh), and moraine veneer (Mv), with segments of moraine ridges. Two short ridges immediately south of the lake mark the continuation of the two larger ridges to the north. The hummocky stagnant ice deposits (i) include several well-developed ring-shaped features and minor ridges. Except for two parallel moraine ridges farther south (j), the minor ridges might be interpreted as crevásse fillings or other ridges associated with stagnant ice. The irregular shape of the large lakes in this area also suggests the late existence of isolated blocks of ice.

The western, innermost moraine ridge continues south-westerly with few breaks to terminate on the upland east of the Tatsiujarjualaq River valley (k). The broad northern segment of the ridge descends from 300 m near the large lake (b) to 220 m at the first break. Segments south of this gap are narrower and lie mostly at 240 to 260 m. The section of the ridge examined in the field (site 2, Fig. 4) is about 60 m wide at the base and up to 10 m wide at the crest, though some segments are narrower and steep-sided (Fig. 6). Large, partly buried boulders are scattered on the surface. The ridge is composed of coarse sandy till
(site 2, Table 2).

A broad arcuate ridge extends across part of the Tatsiujarjualaq River valley between the bedrock upland and an outcrop in the valley (1), apparently as a continuation of the inner moraine ridge (see Fig. 21 and Fig. 22 for deposits in the valley). The flat, boulder-covered crest of this ridge lies about 7 m above the valley floor, at an elevation of 38 m aht. The ridge is composed of medium sandy till (site 3, Table 2). A narrow, sharp-crested ridge sits upon the flat ridge, with its crest at 43 m aht. The material of the higher ridge is coarse, moderately well sorted sand (site 4, Table 2). A 40-m aht

Fig. 6. Proximal side of narrow segment of moraine ridge near site 2 of Fig. 4.
washing limit was observed on the south side of the higher ridge. Large rounded boulders are scattered on the north side but are rare on the south side, especially below the washing limit. Andrews (1963) has described similar superimposed till ridges from Labrador. He interpreted the lower ridges as subaqueous deposits formed in glacially dammed lakes and the higher ridges as subaerial deposits. The complex till feature in the Tatsiujarjualaq River valley fits this interpretation. The broad, washed lower ridge was deposited subaqueously in a lacustrine environment while the coarse upper ridge remained above water during deposition at the ice margin. At the eastern end of this feature, a narrow moraine ridge extends up and across the bedrock outcrop, reaching an elevation of approximately 100 m and terminating above the valley floor.

Three short, narrow moraine segments (m) lie west of and parallel to the near-continuous inner ridge, interrupted by gaps more than 4 km long that are occupied by a continuous or discontinuous cover of moraine veneer. The ridges may mark a synchronous ice-front position 500 to 750 m west of the larger ridge. Alternatively, the widely separated moraine segments could represent minor local fluctuations during the recession from the main ridge.

A zone of multiple moraine ridges and a zone of thick, hummocky till occur on the higher land of southeastern Laird Peninsula. Several ice-front positions are delimited
by the moraines, perhaps corresponding to the wide multiple ridges in the northern map area. The northern traces of the Laird Peninsula moraines are lost in the sediment-filled valley north of Burton Bay. The ridges occur at about 150 m near Burton Bay, reach a maximum elevation of 240 m, and decline to 60 m toward the southern edge of the peninsula. Till sampled from the outermost moraine on Laird Peninsula (site 5, Table 2) consisted primarily of silty sand with a large proportion of coarse material. Thick till covers the high land on southeastern Laird Peninsula, mostly at elevations above 120 m. An isolated moraine ridge farther north (n) lies at an elevation of 120 to 165 m. No other major till deposits are found on the peninsula.

Extensive till deposits are found in the northwest section of the mapped area near the Sylvial Grinnell River. These moraine veneer and moraine blanket deposits continue northward along the river, but no moraine ridges are found there. The next identifiable moraine ridges are just south of Amadjuak Lake (Fig. 7).

3.2.3 Glaciofluvial and Fluvial Deposits

Many of the modern rivers occupy the same channels the major meltwater streams occupied, and it is difficult to differentiate fluvial deposits from glaciofluvial deposits along those channels. As a general rule, terraces in former meltwater channels are designated as
glaciofluvial if they are of ice-contact origin or if they are related to a higher water level associated with greater discharge during deglaciation. Lower deposits, closer to the level of the modern streams, are considered to be fluvial. Raised features that were deposited below the local marine limit are classified as marine.

Glaciofluvial deposits within the study area occur mostly in the north, between the Tatsiujarjualaq and Sylvia Grinnell rivers. High-level terraces flank the upper parts of these two rivers, and smaller deposits are found in minor meltwater channels north of Koojesse Inlet. Two small outwash plains, about 0.6 km², occur among the broad moraine ridges and thick till cover (o). In the northwest, the glaciofluvial deposits are mostly mixed units which include small eroded terraces, several eskers, and a thin, discontinuous cover of glaciofluvial material overlying bedrock.

Glaciofluvial deposits are found along the main channel that carried meltwater beyond the outer moraine ridge toward Lewis Bay (p). In the middle section of the Tatsiujarjualaq River, a large complex of glaciofluvial deposits occurs in an area where meltwater channels leading beyond the moraine ridge on the highland to the west meet several parallel channels from the northwest (q). Glacio-fluvial deposits that can be identified on 1:60,000 scale air photos are rare on Laird Peninsula. One small outwash terrace identified in the southeast consisted of very poorly sorted gravel and coarse sand (site 6, Table 2).
Minor fluvial deposits are found in the bedrock channels east of the Frobisher Bay moraine system. Large plains and terraces are associated with the major rivers such as the Sylvia Grinnell and sections of the upper Tatsiujaqjaq. Active fluvial deposition occurs along the incised meanders of the lower Tatsiujaqjaq River, but the features are too small to map at the scale used.

3.2.4 Glaciolacustrine and Lacustrine Deposits

Deltas are classified as glaciolacustrine features if they were deposited in contact with glacier ice in standing fresh water or by meltwater streams in ice-dammed lakes. A large glaciolacustrine delta complex lies along the Tatsiujaqjaq River between two closely spaced moraine ridges (near h). The surface of the higher northern section of the delta lies at 180 to 200 m, and the lower section lies below 160 m. The lower delta was formed in a lake created along the Tatsiujaqjaq River when the ice margin lay at the western moraine ridge. The higher deposits may be glaciofluvial, since there is no other depositional or erosional evidence that the lake level was greater than 160 m. Another glaciolacustrine delta (r) marks a temporary lake formed when the ice margin blocked the channel leading from the stagnant ice area south of the Tatsiujaqjaq River to the large marine delta in Burton Bay.

Terraces along the lower Tatsiujaqjaq River (s)
are considered to be remnants of a delta formed in an ice-dammed lake when ice advanced into the valley from the west (Fig. 7). The southernmost terrace reaches 53 m aht, whereas the apparent washing limit farther south in the valley is 40 m. A sediment sample from the deposit consisted of very coarse sand with a large proportion of gravel (site 7, Table 2). Foreset beds dipping in a down-valley direction were visible in an exposure farther north. The formation of this feature is further discussed in Chapter 4.

Fig. 7. Eroded remnants of the glaciolacustrine delta in the lower Tatsiujarjualaq valley (site 4, Fig. 4), looking northward from the 53-m aht surface.
3.2.5 Marine and Glaciomarine Deposits

Marine deposits cover large areas inland from the present coastline. The modern shoreline configuration is much different from that of immediate post-glacial times when many of the peninsulas would have been islands. Post-glacial emergence has exposed marine deposits at the heads of most major inlets. Other areas have been washed by a higher relative sea level but show no signs of deposition. Marine deposits were mapped with the knowledge of local marine limits, and all waterlain deposits found below the marine limit were classified as marine.

Marine deposits extend into the lowlands of Lewis Bay and Porter Inlet, and moraine ridges in these areas have been washed by the sea. At the head of Lewis Bay, a large raised delta lies at 42 m aht. A smaller delta 3.5 km upstream at an elevation of 96 m is also mapped as marine.

The marine incursion covered a large area in the lowlands north and east of Burton Bay. A raised delta occurs at the mouth of a former meltwater channel at the head of the bay (t). The main surface of the delta is at 35 m aht, but a higher section reaches 43 m. Marine deposits fill the lower Tatsiujarjualaq valley and the connecting valley at the head of Tarr Inlet. The sandy marine deposits in the Tatsiujarjualaq valley have been cut by gullies and the incised meanders of the river to form terraces that grade from 28 m aht to 15 m near Burton
Bay (sites 8-11). Higher terraces that lie north of the meandering section reach 38 and 39 m aht. They are mapped as marine features because they lie below a 40-m washing limit. Two near-surface sediment samples taken at 32 and 37 m aht from these high terraces consisted primarily of coarse to very coarse sand with cobbles up to 12 cm long (sites 12 and 13). Another sample from 29 m showed alternating bands of medium sand and coarse sand containing gravel (site 14). The high proportion of coarse material, up to 51 percent from the deposit on the west bank of the river (Fig. 8), suggests the ice margin was nearby during deposition. The central depression of the western feature likely resulted when a block of buried ice melted.

The small raised feature at the head of Tarr Inlet is mapped as a glaciomarine terrace. It appears to be a delta-moraine, formed at a stationary ice margin that terminated in the sea. A steep ice-contact slope faces Tarr Inlet, and the sea probably occupied the valley to the east during formation. The deposit reaches a maximum elevation of 27 m aht, which is less than the marine limit elevations to the east and west. Sediment analysis (site 15, Table 2) indicates that the feature is similar in composition to the subaerial part of the broad moraine ridge in the Tatsiujarjualaq River valley (site 4), but the Folk and Ward parameters are more like those of the subaqueous part of that deposit (site 3).

A marine terrace, eroded in places by small streams,
Fig. 8. Glaciomarine deposit on the west bank of the lower Tatsiujjarjualaq River. Top of the deposit reaches 38 m a.s.l. The lower marine terrace lies at 23 m.

lies along the north shore of Tarr Inlet, and a 24-m-high delta occurs farther west near Apex Hill (u). Thick marine sediments underlie most of the settlement of Frobisher Bay at the head of Koojesse Inlet, and a thin veneer covers part of the bedrock peninsula to the west. The only area of prominent raised beaches in the study area is north of Peterhead Inlet, at the western margin of the mapped area.

3.2.6 Colluvial Deposits

Unconsolidated materials accumulated on slopes are classified as colluvial deposits. They are hard to
distinguish from till, and in most cases the deposits are derived from till. Continuous colluvial or till deposits on steep slopes are classified as Cv. Colluvium and till combinations on gentler slopes are labelled as Cv/Mv or Mv/Cv, and discontinuous colluvial or till deposits on steep bedrock slopes are classified as Cv/B or B/Cv. Most colluvial terrain units are mixed units that occur on channel slopes or along the slopes of bedrock uplands.

3.2.7 Bedrock

The outer Frobisher Bay moraine ridge forms a distinct boundary between bedrock to the east and drift to the west. Bedrock is exposed on most of the peninsulas within the study area and often appears with a discontinuous cover of thin till. In a broad zone west of the innermost moraine ridge, moraine veneer overlies bedrock which strongly controls the topography.

3.3 Relative Dating by Lichenometry

3.3.1 Introduction

The moraine ridges in the study area are discontinuous and the pattern is complex. Contemporaneous ice-front positions cannot be definitively identified solely on the basis of the distribution of surficial deposits. Although datable organic materials have been found in former marine areas, they do not occur in appropriate locations to enable a correlation of coastal and inland moraine segments. Relative dating techniques were there-
fore considered as a potential means of correlating deposits.

In the Burton Bay region, the relative dating methods must be able to differentiate moraines deposited within a 1500 year interval approximately 6700 to 8200 years ago. Few methods have this capability. Techniques based on various boulder weathering characteristics or the degree of soil development are best used for deposits of glacial events separated further in time (e.g. Dugdale, 1972; Carrara and Andrews, 1972; Dyke, 1979). Lichenometry is frequently used for dating late Wisconsin and Neoglacial deposits in high altitudes and latitudes, but the technique is most successful for deposits formed near the present time. The Frobisher Bay moraines may be at or beyond the upper age limit for using lichenometry, but this procedure was considered to have the best potential for correlating moraine segments.

Lichenometry is a technique for estimating the relative or absolute age of a substrate using the lichens growing on the rocks. Beschel (1961) and Andrews and Webber (1964) have discussed the assumptions underlying lichenometry, and Jochimsen (1973) has pointed out some problems in the application to glacial studies. The basic assumption for dating glacial deposits are that no lichens were present immediately before deglaciation and that the growth of lichens found in similar environments is a function of time. The largest lichens on a substrate
should represent the minimum age of exposure of the sub-
strate.

*Rhizocarpon geographicum*, a slow-growing crustose
lichen found in arctic and alpine environments, is the
species most commonly used for dating old substrates
( Locke and others, 1979). Since the taxonomy of *Rhizocarpon*
is complex and field identification is difficult, the
various green-and-black species and subspecies are often
grouped together as *R. geographicum* for dating purposes.
These long-living lichens may be useful for dating deposits
up to 9000 years old in the Arctic (Miller and Andrews,
1972), although they are most successful for dating deposits
of Neoglacial age or younger (Webber and Andrews, 1973).

3.3.2 Method

*R. geographicum* were measured on boulders on five
moraine segments in the Burton Bay area (Fig. 9). Three
sections of the western moraines were sampled: north of Tarr
Inlet (A), in the valley northwest of Burton Bay (B), and
on Laird Peninsula (C). An intermediate ridge farther
inland was reached (D), and the outer moraine was sampled
on Pichit Peninsula (E).

Sampling methods for relative dating by lichenometry
vary, but any of several procedures applied consistently
within a restricted area should produce useful data (Locke
and others, 1979). Common sampling techniques include
measuring the diameter of the largest lichen on a moraine,
averaging the largest lichens from several sites, and
Fig. 9. Location of lichen measurement sites and radiocarbon-dated samples in the Burton Bay area.
determining size/frequency distributions. For this study, lichens were examined within several sites of 10-m radius spaced 40 to 60 m apart along a transect on the proximal side of a moraine ridge. Sampling was restricted to well-drained sites and similar rock types to minimize possible variations in lichen size due to factors other than time. All rock surfaces within a site were searched and the smallest diameters of large, roughly circular *Rhizocarpon geographicum* thalli were measured (Fig. 10). The five largest lichens for each site were recorded (see Appendix, Table 8).

The largest lichen at each of several sites on the moraine segment was used to calculate a mean lichen diameter

![Image](https://example.com/image.png)

**Fig. 10.** *Rhizocarpon geographicum* thallus (center) measuring 78 mm in diameter found on moraine ridge on Laird Peninsula (site C, Fig. 9).
to represent the moraine. This procedure is based on the assumption that a substrate that has been exposed longer will have lichens of greater average size. Mean values underestimate the true age of the moraines, since only the largest lichens can be as old as the exposed substrate (Beschel, 1961), but they are adequate for relative dating.

3.3.3 Results

The lichen data are summarized in Table 3. The mean values of maximum lichen diameters range from 57.0 mm for the middle ridge (D) to 82.0 mm for the outer ridge on Pichit Peninsula (E). Values for inner segments range from 66.2 mm to 69.4 mm. The largest lichens found within sample sites or nearby are also listed for each moraine.

**TABLE 3**

**SUMMARY DATA FOR LICHEN MEASUREMENTS**

<table>
<thead>
<tr>
<th>Moraine segment</th>
<th>Number of sample sites</th>
<th>Mean (mm)</th>
<th>Standard deviation (mm)</th>
<th>Largest lichen (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>7</td>
<td>69.4</td>
<td>3.3</td>
<td>85&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>B</td>
<td>6</td>
<td>66.8</td>
<td>3.8</td>
<td>70</td>
</tr>
<tr>
<td>C</td>
<td>5</td>
<td>66.2</td>
<td>3.8</td>
<td>78&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>D</td>
<td>6</td>
<td>57.0</td>
<td>4.0</td>
<td>60</td>
</tr>
<tr>
<td>E</td>
<td>7</td>
<td>82.0</td>
<td>9.5</td>
<td>92</td>
</tr>
</tbody>
</table>

<sup>a</sup>Found beyond a sample site
Analysis of variance and t-tests show significant
differences in mean lichen size between the three groups of
moraines at the 0.05 significance level (see Appendix). If
mean lichen values represent the ages of the moraines, these
results indicate that the outer moraine on Pichit Peninsula
is older than the others. Moraines north of Tarr Inlet, in
the valley northwest of Burton Bay, and on Laird Peninsula
are of similar ages, according to these results, and may
represent a single ice-front position. Also, the middle ridge
is younger than the others, suggesting residual ice remained
in the area after the ice-front retreated farther west.

The largest lichens were much smaller than might be
expected for deposits of this age, however, and it is
likely that the lichens do not represent the date of moraine
formation. Miller and Andrews (1972) have constructed a
growth curve for *R. geographicum*, plotting maximum lichen
diameter versus age, for northern Cumberland Peninsula,
Baffin Island. According to this curve, the largest lichen
measured on any moraine in the Burton Bay area (92 mm) is
less than 3000 years old. One larger lichen observed on the
raised shoreline at Porter Inlet, measuring 120 mm in
diameter, would be 4000 ± 800 years old. Lichens old
enough to date deglaciation more than 6750 years ago would
be greater than 200 mm in diameter. No lichens that large
were seen in the study area, although maximum lichen
diameters of 185 to 280 mm have been reported from a
number of studies on Cumberland Peninsula (Dugdale, 1972;

3.3.4 Discussion

One explanation for relatively small lichens on an apparently old substrate is that the largest lichens were overlooked by the quadrat method of sampling. It seems unlikely that no lichens between 92 and 200 mm would be found in any of the sample sites, though. Another explanation might be that the Cumberland Peninsula lichen curve is not applicable in the Burton Bay area. If the lichen growth rate was considerably slower in the study area, smaller lichens could represent the same relative date as larger lichens on Cumberland Peninsula. But growing conditions are more favorable in the Burton Bay area, where the growing season is longer and precipitation during the growing season is greater. If growth rates differed, they would probably be higher here, and hence larger lichens would be expected at any date.

If the present largest lichens did not colonize the moraines until about 3000 years ago, then they do not represent the date of moraine formation. Radiocarbon dates on shells indicate that the ice front had retreated beyond the study area by about 6750 yr B.P. and that the final ice mass in the area east of Amadjuak Lake may have disappeared after about 4500 yr B.P. (Blake, 1966). Neither residual ice masses, if they could survive until 3000 yr B.P., nor permanent snow banks could have existed at all moraines throughout the study area because of differences
in elevation, exposure, and proximity to the sea. Late exposure of the moraines is probably not the reason for the small lichens.

Maximum lichen sizes might date the stabilization of the substrates, perhaps after an ice core melted out from the moraines. Another possibility is that the young ages indicate that climatic conditions were unfavorable for lichen growth before 3000 yr B.P. and that older lichens, if they had existed, had been killed. In either case, the lichen data would not be valid for determining the relative ages of the moraine ridges.

Some studies have found that *R. geographicum* thalli reach an effective upper limit at a relatively small size. Andrews and Webber (1964) noted that lichens seldom exceeded 120 mm in diameter and most maximum values were less than 80 mm at the northwest margin of the Barnes Ice Cap in north-central Baffin Island. They attributed the small sizes to weathering of the rock surfaces. Løken and Andrews (1966) found that 70 to 80 mm was the upper limit of usefulness at the south margin of the Barnes Ice Cap because larger thalli tended to coalesce. In the Burton Bay area, though, most rocks were only slightly altered by weathering, and coalescence was not a major problem.

Although a pattern of variation in mean lichen diameters was shown for the moraine ridges in the study area, the lichens probably do not date the formation of the ridges. The glacial deposits are older than the upper
limit of usefulness for lichenometry here, which is about 3000 yr B.P. if the growth curve of Miller and Andrews (1972) is valid in this area. In the absence of other suitable dating methods, inland moraine segments can only be correlated by their position and relationship to surrounding surficial deposits. Moraines near the coast can be dated through their association with marine limit features, and radiocarbon dates from these marine limit features form the basis of the chronology of deglaciation.

3.4 Marine Limits and Radiocarbon Dates

3.4.1 Introduction

The marine limit marks the highest level reached by the sea on a glacio-isostatically depressed coast (Andrews, 1975). It can be delimited by the highest glaciomarine delta built by glacial meltwater and sediments, by the washing limit, which is the lowest level at which perched boulders or undisturbed ground moraine are found, by the uppermost beach ridge, and by the highest occurrence of marine fossils. In the study area, marine limits were determined from the highest marine deltas and from washing limits. Elevations were measured with a hand level or an altimeter as the height above the known limit of the previous high tide or as the height above the upper level of shore-rafted debris, which approximates the large high tide. All elevations were converted to the height above mean high tide using Canadian Tide and Current Tables.
The various features which indicate the marine limit in an area may represent different levels. Whereas raised marine deltas are constructional features that probably formed near low tide level (Andrews, 1970a, p. 77), washing limits are erosional features that mark the former high tide level. Delta elevations in the study area are thus minimum values for the marine limit, and they may underestimate the actual value by as much as the tidal range. Although the tidal range may have been different in the past, delta elevations can be made roughly comparable with washing limit elevations by adding to them the present mean tidal range of 7.8 m for upper Frobisher Bay.

Radiocarbon dates and associated marine limits are listed in Table 4. Marine limit elevations determined from raised deltas are minimum values; no corrections for tidal range have been made in the table. The dates do not necessarily date the marine limit listed for the site, but may refer to a lower relative sea level. Details of the dates and elevations are described below, proceeding from the distal side of the Frobisher Bay moraine system to the head of Frobisher Bay. Dated sites located within the study area are shown on Figure 9.

3.4.2 Discussion of Dates

Radiocarbon dates from marine shells in deposits of the Hall moraine provide evidence of early retreat of the outlet glacier in lower Frobisher Bay (Miller, 1980). Ice
<table>
<thead>
<tr>
<th>Lab. no.</th>
<th>Date (yr B.P.)</th>
<th>Location</th>
<th>Elevation (maht)</th>
<th>Source a</th>
</tr>
</thead>
<tbody>
<tr>
<td>QC-480c</td>
<td>10,760 ± 150</td>
<td>Warwick Sound</td>
<td>61</td>
<td>(1)</td>
</tr>
<tr>
<td>GSC-2725</td>
<td>10,100 ± 110</td>
<td>Gold Cove</td>
<td>73</td>
<td>(1)</td>
</tr>
<tr>
<td>GX-8159</td>
<td>8450 ± 190 b</td>
<td>Lewis Bay</td>
<td>37-39</td>
<td>(2)</td>
</tr>
<tr>
<td>GSC-462</td>
<td>8230 ± 240</td>
<td>Cape Rammelsberg</td>
<td>88</td>
<td>(3)</td>
</tr>
<tr>
<td>GX-8696</td>
<td>8225 ± 450 b</td>
<td>Easy Lake</td>
<td>1</td>
<td>(5)</td>
</tr>
<tr>
<td>QC-905</td>
<td>7800 ± 150</td>
<td>Burton Bay</td>
<td>5</td>
<td>(2)</td>
</tr>
<tr>
<td>QC-902</td>
<td>7510 ± 320</td>
<td>Burton Bay</td>
<td>34</td>
<td>(2)</td>
</tr>
<tr>
<td>GSC-2771</td>
<td>7380 ± 220</td>
<td>Burton Bay</td>
<td>14</td>
<td>(2)</td>
</tr>
<tr>
<td>QC-901</td>
<td>7340 ± 135</td>
<td>Tatsiujarjuslaq R.</td>
<td>13</td>
<td>(2)</td>
</tr>
<tr>
<td>GX-8160</td>
<td>7080 ± 175 b</td>
<td>Peterhead Inlet</td>
<td>16</td>
<td>(2)</td>
</tr>
<tr>
<td>GSC-464</td>
<td>6750 ± 170</td>
<td>Apex Hill</td>
<td>14-17</td>
<td>(3)</td>
</tr>
<tr>
<td>GSC-533</td>
<td>6440 ± 160</td>
<td>Probisher Bay</td>
<td>3</td>
<td>(3)</td>
</tr>
<tr>
<td>GX-8695</td>
<td>6430 ± 225 b</td>
<td>Easy Lake</td>
<td>2</td>
<td>(5)</td>
</tr>
<tr>
<td>GSC-503</td>
<td>6140 ± 170</td>
<td>Apex Hill</td>
<td>15</td>
<td>(3)</td>
</tr>
<tr>
<td>GSC-849</td>
<td>4140 ± 130</td>
<td>Sylvia Grinnell R.</td>
<td>15</td>
<td>(4)</td>
</tr>
</tbody>
</table>

a1) Miller, 1980  
b) Quoted error term of 1 sigma; other dates are given with 2 sigma error.  
2) Andrews, 1983  
3) Blake, 1966  
4) Maxwell, 1973  
5) Jacobs and Mode, unpublished data  
Representing former low-tide level; other marine limits represent high tide level.
withdrew from the Hall moraine near Warwick Sound, approximately 160 km down-bay from the outer Frobisher Bay moraine, at about 10,760 ± 150 yr B.P. (QC-480c). Deglaciation from Gold Cove, 20 km farther up-bay (Fig. 3, inset), occurred at 10,100 ± 110 yr B.P. (GSC-2725). Between the terminus of the Hall moraine and the outer Frobisher Bay moraine, the marine limit rises steadily from 73 m to 119 m, implying continuous ice recession to the Frobisher Bay moraine (Miller, 1980).

On the southwest side of Frobisher Bay near Cape Rammelsberg (Fig. 3, inset), Blake (1966) found marine shells which were dated at 8230 ± 240 yr B.P. (GSC-462). The shells lay at 88 m aht in silts associated with the outer moraine of this section of the Frobisher Bay system. The marine limit at the site is about 103 m aht (Lowden and others, 1967, p. 182). Washing limits 12 km southeast of this moraine and 19 km southeast of the outer moraine on Pichit Peninsula are close to 120 m (Blake, 1966). The date represents a minimum date for deglaciation and hence for the marine limit on the distal side of the moraine.

At the head of Lewis Bay, whole valves and fragments of marine shells were found at 37 to 39 m aht in bottomset beds beneath the surface of a 42-m-high delta (Andrews, 1983, p. 25). The delta and the streams leading to the delta lie immediately east of the outermost moraine. The shells, dated at 8450 ± 190 yr B.P. (GX-8159), are similar in age to shells from the southwest side of Frobisher Bay,
but the surface of the delta is well below the 103-m washing limit there (Fig. 11). However, the Lewis Bay shells are from bottomset beds, and sea level at the time of deposition could have been considerably higher. Nearly 3.5 km farther upstream from this delta, remnants of a smaller delta lie at 96 m (Fig. 12). No marine shells have been found here, but the local washing limit is 119 m aht (Jacobs and Mode, 1982, pers. comm.). The shells from the lower delta thus provide a minimum date for the 119-m marine limit.

At the head of Porter Inlet the western arm of the outer moraine ridge dams the southeastern end of Easy Lake. The moraine crest is 21 m high and has been washed by the sea (Fig. 13). Shorelines cut the seaward side at 19 m and the proximal side at 18 m. A small raised delta lies at 14 m aht on the seaward side. The northwestern margin of Easy Lake is blocked by a 9-m-high washed, bevelled moraine ridge. A washing level near the northwestern end of the lake is at 28 m, but this elevation is probably not the marine limit since it is lower than the marine limits in nearby Lewis Bay and Burton Bay. On the western, proximal side of the moraine on Pichit Peninsula, the washing limit is 90 m (Fig. 14).

A core from Easy Lake, believed to represent a complete post-glacial section, yielded two radiocarbon dates that relate to the marine incursion (Jacobs and Mode, unpublished data). The 1.5-m core spans a depth/
Fig. 11. East bank of the 42-m aht marine delta at the head of Lewis Bay, where shells dated 8450 ± 190 yr B.P. were found.

Fig. 12. High delta at 96 m aht (open arrow) and local washing limit of 109 m (closed arrow) in the channel 3.5 km north of the lower delta at Lewis Bay.
Fig. 13. Washed moraine ridge and 14-m marine delta lying between Easy Lake (off photo to left) and Porter Inlet (right).

Fig. 14. Prominent 90-m washing limit on the eastern flank of the moraine ridge on Pichit Peninsula, which also appears as the marine limit on the western flank of the ridge.
elevation range of 2.5 to 1.0 m aht. A date of 8225 ± 450 yr B.P. (GX-8696) was obtained from organic material in the lower 20 cm of the core. This date, determined from 20 cm of sediments, would include materials of different ages and therefore is a minimum date for deposition at the bottom of the core. Easy Lake was thus ice-free and occupied by marine waters by at least 8225 ± 450 yr B.P. The date here is similar to the date for the deglaciation of Lewis Bay, but the position of the two sites in relation to the outermost moraine ridge requires that the Lewis Bay site was ice-free first. The 90-m washing limit on the proximal side of this moraine must be older than the Easy Lake date of 8225 ± 450 yr B.P., but younger than the Lewis Bay marine limit, which formed at least 8450 ± 190 years ago.

The second Easy Lake date, from organic material in marine sediments 20 cm above the older sample, shows that the site was still open to the sea at 6430 ± 225 yr B.P. (GX-8695). Relative sea level was at least higher than the 9-m washed moraine that forms the lowest entrance to the lake at that time.

A large raised delta lies at the head of Burton Bay, east of the Tatsiujarjualaq River and its tidal flats, at the south end of a bedrock channel that extends from an area formerly occupied by stagnant ice. The main surface of the Burton Bay delta measures 35 m aht and a smaller section lies at 43 m. The higher value indicates a marine
limit of 43 m, or 51 m if adjusted for the modern tidal range.

Three radiocarbon dates are available from the delta (Andrews, 1983). The oldest date of 7800 ± 150 yr B.P. (QC-905) came from shells found at only 5 m aht. Another shell sample, dated at 7510 ± 320 yr B.P. (QC-902), occurred at 34 m aht. If the bivalve molluscs of the sample lived in abundance only at low tide level or below as they presently do in upper Probus Bay (Ellis and Wilce, 1961), and if the tidal range was similar to the present range of nearly 8 m, then these shells represent a minimum relative sea level of 42 m, only 9 m below the marine limit. The older shells found at 5 m aht were probably laid down in deeper water when sea level was at or near the marine limit. They indicate that the area was deglaciated by 7800 ± 150 yr B.P. Shells from 14 m aht at another site on the delta were dated at 7380 ± 220 yr B.P. (GSC-2771). From the above assumptions, these shells would date a relative sea level of 22 m or higher.

In the valley northwest of Burton Bay, shells from 13 m aht in the sandy bluffs along the Tatsiujarjualaq River have been dated at 7340 ± 135 yr B.P. (QC-901). The bluffs are part of the terraced marine plain that grades from 28 m north of the shell location to 15 m near Burton Bay. A 40-m washing limit in the lower valley is probably not the local marine limit, since marine deposits occur up to 39 m aht, but a 40-m washing limit on the proximal side of the
moraine ridge in the valley probably represents the marine limit after deglaciation from that part of the valley. The 40-m sea level in the valley should be younger than the 7800 ± 150 yr B.P. Burton Bay delta and may have formed near 7340 ± 135 yr B.P., when the adjusted relative sea level was 21 m or more.

Two dates have been published for the delta near Apex Hill, which lies west of Tarr Inlet and the inner ridge of the Frobisher Bay moraine system. The older date, 6750 ± 170 yr B.P. (GSC-464), was previously the oldest estimate for deglaciation from the inner moraine (Blake, 1966). Fragments of marine shells from the 14- to 17-m level were collected on the 21- to 24-m-high delta for the dated sample. The washing limit here is close to 30 m. Blake (1966) considered it possible that the sea had penetrated the inner moraine earlier than 6750 years ago because the shells were much lower than the marine limit, but after the tidal range correction is added to the shell elevation, the minimum high tide level of 25 m is only 5 m below the high tide level marked by the washing limit.

Highly weathered shell fragments found at 15 m aht in the Apex Hill delta have been dated at 6140 ± 170 yr B.P. (GSC-503; Blake, 1966; Matthews, 1967). The elevation is similar to that of the 6750 ± 170 yr B.P. sample (GSC-464), but higher than a 3-m aht site near Frobisher Bay airport dated at 6440 ± 160 yr B.P. (GSC-533; Blake, 1966).
One explanation for the difference in dates is that the shells found at 3 m aht are from deep-water sediments associated with a higher sea level, and the weathered, young shells from Apex Hill are contaminated and actually represent deposits similar in age to sample GSC-464 (Matthews, 1967). Matthews has identified a poorly developed terrace at 15 m aht at Apex Hill, and an alternative explanation for the dates is that the weathered $6140 \pm 170$ yr B.P. shells date the 15-m shoreline cut into the older deltaic sediments.

Materials from an archaeological site near the Sylvia Grinnell River, west of the town of Frobisher Bay, were dated at $4140 \pm 130$ yr B.P. (GSC-849; Maxwell, 1973). The site was at 15 m aht, indicating that sea level has been below this elevation for at least 4000 years.

Shells found underlying a 16-m aht terrace in eastern Peterhead Inlet were dated at $7080 \pm 175$ yr B.P. (GX-8160), but it is unclear whether the terrace was formed by erosion or deposition (Andrews, 1983, pp. 25-26). If the 16-m terrace is erosional, it may be a younger feature cut into a deposit that formed at or near the time of the marine limit. A higher terrace is found at 24 m aht, and the local marine limit is 29 m. If the 16-m terrace is a depositional feature, it indicates a relative sea level for low tide near 16 m about 7080 years ago and an earlier marine limit of 29 m. These possibilities are considered later during the discussion of an emergence curve. The
date is now the oldest evidence for deglaciation from the inner side of the Frobisher Bay moraine system.

3.4.3 Marine Limit Trends and Ice Retreat

Spatial variations in marine limit elevations are shown on Figure 15, where data from the study area have been projected onto the axis of Frobisher Bay. Observations from upper Lewis Bay and Burton Bay have been projected onto the axis of Frobisher Bay from the mouths of their respective inlets to preserve the relative position of marine limit observations and moraine ridges. Distances are measured from the head of Frobisher Bay.

The data from the study area corroborate the observations of Blake (1966) and Miller (1980) that the marine limit declines across the Frobisher Bay moraine system toward the head of the bay. From a high of 119 m in Lewis Bay, beyond the outermost ridge, the marine limit declines to Burton Bay at an average gradient of 8 m km$^{-1}$. From Burton Bay to the proximal side of the inner moraine, the gradient is 1-2 m km$^{-1}$. Beyond Apex Hill, the variation in marine limit elevations is slight.

The decline in marine limits at the Frobisher Bay moraine is a reversal of the trend between the Hall moraine and Lewis Bay, in which marine limits rise up-bay from 75 m to 119 m (Miller, 1980). The large drop across the Frobisher Bay moraine system reflects a long period of slow ice retreat, marked by stillstands at the various moraine
Fig. 15. Marine limits in upper Frobisher Bay. Moraine ridges are identified by phases shown on Fig. 20.
ridges, in which a total of 90 m of restrained rebound took place. Radiocarbon dates from the proximal and distal sides of the moraine system put a limit of about 1370 years on this period. The average rate of ice retreat between Lewis Bay and Apex Hill was thus about 18 m yr\(^{-1}\), much slower than the earlier retreat from the Hall moraine to the outer Frobisher Bay moraine which averaged 85 m yr\(^{-1}\) (Miller, 1978. The consequences of slow ice retreat and restrained rebound for the emergence history of upper Frobisher Bay are discussed in the next section.

3.5 Emergence Curves for Upper Frobisher Bay

3.5.1 Introduction

Radiocarbon-dated materials related to the marine limit and to lower sea levels provide information on the nature of post-glacial emergence of the land. An emergence curve is produced by plotting and joining time/elevation points for sea level. The curve can be used to estimate the age of undated marine features and to fix relative sea level at any given date.

An emergence curve is ideally drawn from many samples that date definite sea levels at a single site. Within a larger area, strandlines incline upward toward the former area of greatest ice thickness, so that marine features of the same age may now occur at different elevations in different locations. Separate emergence curves from several sites would depict the regional pattern of emergence
more accurately than a single curve.

In the upper Frobisher Bay area, there is no site with enough dated samples related to specific sea levels to draw a well-controlled emergence curve. However, for sites where the marine limit and date of deglaciation are known, curves can be approximated using Andrews' (1970a) general equation for the Canadian Arctic. This method would allow curves to be drawn for several sites in the study area. Another alternative is to use all available dates and elevations to draw a single regional curve for upper Frobisher Bay. Both alternatives are presented here.

3.5.2 Predicted Emergence Curves

Andrews (1970a) found that post-glacial uplift curves for the Canadian Arctic had a similar exponential form which can be approximated by the equation:

\[ U_p(t) = A(1.521^t - 1) / 0.521 \]  \hspace{1cm} (1)

where \( U_p(t) \) is predicted uplift at time \( t \), \( t \) is expressed in \( 10^3 \) yr B.P., and the coefficient \( A \) is calculated by multiplying the total uplift (which equals the marine limit plus a correction for eustatic sea level at the date of deglaciation) by a variable related to the date of deglaciation (Andrews, 1970a, Fig. 3.8). Post-glacial uplift, which is the glacio-isostatic recovery of the crust following the removal of the ice load, consists of the observable emergence and the rebound masked by the general rise in world sea level that resulted from the melting of the late
Wisconsin ice sheets. Emergence curves are derived from predicted uplift curves by subtracting a correction for eustatic sea level change at each date. Andrews (1970a) used estimates from Shepard (1963) for the eustatic correction in the uplift equation and for converting uplift curves to emergence curves. Although there is no general agreement on the appropriate eustatic sea level corrections (Andrews, 1970a, p. 23), the use of Shepard's values with a form of Andrews' equation has provided good estimates to field data in several areas of the Arctic (Andrews and Falconer, 1969; Andrews and others, 1970).

A major limitation in predicting emergence curves from the empirical equation and only one point is that the smooth exponential curve cannot depict relative sea level fluctuations that might result from glacial readvances or shifting centers of uplift. Neither can local variations that might exist because of complex geological structure or possibly the gravitational attraction of the ice mass on the sea (Clark, 1976) be represented. As already noted, the available radiocarbon dates are too few to define an emergence curve, therefore they are also unable to show sea level fluctuations.

Predicted emergence curves are drawn for three sites in upper Frobisher Bay where estimates of the age and elevation of the marine limit are available. The sites occupy distinct positions with respect to the Frobisher Bay moraine system: Lewis Bay is beyond the outer moraine
ridge, the Burton Bay delta lies within the moraine system, and Peterhead Inlet is on the proximal side of the inner ridge. Three uplift curves were calculated using equation (1) and the data in Table 5, and values for the eustatic sea level correction (Andrews, 1970a, Table II-4) were subtracted to give the emergence values graphed in Figures 16-18.

In addition to the marine limit date, the emergence curve for each site shows dates of other nearby samples to help evaluate the agreement between predicted and observed emergence values. In this paper, no corrections have been made to the radiocarbon dates for the variation of radiocarbon years from calendar years, the correct half-life of $^{14}$C, or other factors, in the absence of a general agreement on the corrections to be applied. The use of uncorrected dates makes the emergence curves more comparable to other curves from Arctic Canada. Dates are reported in $^{14}$C yr B.P. with the standard error assigned by the dating laboratory.

On the emergence curves, dates associated with a narrow sea level range are plotted as rectangles, with the width indicating the laboratory standard error for the date and the height representing the possible range of sea level. Samples that date a minimum or maximum sea level are plotted to show the dating error and the relative position of sea level. The width of the emergence curve reflects the standard error of the date of the marine limit,
TABLE 5
DATA TO CALCULATE UPLIFT CURVES FOR SITES IN UPPER FROBISHER BAY

<table>
<thead>
<tr>
<th>Date of deglaciation</th>
<th>Lewis Bay 8450±190 BP</th>
<th>Burton Bay 7800±150 BP</th>
<th>Peterhead Inlet 7080±175 BP</th>
</tr>
</thead>
<tbody>
<tr>
<td>Marine limit</td>
<td>119 m</td>
<td>51 m</td>
<td>29 m</td>
</tr>
<tr>
<td>Eustatic correction</td>
<td>19 m</td>
<td>15 m</td>
<td>10 m</td>
</tr>
<tr>
<td>Percentage</td>
<td>1.55</td>
<td>2.06</td>
<td>2.80</td>
</tr>
<tr>
<td>A-value(^a)</td>
<td>2.14</td>
<td>1.36</td>
<td>1.09</td>
</tr>
</tbody>
</table>

\(^a\) A = (marine limit + eustatic correction) \times percentage

but this is only one of the several sources of errors involved in constructing emergence curves. Other possible errors, discussed by Andrews (1970a), include identifying the marine limit, relating shells and other organic materials to their contemporary sea level, measuring the elevation of the marine limit, and correcting for eustatic sea level changes, as well as problems inherent in the radiocarbon dating method. Since the combined effects of these factors is unknown, it is impossible to put stricter error limits on the emergence curves (Andrews, 1968b).

Marine limit dates for each curve and the elevations assigned to other dated samples are reviewed below.

At Lewis Bay, the emergence curve is based on shells from 39 m aht which provide a minimum age of 8450 ± 190 yr B.P. (GX-8159) for the 119-m marine limit (Fig. 16). Other
dates shown with this curve are from Easy Lake and Cape Rammelsberg. Organic material in the core from Easy Lake places sea level at less than 90 m at 8225 \( \pm \) 450 yr B.P. (GX-8696). The Cape Rammelsberg site is on the opposite shore of Frobisher Bay, but the date gives some control on the position of sea level during the formation of the outer Frobisher Bay moraine. At Cape Rammelsberg, sea level was between 88 and 103 m at 8230 \( \pm \) 240 yr B.P. (GSC-462).

In Burton Bay, relative sea level was between 42 m and the 51-m marine limit at 7510 \( \pm \) 320 yr B.P. (QC-902) and closer to the marine limit at 7800 \( \pm \) 150 yr B.P. (QC-905). The latter date is used as the estimate of the 51-m marine limit in constructing the emergence curve (Fig. 17). The sample from 14 m in the Burton Bay delta dates a relative sea level of at least 22 m at 7380 \( \pm \) 220 yr B.P. (GSC-2771). The shells at 13 m along the Tatsiu- jarjualaq River indicate sea level was greater than 21 m at 7340 \( \pm \) 130 yr B.P. (QC-901).

On the proximal side of the moraine system, the Peterhead Inlet date is used to draw the emergence curve (Fig. 18). The oldest sample from Apex Hill places sea level between 25 m and the 30-m marine limit at 6750 \( \pm \) 170 yr B.P. (GSC-464). The Peterhead Inlet date of 7080 \( \pm \) 175 yr B.P. (GX-8160) refers to a similar range of 24 to 29 m. Given the narrow range of sea level for the two dates, it is likely that the Peterhead Inlet sample closely dates the 29-m marine limit and the Apex Hill date refers
Fig. 16. Predicted emergence curve for Lewis Bay.
Fig. 17. Predicted emergence curve for Burton Bay.
Fig. 18. Predicted emergence curve for Peterhead Inlet.
to a level nearer to 25 m. This conclusion implies that the 16-m terrace in which the shells at Peterhead Inlet were found is erosional, cut into a deposit formed near the time of the marine limit. The younger shell sample from Apex Hill dates a sea level of at least 15 m at 6140 ± 170 yr B.P. (GSC-503), and the date from the Sylvia Grinnell River shows that relative sea level was less than 15 m at 4140 ± 130 yr B.P. (GSC-849).

The three predicted emergence curves are unique and there are no marked discrepancies with the few dated observations. Little dating control is available to evaluate the part of the curves between 6000 yr B.P. and the present, though. Emergence rates are greatest in the period immediately following deglaciation and decrease toward the present, as demanded by the general exponential equation. The separate curves suggest that emergence proceeded at different rates at the three sites. Initial emergence was most rapid in Lewis Bay, with about 70 m of emergence occurring during the first 2000 years after deglaciation. In Burton Bay, where total emergence was only 51 m, emergence during the first 2000 years was about 30 m. Peterhead Inlet experienced the slowest emergence rates after deglaciation, with only 17 m of emergence during the initial 2000 years.

The large variation in emergence rates is predictable, given the 90-m drop in marine limits between Lewis Bay and Peterhead Inlet during the relatively short period of less
than 1400 years. The relative position of the three curves is unexpected, however. According to the emergence curves, a strandline formed at any specified date in upper Frobisher Bay would tilt upward in a down-bay direction. For example, marine features formed close to high-tide level at 6000 yr B.P. would now be found at 17 m in Peterhead Inlet, 23 m in Burton Bay, and 40 m in Lewis Bay. The theory of isostasy dictates that these strandlines should incline upward roughly toward the northwest, in the direction of the center of isostatic depression associated with the Foxe Basin or Amadjuak Lake ice domes. This is the case farther down-bay, where Miller (1978) found that terraces dated between 9200 and 9700 yr B.P. indicated upward tilting in a westerly to northwesterly direction. Assuming that strandlines in upper Frobisher Bay must also tilt upward toward the presumed ice center northwest of the bay, then either the radiocarbon dates and marine limit elevations used to calculate the emergence curves have been misinterpreted, or Andrews' general equation does not correctly predict emergence in this area.

The rationale for assigning radiocarbon dates to the specific marine limit elevations has already been discussed. A few additional points in reference to the emergence curves can be made. If the sample from Peterhead Inlet dates the 16-m terrace instead of the 29-m marine limit, then the marine limit would be older than 7080 yr B.P. An older date would shift the emergence curve farther to the right,
increasing the discrepancy between the Peterhead Inlet and Lewis Bay curves. In Burton Bay, the date is probably a close estimate for the formation of the marine limit. If the 8-m addition to convert the delta elevation to a high-tide water level is removed, the marine limit would become 43 m. The lower value would move the Burton Bay curve to the right also, closer to the Peterhead Inlet curve. At Lewis Bay, the shells from 39 m are a minimum date for the 119-m marine limit. A marine limit date of 10,000 yr B.P. or earlier would be needed to shift the calculated emergence curve for Lewis Bay around the marine limit date of Peterhead Inlet so that the strandline formed at 7080 yr B.P. tilts up toward Peterhead Inlet. At this time, however, according to Miller's (1980) evidence, the outlet glacier in Frobisher Bay was at the Hall moraine, nearly 140 km beyond Lewis Bay. Thus, despite some uncertainties in the marine limit dates, the data cannot be interpreted to make the three predicted emergence curves correspond to the expected regional pattern of differential uplift. It seems, therefore, that equation (1) does not provide an accurate representation of emergence in upper Frobisher Bay. An alternative approach which may provide a more realistic picture of emergence in the region is to draw a single curve using all of the available dates.

3.5.3 Regional Emergence Curve

A single emergence curve was subjectively drawn to accommodate eleven radiocarbon dates from upper Frobisher
Bay (Fig. 19). The width of the curve represents a range of approximately \( \pm 175 \) radiocarbon years. The interpretation of dates and elevations is the same for the regional curve as for the predicted emergence curves. Possible sources of error are also similar, with the exception of eustatic sea-level estimates. No addition of eustatic sea level was required in fitting the regional curve directly to the dated sea-level observations.

The position of the emergence curve is uncertain between 6500 yr B.P. and the present because dates are lacking. Dyke (1979) found that Cumberland Sound is currently undergoing submergence which began less than 1000 years ago at the head of the sound and perhaps more than 3000 years ago beyond the Ranger moraine. Miller and others (1980) noted that recent submergence has occurred in outer Frobisher Bay, but found no unambiguous evidence of submergence at the head of the bay. The present rate of uplift in east Baffin Island has been estimated at 0.2 to 0.3 cm yr\(^{-1}\) (Andrews, 1968a). The regional emergence curve proposed here for upper Frobisher Bay is drawn to show continuous post-glacial emergence.

The single curve is applicable throughout the study area only if strandlines are not tilted between Lewis Bay and Peterhead Inlet. The available dates are insufficient to trace a particular strandline here, however. The date best represented is about 7200 yr B.P. At Peterhead Inlet, marine deposits from about 7100 yr B.P. date a 29-m aht
Fig. 19. Regional emergence curve for upper Frobisher Bay.
level. Deposits of the same age at Apex Hill would be between 25 and 30 m high. In the Tatsijuńjarjuaq River valley and at the Burton Bay delta, the shoreline of about 7350 yr B.P. is now at least 21 m but less than 42 m aht. Within the limits implied by such sparse information, the single curve of Figure 19 may be viewed as a tentative model of regional emergence.

3.5.4 Discussion

The regional emergence curve is extremely steep in its upper portion, with almost 100 m of emergence occurring during the first 2000 years after the deglaciation of Lewis Bay. Only 20 m of emergence took place during the last 6500 years. If three separate curves had been drawn to allow for regional tilting and to fit the marine limit dates, the curve for Lewis Bay would be steeper yet. The rapid rate of emergence is one of the highest in the eastern Arctic. Emergence during the first 1000 years following the deglaciation of Lewis Bay was at least 80 m, about twice as much as in Cumberland Sound and west Baffin Island, and three to four times as much as in east Baffin Island (Table 6). The proportion of total emergence accomplished during this interval is also markedly greater in upper Frobisher Bay. As a result of the anomalous high rate of initial emergence, Andrews' (1970a) general equation cannot adequately predict emergence here. Emergence may still be approximated by an exponential curve, but the parameters would be different from those averaged from the 21 sites.
<table>
<thead>
<tr>
<th>Location</th>
<th>Source</th>
<th>Emergence&lt;sup&gt;a&lt;/sup&gt;</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Total (m)</td>
<td>Within 1000 yr after deglaciation (m)</td>
<td>% during first 1000 yr</td>
<td></td>
</tr>
<tr>
<td><strong>East</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inugsuin Fiord</td>
<td>Løken (1965)</td>
<td>65</td>
<td>23</td>
<td>35.4</td>
<td></td>
</tr>
<tr>
<td>Kangok Fiord</td>
<td>Andrews and others (1970)</td>
<td>79</td>
<td>29</td>
<td>36.7</td>
<td></td>
</tr>
<tr>
<td><strong>West</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Isortoq Fiord</td>
<td>Andrews (1966)</td>
<td>95</td>
<td>33</td>
<td>34.7</td>
<td></td>
</tr>
<tr>
<td>Piling Lake area</td>
<td>Andrews (1970b)</td>
<td>102</td>
<td>38</td>
<td>37.3</td>
<td></td>
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<tr>
<td><strong>Southeast</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cumberland Sound</td>
<td>Dyke (1979)</td>
<td>88</td>
<td>40</td>
<td>45.2</td>
<td></td>
</tr>
<tr>
<td>Pribishier Bay</td>
<td>--</td>
<td>119</td>
<td>80</td>
<td>67.2</td>
<td></td>
</tr>
</tbody>
</table>

<sup>a</sup>Estimated from emergence or uplift curves
originally used to develop the equation (Andrews, 1968b).

High emergence rates at Lewis Bay could be related to the geological structure. Frobisher Bay has undergone Tertiary faulting (Miller and others, 1980), and local post-glacial tectonic activity may have occurred. Another possible cause of rapid emergence, proposed by Clark (1976) to explain anomalous high uplift rates in Greenland, is the gravitational attraction of an ice sheet on the sea. According to Clark, as the mass of the Greenland ice sheet decreased and the ice margin there retreated, the reduced gravitational attraction of the ice would have caused a rapid fall in relative sea level. On Baffin Island, the large drop in sea level at Lewis Bay may similarly reflect the decrease in the mass of the Foxe Basin ice dome after the Cockburn Substage, before the sea entered Foxe Basin (Prest, 1970).

In addition to providing estimates of emergence after deglaciation, the regional emergence curve also shows restrained rebound for Burton Bay and Peterhead Inlet between 8450 yr B.P. and the date of deglaciation. Glacio-isostatic rebound in upper Frobisher Bay began at least 10,760 years ago, when the outlet glacier started to thin and recede from its maximum late Foxe position at the Hall moraine, but only the last phase of restrained rebound occurring after 8450 yr B.P. can be measured in the study area. Marine limits on the proximal and distal sides of the Frobisher Bay moraine system indicate that 90 m of
restrained rebound took place beneath the ice at Peterhead Inlet while the ice margin retreated from Lewis Bay. The emergence curve shows that most of this occurred before about 7500 to 7800 yr B.P., when the steep gradient of the curve begins to decline. The average rate of restrained rebound before 7800 yr B.P. was 10.5 cm yr⁻¹, similar to an estimate of 6-10 cm yr⁻¹ during the ice stillstand at the Cockburn-age Ekalugad moraine in east Baffin Island (Andrews, 1970a). Between 7800 and 7100 yr B.P., the rate of restrained rebound in upper Frobisher Bay was only 3.1 cm yr⁻¹. Sites deglaciated during this later period, including Burton Bay and Peterhead Inlet, have lower rates of initial emergence because much of the total glacio-isostatic recovery had been accomplished beneath the ice sheet. Slow ice retreat accounts for the later dates of deglaciation at these sites, and hence for the lower emergence rates.
CHAPTER FOUR

CHRONOLOGY OF DEGLACIATION

4.1 Ice Retreat in the Burton Bay Area

The first major stillstand or readvance of the ice sheet after retreat from the Hall moraine at 10,100 ± 110 yr B.P. is marked by the outer ridge of the Frobisher Bay moraine system (phase A, Fig. 20). During the stillstand, the ice margin lay east of the upper Tatsiujarjualaq River and extended south-southwesterly to Pichit Peninsula. Meltwater issuing from several sources along the ice front flowed into a channel that paralleled the ice margin and ended 5 km north of the present shore of Lewis Bay ('a' on Fig. 20). The 119-m marine limit and the 96-m-high delta at the mouth of the channel formed while the ice front was at this position. A minimum date for the stillstand is provided by marine shells from the large lower delta 3 km downstream, which indicate that Lewis Bay was ice-free at 8450 ± 190 yr B.P. (GX-8159). If relative sea level was near the marine limit of 119 m during formation of the outer ridge, much of the southern 7.5 km of the ridge was deposited close to or below sea level. Initial post-glacial emergence was rapid, however, possibly as much as 15 cm yr⁻¹ at Lewis Bay.
Fig. 20. Phases in the deglaciation of the Burton Bay area. (Note that the lettering scheme for ice-front positions is independent of the letter designations used in the lichen sampling in Sect. 3.3).
While the ice margin remained stable along most of the outer ridge, the marine-based part withdrew 1 to 2 km westward from Pichit Peninsula and the head of Lewis Bay to form the western arm of the outer ridge (phase A'). Several sections of the ridge were formed in contact with the sea, including the 21-m-high curved segment along the southeast shore of Easy Lake and a broad segment more than 2 km beyond the present head of Lewis Bay (b). The age of this stillstand can be estimated from two sources. Firstly, a radiocarbon date from Easy Lake indicates that the ice had retreated from the western arm of the outer ridge by 8225 ± 450 yr B.P. (GX-8696). Secondly, the stillstand may be synchronous with the prominent 90-m washing limit on the proximal side of the ridge on Pichit Peninsula, which can be dated by the regional emergence curve as between 8100 and 8450 yr B.P.

After a period of retreat the ice stood at position B, which is defined by a broad moraine ridge in the north but is uncertain in the south. The ice lay west of the upper Tatsiujarjualaq River in the north, blocking the river near its westward bend (near c) and depositing the moraine ridge. Hummocky till deposits associated with stagnant ice occur on the distal side of the moraine ridge parallel to the ice margin (d). The relationship of the moraine ridge and the stagnant ice deposits suggests that a narrow terminal zone of the ice sheet thinned and eventually stagnated while the ice behind remained active.
Debris thrust onto the stagnant ice from the active margin formed hummocky deposits with rounded hills, ring-shaped mounds, and small kettles when the underlying ice melted. The moraine ridge marks the stillstand position of the active ice. Abandoned meltwater channels cross the drift to the upper Tatsiujarjualaq River, where high-level glaciofluvial terraces were formed.

South of the bend in the Tatsiujarjualaq River, the trace of the ice margin is obscured by a zone of stagnant ice deposits (e). While glacier ice dammed the river, drainage was directed southward around or through this stagnant ice zone to a channel that leads to Burton Bay, east of the present Tatsiujarjualaq River. A moraine ridge forming at the ice margin blocked the channel at a point about 5 km north of the bay, and a large delta formed in the ice-dammed lake (f). From there, meltwater reached Burton Bay by way of a narrow channel cut deeply into the bedrock. At the mouth of the channel, the high section of the large marine delta was deposited at the local marine limit, when relative sea level was at least 43 m and possibly near 51 m. The marine limit at the Burton Bay delta is dated at about 7800 ± 150 yr B.P. (QC-905), and the stillstand marked by position B occurred at that time or shortly thereafter.

1Marine limits and other elevations that refer to a low-tide level have been converted to possible high-tide values by adding the present mean tidal range at Frobisher, which is about 8 m.
Since upper Burton Bay was ice-free during phase B, the southern continuation of the ice margin must have been on Laird Peninsula. It is not clear whether ice occupied the lower valley of the present Tatsiujarjualaq River during this stage or part of the valley was ice-free. The trend of the moraine ridges on Laird Peninsula and north of Burton Bay suggests that only Burton Bay would have been ice-free. The ice margin during phase B is drawn across upper Burton Bay on Figure 20, linking the outer moraine ridge on Laird Peninsula to the outer of the two parallel ridges north of Burton Bay, though the relative age of these two segments has not been determined.

A period of retreat followed phase B, and the ice stood 1 to 3 km to the west at position C. In the north the ice front is marked by a wide ridge that was overrun near the large lake (g) during a later ice advance. The ridge can be traced southward across the Tatsiujarjualaq River to the zone of stagnant ice deposits where the position of the ice margin is lost. The innermost of the two contiguous moraine ridges near the glaciolacustrine delta in the channel leading to Burton Bay (f) is considered to be synchronous with this phase.

Meltwater flowed southeasterly beside the northern segment of the moraine ridge to the impounded Tatsiujarjualaq River. A kettled glaciofluvial terrace was deposited between the ridges of phases B and C (h), and a large glaciolacustrine delta complex was deposited in the
ice-dammed river (c). Outflow from the lake probably escaped southward through the stagnant ice zone to the channel leading to Burton Bay. The lower part of this channel was not blocked by ice during phase C, and the small dammed lake that had been held in the channel during phase B was released. Meltwater cut through the glacio-lacustrine deposits that had formed here (f) and occupied a new channel 500 m to the west to flow toward the Burton Bay delta.

The position of the ice margin farther south is placed at the complex zone of multiple ridges on southeastern Laird Peninsula, nearly 1 km west of the position assigned to phase B. This correlation is based on the trend of moraine ridges and on an assumption that the ridges of phases B and C occur close together on Laird Peninsula, as they do in the northern and probably the central areas. The lower Tatsiujarjualaq valley may still have been covered by ice during phase C, as was the middle section of the river. Meltwater from inland reached the sea at the Burton Bay delta, and the extensive deltaic deposits at 35 m aht may have formed during this period. There are no raised marine deposits on southeastern Laird Peninsula that relate to either phase B or phase C.

The radiocarbon date from the lower Tatsiujarjualaq valley indicates that the valley was ice-free by at least 7340 ± 135 yr B.P. (C-901). If the Tatsiujarjualaq valley was still occupied by ice during phase C, the date
is a minimum date for this stillstand. A maximum date is provided by the date for phase B, which is 7800 ± 150 yr B.P. (QC-905). Furthermore, if the 35-m delta at Burton Bay formed during phase C, relative sea level may have been near 43 m. Shells from 34 m in the delta would then date this phase at about 7510 ± 320 yr B.P. (QC-902).

After retreating from position C, the ice readvanced to position D, which is represented only by a broad moraine ridge that arcs around the southern end of the large lake in the northwest (g) and the associated glaciofluvial deposits. Any southern extensions of the ridge have been overrun by the later readvance of phase E. During phase D, meltwater from the ice lobe in the north flowed eastward and entered the upper Tatsiujaqjualaq River, breaching the multiple ridge's that had formed during local fluctuations of the ice margin in phase B (i). Meltwater also flowed southward beyond the ridge, on both sides of the ridge formed in phase C, toward the present-day Tatsiujaqjualaq River. An outwash plain adjacent to the ridge (j) and two small glaciofluvial terraces shown on Figure 3 are associated with drainage beyond the lobate ice front.

Except for the lobe in the northwest, the ice front during phase D lay somewhere west of the innermost moraine ridge. After the ice retreated from position C, the ice-dammed lake along the Tatsiujaqjualaq River was lowered and the river occupied its present course in the upper and middle sections. The stagnant ice east of the river
(e) may still have persisted, with meltwater entering the channel to the Burton Bay delta or joining the river.

The lower valley of the modern Tatsiujarjualaq River was inundated by the sea during phase D, but sea level at this stage is unknown. If the regional trend of marine limits declined during successive phases of moraine formation because of ongoing uplift during each stillstand (see Sect. 2.3), the marine limit in the lower valley should be lower than the 51-m marine limit of phase B and the possible 43-m sea level of phase C and higher than the level of phase E. The 40-m washing limit observed in the valley appears to be related to a sea level after phase E and not the marine limit at deglaciation. Except for thick marine deposits on the valley floor, the surficial deposits in the valley are attributed to phase E.

No precise date can be assigned to phase D. The regional emergence curve dates a relative sea level of 40 to 50 m at 7350 to 8000 yr B.P. (Fig. 19). An upper limit of 7800 ± 150 yr B.P. is given by the marine limit date of phase B, and if the 35-m delta at Burton Bay formed in phase C, the upper limit may actually be 7510 ± 320 yr B.P. (QC-902). Since the lower valley is known to be ice-free at 7340 ± 135 yr B.P. (QC-901), this is the tentative date assigned to phase D, though it may be somewhat young.

In phase E, the ice front throughout most of the study area readvanced to a position at or just west of the divide separating drainage to Burton Bay via the Tatsiujar-
jualaq River from drainage into the upper part of Probisher Bay. The position of the ice margin is marked by the near-continuous inner ridge, which overruns part of the moraine loop of phase D in the northwest and extends south-southeasterly for about 19 km. Meltwater flowed off the divide into the middle Tatsiuajarjualaq valley, where extensive glaciofluvial terraces were deposited (k). Remnants of the stagnant ice to the east may have blocked the channel here for a period.

The moraine ridge across the lower Tatsiuajarjualaq valley (l) appears to be a continuation of the ice front in phase E. Northwest of the ridge, eroded remnants of a delta lie at least 53 m aht. The delta is too high to have formed during phase D, when the Tatsiuajarjualaq River entered the sea there. It is 10 m higher than the maximum elevation of the Burton Bay delta and higher than all other marine deposits west of Lewis Bay. Therefore, the delta is considered to be a lacustrine deposit that formed in a lake held up by glacier ice in the lower valley, at or near the position of the arcuate moraine ridge (Fig. 21 and Fig. 22).

The ridge across the valley is a complex feature that could not have formed wholly in a lake level of 53 m aht. The broad, flat-crested lower section that reaches 58 m aht was deposited below water, but the narrow upper ridge that reaches 43 m remained above water during deposition at the ice front. Either the ice lay slightly south of the ridge during the maximum lake level and
Fig. 21. Middle and lower Tatsiujarjualaq River (location shown on Fig. 4). Features related to phase E are sketched on Fig. 22. Other features shown here are: zone of stagnant ice deposits (A), channel to Burton Bay delta (B), glaciolacustrine delta of phase E (C), contiguous moraine ridges of phases B and C (D), marine terraces in lower valley (E), multiple ridges of phase C on Laird Peninsula (F). Scale 1:76,000 (approx.). Part of Government of Canada air photo A16166-85.
Fig. 22. Ice position in the lower Tatsiujarjualaq valley during phase E.
advanced to deposit the entire feature when water level had fallen to 38–40 m, or the lower ridge formed during the higher lake level and the upper ridge formed only after water level had fallen.

At the east end of the arcuate ridge, a moraine ridge continues onto the bedrock outcrop in the valley, reaching a height of about 100 m. The outlet of the lake to the sea was a narrow bedrock channel east of the outcrop, which the present Tatsiujarjualaq River occupies. The lower valley remained open during the deposition of the high terraces immediately south and southeast of the bedrock outcrop, so the ice margin is drawn to the isolated moraine ridge on Laird Peninsula. Beyond the ridge, the ice position on Laird Peninsula is conjectural, as there are no apparent ice-marginal deposits.

The high terraced deposits at 38 and 39 m aht in the lower valley formed in the sea during phase E, when ice partly blocked the valley and held up the lake to the north to at least 39 m (Fig. 22). They lie at the mouth of the outlet channel of the lake and close to the ice front, which was the source of the very coarse material found particularly in the deposit on the west bank. The 39-m marine terrace is indicative of a relative sea level of at least 39 m, and possibly up to 47 m, during phase E. The washing limit on the proximal side of the ridge in the valley, which formed later on retreat from phase E, is 40 m aht. The 40-m washing limit farther down-valley may also
correspond to the lower relative sea level during ice retreat.

Phase E cannot be closely dated. The ice had retreated to Peterhead Inlet by 7080 ± 150 yr B.P. (GX-8160), so the stillstand occurred somewhat earlier than this. Although the date of 7340 ± 135 yr B.P. (QC-901) is assigned to phase D on the basis of shells in the lower marine terraces, those shells could have been deposited during phase E.

Phases C, D and E all occurred between 7800 ± 150 yr B.P. and 7080 ± 150 yr B.P., but the specific dates assigned to the phases are tenuous (Table 7). Although several radiocarbon dates fall within that time period, the shells cannot be associated with a definite sea level (e.g. GSC-2771 and QC-901) or the deposit in which they occur cannot be related to a particular phase (e.g. QC-902). Sea level values for those phases are also problematic. If the rate of isostatic uplift remained greater than the rate of eustatic sea level rise during deglaciation, and relative sea level thus fell continuously throughout the period, sea level values for phases C, D and E should be in the range of 40 to 50 m, decreasing in each phase. With the 35-m level of the Burton Bay delta assigned to phase C and a minimum sea level of 40 m in phase E, sea level in phase D would then be narrowly limited to between 40 and 43 m. The closeness of values in these phases suggests the possibility of a sea level stillstand, likely associated
TABLE 7

DATES AND RELATIVE SEA LEVELS ASSOCIATED WITH PHASES OF DEGLACIATION IN THE BURTON BAY AREA

<table>
<thead>
<tr>
<th>Glacial Phase</th>
<th>Approx. Date (yr B.P.)</th>
<th>Relative Sea Level (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>≥8450 ± 190</td>
<td>119</td>
</tr>
<tr>
<td>A'</td>
<td>&gt;8225 ± 450</td>
<td>≤90</td>
</tr>
<tr>
<td>B</td>
<td>7800 ± 150</td>
<td>51</td>
</tr>
<tr>
<td>C</td>
<td>7510 ± 320</td>
<td>43 (?)</td>
</tr>
<tr>
<td>D</td>
<td>7340 ± 150</td>
<td>?</td>
</tr>
<tr>
<td>E</td>
<td>&gt;7080 ± 150</td>
<td>≥40</td>
</tr>
</tbody>
</table>

with the ice readvance from phase D to E, caused by a reduction in the rate of uplift or an increase in the rate of eustatic sea level rise or both. If relative sea level in phase E is actually closer to 47 m, a marine transgression is indicated. However, available dates and sea level elevations are too few to support this or to justify modification of the regional emergence curve (Fig. 19).

After phase E and the formation of the 40-m washing limit on the south side of the moraine ridge in the Tatsiu-jarjualaq valley, the ice retreated from the prominent ridges of the Frobisher Bay moraine system. During retreat, ice in Frobisher Bay appears to have held up a lake in Koojesse Inlet while the land to the north was ice-free. Dale (1982) found an indurated clay layer beneath the modern tidal flats of Koojesse Inlet which she interpreted as having been deposited in an ice-marginal lake. There is little evidence in the study area on which to base the
position of the ice margin at that time. Thick till deposits around the Sylvia Grinnell River in the northwest and extending beyond the mapped area may signify a pause in the general ice retreat. Small eskers and other glaciofluvial deposits are also found in this area. A tongue of ice in Tarr Inlet deposited the delta kame at 27 m aht and kept the sea from entering Koojesse Inlet by way of the inundated Tatsiujarjualaq valley and adjoining valley.

When the ice retreated to release the marginal lake in Koojesse Inlet, the 30-m marine limit was established. Similar marine limits of 29 to 30 m at Peterhead Inlet and the head of the bay indicate rapid withdrawal of the ice lobe out of upper Frobisher Bay into the lowland to the northwest at about 7080 ± 175 yr B.P., the date of deglaciation for eastern Peterhead Inlet. The marine delta at Apex Hill and the 24-m terrace along Tarr Inlet formed when the ice had retreated from that area and sea level was close to the marine limit. The oldest date for the delta is 6750 ± 170 yr B.P. from shells at 15 m (GSC-464). The 24-m terrace at Peterhead Inlet, which also formed close to the marine limit, was forming at 7080 ± 175 yr B.P. (GX-8160).

The Frobisher delta may have begun to form in the ice-dammed lake occupying Koojesse Inlet, since sediments at the northwest end (mapped as marine in Fig. 3) lie above the 30-m (100-ft) contour, the approximate elevation
of the marine limit. Meltwater reached the delta by way of the Sylvia Grinnell River, which entered Koojesse Inlet then, and also from another channel farther east. Koojesse Inlet would have been open to the sea at about 7080 ± 175 yr B.P., but the only radiocarbon date from the much-disturbed delta is relatively young, 6440 ± 160 yr B.P. (GSC-533), obtained from deep-water sediments at 3 m aht. Other undated shells have been found up to 12 m (Matthews, 1967).

During ice retreat from the Tatsiujaŋjuaŋjaq River to Peterhead Inlet the marine limit fell from 40 m to 29 m, as indicated by washing limits. The delta kame in Tarr Inlet at 27 m aht represents an intermediate sea level. The fall in relative sea level resulted in the initial recession of the sea occupying the lower Tatsiujaŋjuaŋjaq valley. Deposits on the valley floor south of the arcuate ridge which lie at 30 to 32 m aht would have emerged above high-tide level at about 7080 yr B.P. Marine deposits that form the terraces along the modern river would have been exposed at low tide in much of the valley. As the marine recession proceeded, a meandering river was established on the marine sediments, and the meanders became incised as the isostatic uplift of the land continued. The extensive marine deposits in the lower valley were thus deposited in the relatively short period between ice retreat from phase C at perhaps 7500 yr B.P. and emergence above high tide, estimated from the regional
emergence curve at about 6700 yr B.P. in the middle section of the lower valley. An average rate of sedimentation can be calculated at one site where shells occur in sediments at 13 m and the upper marine deposits lie at 23 m. The shells were deposited at 7340 ± 135 yr B.P., and if the upper sediments were above sea level at 6700 yr B.P., the average rate of deposition was about 1.6 cm yr⁻¹.

Little can be said about the period after 7080 ± 175 yr B.P., when the ice margin had retreated beyond the study area. There is some evidence for a period of shorel ine formation throughout the area that would reflect a temporary balance in the rates of uplift and eustatic sea level rise at about 6140 ± 225 yr B.P. The 16-m erosional terrace at Peterhead Inlet, the poorly developed strandline at 15 m at Apex Hill (Matthews, 1967), shorelines cut into a moraine ridge at 18 and 19 m at the head of Porter Inlet, and the 14-m delta at Porter Inlet may be synchronous features. The date is provided by the youngest, highly weathered shells at 15 m in the Apex delta (GSC-503). These features may be contemporaneous with a 20-m strandline reported by Mercer (1956) at several sites beyond Cape Rammelsberg on the opposite shore of Frobisher Bay. However, if shorelines tilt upward toward the head of Frobisher Bay, the 20-m strandline may instead relate to the 24-m-high deltas and terraces at the head of the bay which were forming about 6750 to 7080 yr B.P.
4.2 Regional Comparisons

The five main phases of ice stillstand or readvance within the Frobisher Bay moraine system occurred between about 8450 and 7080 yr B.P. The dates assigned to each phase are provisional (Table 7), particularly phases C, D and E, because of problems in identifying the ice-front position throughout the area and the lack of dated marine shells that can be related to the sea level at deglaciation. In addition, few ice positions or sea levels can be dated more closely than \( \pm 200 \) years because of the error term associated with the radiocarbon date. Dates from some phases overlap, though the distinct moraine ridges in the north clearly indicate five separate periods of stillstand or readvance. Correlations of specific glacial and sea level events with those of other areas are therefore necessarily general.

The outer moraine ridge, which marks the first major ice stillstand after retreat from the Hall moraine, was forming in phase A at about 8450 yr B.P. It is roughly contemporaneous with major terminal and recessional moraines that were forming throughout eastern Baffin Island during the Cockburn Substage (8000 to 9000 yr B.P.). The date of moraine formation varied by a few hundred years in different areas along the east coast (Andrews and Ives, 1978). In Cumberland Sound, the nearest area where moraines of Cockburn age have been dated, the ice was retreating from its maximum late Foxe position at the Ranger moraine.
at about 8700 yr B.P. (Dyke, 1979).

The sea entered Hudson Bay at the end of the Cockburn Substage, initiating the final breakup of the Laurentide ice sheet, but the date of 8000 yr B.P. is not significant in terms of moraine formation in the Baffinland Stade (Andrews and Ives, 1978). In the Burton Bay area, moraine ridges were deposited until shortly before 7080 ± 175 yr B.P. during phases B-E. At the head of Cumberland Sound, Laurentide ice retreated slowly from the Ranger moraine, forming eleven recessional moraines before retreating inland at 5700 yr B.P. (Dyke, 1979). In the Home Bay area on the east coast, several recessional moraines formed during retreat from the Ekalugad moraine between 8000 and 5700 yr B.P. (Andrews and others, 1970).

Major shoreline complexes have been associated with moraines of Cockburn age on the east coast of Baffin Island and with later phases of moraine formation on both the east and west coasts (Andrews, 1970b; 1980; Andrews and others, 1972). On the east coast, periods of shoreline formation occurred within the Inugssuit marine regression (1000 to 8000 ± yr B.P.), which followed the deglaciation from moraines of Cockburn age (Andrews, 1980). Several of the strandlines are correlative throughout Baffin Island, and some may be synchronous with glacial events and shoreline deposits in the Burton Bay area. The shorelines may represent a balance between uplift and eustatic sea level rise, as is proposed for shorelines formed in the Cockburn
Substage on the east coast (Andrews, 1980), or they may be related to climatic responses such as a change in effective wave action or an increase in sediment influx (Andrews, 1970a; 1980).

A period of shoreline building at about 7800 yr B.P., recognized on the east coast (Andrews and others, 1972) and in Cumberland Sound (Dyke, 1979), is synchronous with phase B of moraine formation in the study area. The 43-m level of the delta in Burton Bay and possibly the 42-m delta in Lewis Bay formed at this time. The extensive 35-m level of the delta in Burton Bay, which formed about 7500 years ago in phase C, may be contemporaneous with strandlines in Home Bay (Andrews and others, 1970) and on the Ottawa Islands in northeastern Hudson Bay (Andrews and Falconer, 1969).

Moraine formation had ceased before 7080 yr B.F. in the study area, when the ice retreated northwest of Frobisher Bay. In Cumberland Sound and in Home Bay, Laurentide ice remained in contact with the sea until at least 5700 yr B.P. (Dyke, 1979; Andrews and others, 1970). On the west coast of Baffin Island, the sea had entered Foxe Basin by about 7500 to 7000 yr B.P., and the western margin of the Baffin Island remnant of the Laurentide ice sheet lay on the west coast at about 6900 to 6700 yr B.P. (Prest, 1970). During the Isortoq phase, at about 6700 yr B.P., a major moraine system and associated strandlines formed on the west coast during a glacial readvance.
(Andrews, 1970b). The ice was beyond the Burton Bay area at the time of the Isortoq phase, but marine deposits at the head of the bay may be related to the readvance. Terraces and deltas at 24 m at Peterhead Inlet and Apex Hill contain shells dated at $7080 \pm 175$ and $6750 \pm 170$ yr B.P. The 20-m strandline down-bay from Cape Rammelsberg (Mercer, 1956) may also be contemporaneous with the Isortoq phase. Dyke (1979) shows evidence of a strandline in Cumberland Sound that can be dated from his emergence curves at about 6800 yr B.P., and a strandline dated at 6800 yr B.P. occurs in Home Bay (Andrews and others, 1970).

The extensive deposits in the Tatsiujarjualaq River valley cannot be dated more precisely than between about 6700 and 7500 yr B.P. Some deposition may be related to the Isortoq phase, but a significant amount of accumulation had occurred before $7340 \pm 135$ yr B.P. (QC-901), probably during phases D and E.

Erosional terraces and shorelines at 13 to 16 m between Porter Inlet and Peterhead Inlet may be dated by the $6140 \pm 170$ yr B.P. shells at 15 m at Apex Hill (GSC-503). If the date is accurate, the features may be synchronous with strandlines recognized in Home Bay and on the Ottawa Islands (Andrews and others, 1970; Andrews and Falconer, 1969) and with the Eqe phase (6300 yr B.P.) or the Gillian phase (6000 yr B.P.) of stillstand on the west coast of Baffin Island (Andrews, 1970b).
CHAPTER FIVE

CONCLUSIONS

5.1 Summary

The Frobisher Bay moraine system represents a long period of stillstands or readvances of the Laurentide ice sheet after steady retreat from the maximum late Foxe position at the Hall moraine. Detailed mapping of the surficial deposits in the Burton Bay area has led to the recognition of five main phases of ice stillstand and moraine formation within the system. Approximate dates have been assigned to each phase on the basis of the association of moraine ridges and radiocarbon-dated marine shells. Moraine formation began about 8450 yr B.P. in the Cockburn Substage and concluded shortly before 7080 yr B.P. Two later phases of shoreline formation have also been recognized which may be related to stillstands after the ice margin had retreated beyond the head of Frobisher Bay.

The moraine ridges in the Burton Bay area were too old to be dated or correlated by lichenometry, which is a dating technique more useful for Neoglacial deposits. It is unusual, though, that no lichens larger than 92 mm were found on the moraine ridges when lichens measuring
185 to 280 mm have been reported from Cumberland Peninsula (Dugdale, 1972; Miller and Andrews, 1972; Carrara and Andrews, 1972). If the growth rate of *Rhizocarpon geographicum* in the Burton Bay area is similar to that on Cumberland Peninsula (Miller and Andrews, 1972), the largest lichen measured on a moraine ridge is less than 3000 years old.

The ice margin retreated slowly in the Burton Bay area, averaging about 18 m yr\(^{-1}\). The marine limit declines continuously from a high of 179 m beyond the outer moraine ridge to 29 m at the head of Frobisher Bay, reflecting the slow retreat. The regional emergence curve indicates that different rates of emergence prevailed in the periods before and after about 7500 to 7800 yr B.P. Between 8450 and 7800 yr B.P. emergence was rapid, averaging 10.5 cm yr\(^{-1}\); after 7800 yr B.P. the rate of emergence was 3.1 cm yr\(^{-1}\). The available data are insufficient to draw local curves to examine emergence more closely, and attempts to construct local curves using Andrews' (1970a) general equation were unsuccessful because of the large amount of emergence that occurred before about 7500 yr B.P. However, the emergence pattern suggests that a significant amount of isostatic recovery had taken place before deglaciation, so that by the time the ice margin was retreating past Burton Bay and Peterhead Inlet, emergence rates were low.
5.2 Recommendations for Further Study

This paper adds to the regional chronology of late Foxe deglaciation in southeastern Baffin Island presented by Blake (1966) and Milker (1980) by outlining phases of ice retreat within the major moraine system at the head of Frobisher Bay. Additional radiocarbon dates from the area between Apex Hill and the Tatsiujarjualaq River would undoubtedly improve the chronology presented here, particularly if they can be related to specific ice-front positions or relative sea levels. More attention should be focused on identifying and dating deposits below the marine limit in order to determine whether strandlines tilt toward the head of the bay, and more importantly, to document more accurately the relatively long period of rapid emergence before 7500 to 7800 yr B.P.

Additional work is needed to provide a more complete picture of late Foxe deglaciation in Frobisher Bay and southern Baffin Island. The ice retreated rapidly between the Hall moraine and the Frobisher Bay moraine, and no intermediate moraines were deposited. Miller (1978) has identified a prominent strandline dated between 9200 and 9700 yr B.P., but no major strandline associated with moraine formation in the Cockburn Substage has been identified, as in eastern Baffin Island (Andrews and others, 1972; Andrews, 1980). Except for the small high delta in a channel north of Lewis Bay, any evidence of such a strandline would be found down-bay from the Burton
Bay study area. Younger strandlines might also be expected there, related to later phases of ice stillstands identified by the moraines of the Frobisher Bay system.

The chronology of deglaciation presented in this paper could be improved by a comparison with glacial and sea level events represented by deposits within the Frobisher Bay moraine system on the opposite side of the bay. The outer moraine ridge near Cape Rammelsberg was forming at about the same time as the outer moraine ridge on Pichit Peninsula, and marine limits in these areas are similar (Blake, 1966). Later phases of ice stillstand contemporaneous with those identified in the Burton Bay area may be recognizable and perhaps have better dating control. Two recent studies, unavailable when this study was written, should provide additional information on the glacial and sea level events in upper Frobisher Bay (Colvill, 1982; Lind, 1983).

Finally, little is known about ice retreat beyond the head of Frobisher Bay and final disintegration. Blake (1966) suggested that the last remnant of Laurentide ice was located in a crescent-shaped area east of Amadjuak Lake, which was ice-free before 4500 yr B.P. In the Burton Bay area, shorelines formed at about 6750 and 6140 yr B.P. may be synchronous with presently unidentified stillstands of the retreating inland ice.
APPENDIX

Lichenometry was used in an attempt to establish the relative ages of several segments of the Frobisher Bay moraine in the Burton Bay area. Sampling consisted of measuring the largest *Rhizocarpon geographicum* thalli within several sites of 10-m radius spaced 40 to 60 m apart on a transect on the proximal side of a moraine ridge. Five moraine segments were examined: inner ridges north of Tarr Inlet (A), in the valley northwest of Burton Bay (B), and on Laird Peninsula (C), an intermediate ridge further inland (D), and the outer ridge on Fichit Peninsula (E)\(^1\). The five largest lichens at each site were recorded (Table 8). Mean values for each moraine were calculated from the largest values at each of the 5 to 7 sites.

The small number of sites produced representative samples, with the standard deviation within 5 mm of the mean at the 0.95 probability level (Cole and King, 1968, p. 117), for all moraine segments but one. At the outer moraine (E), 22 sites would have been needed for an adequate sample. The large standard deviation here is caused by a clustering of largest lichens into two groups of 70 to 75

\(^1\)Letter designations refer to moraine segments located on Fig. 9 and not the five phases of moraine formation discussed in Chapter 4.
mm and 91 to 92 mm (Table 8). The mean value of 82 mm thus underestimates the date of initial substrate exposure.

A one-way analysis of variance test was used to determine whether significant differences exist in mean lichen size for the five moraine segments. Maximum lichen diameters for each moraine were found to be normally distributed using a Kolmogorov-Smirnov goodness-of-fit test. Bartlett's test of multiple sample variances showed that group variances were approximately equal, in spite of the high standard deviation for one site. As a result, an exact analysis of variance test was run (Table 9). The calculated F-value of 16.90 is greater than the critical F-value of 2.74 at the 0.05 significance level, indicating that the null hypothesis of no significant difference in group means can be rejected. Mean values of maximum lichen diameters therefore differ between moraines.

Paired t-tests were used to determine whether the moraine segments can be grouped according to mean lichen size. Significant differences between moraines are shown in Table 10. Three groups can be distinguished. The outer moraine (E), which has the largest mean value, is separate from the other moraines. The middle segment (D), with the smallest mean, is also distinct. The three inner segments (A, B and C), with intermediate values, form a third group.

Although significant differences exist between the moraines, it is unlikely that the ages of the largest lichens, as determined from the growth curve of Miller and
<table>
<thead>
<tr>
<th>Horaine segment</th>
<th>Maximum diameters (mm) at each site (1-7)</th>
<th>Mean (mm)</th>
<th>Standard deviation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>73 73 70 65 70 65 70</td>
<td>69.4</td>
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<td></td>
<td>70 73 63 56 60 50 65</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>65 65 60 55 55 48 63</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>60 63 55 52 53 45 49</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>60 60 50 50 50 45 45</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>60 65 70 68 70 68 66</td>
<td>66.8</td>
<td>3.82</td>
</tr>
<tr>
<td></td>
<td>55 60 68 63 66 66 66</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>50 58 66 60 63 63 63</td>
<td></td>
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</tr>
<tr>
<td></td>
<td>45 55 55 57 59 60 60</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>42 50 47 55 55 55 53</td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>65 65 70 61 70 60 55</td>
<td>66.2</td>
<td>3.83</td>
</tr>
<tr>
<td></td>
<td>61 64 66 66 57 56 56</td>
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</tr>
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<td></td>
<td>59 61 64 64 50 50 50</td>
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</tr>
<tr>
<td></td>
<td>55 50 55 50 50 50 50</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>50 57 55 60 60 60 60</td>
<td>57.0</td>
<td>4.00</td>
</tr>
<tr>
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<td>45 62 40 47 52 50 50</td>
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<td>40 50 36 45 45 45 50</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>40 40 26 45 45 45 45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>91 92 82 75 70 91 73</td>
<td>82.0</td>
<td>9.45</td>
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<tr>
<td></td>
<td>83 73 73 68 68 60 66</td>
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<td>72 70 70 61 63 75 63</td>
<td></td>
<td></td>
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<td>59 55 55 55 52 61 55</td>
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<td></td>
</tr>
<tr>
<td></td>
<td>54 51 54 51 55 54 54</td>
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</tr>
</tbody>
</table>

*Calculated from largest value of each site*
### TABLE 9
**ONE-WAY ANALYSIS OF VARIANCE FOR LICHEN DATA**

<table>
<thead>
<tr>
<th>Source of variation</th>
<th>Sum of squares</th>
<th>Degrees of freedom</th>
<th>Estimate of variance</th>
<th>F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Total</td>
<td>2928.19</td>
<td>30</td>
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<tr>
<td>Between groups</td>
<td>2114.85</td>
<td>4</td>
<td>528.71</td>
<td>16.90&lt;sup&gt;a&lt;/sup&gt;</td>
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<tr>
<td>Within groups</td>
<td>813.35</td>
<td>26</td>
<td>31.28</td>
<td></td>
</tr>
</tbody>
</table>

<sup>a</sup> Significant at 0.05 level

### TABLE 10
**CALCULATED t-VALUES FOR LICHEN DATA**

<table>
<thead>
<tr>
<th>Moraine</th>
<th>A</th>
<th>B</th>
<th>C</th>
<th>D</th>
<th>E</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>--</td>
<td>1.09</td>
<td>1.28</td>
<td>5.28&lt;sup&gt;a&lt;/sup&gt;</td>
<td>5.52&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>B</td>
<td>--</td>
<td>--</td>
<td>0.23</td>
<td>3.98&lt;sup&gt;a&lt;/sup&gt;</td>
<td>6.37&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>C</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>3.49&lt;sup&gt;a&lt;/sup&gt;</td>
<td>6.32&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>D</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>10.48&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td>E</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
</tbody>
</table>

<sup>a</sup> Significant at 0.05 level
Andrews (1972), represent the age of moraine deposition. What these spatial variations might represent is beyond the scope of this study.
REFERENCES


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SURFICIAL DEPOSITS

Genetic Category | Morphological Modifier | Process Modifier
---|---|---
B bedrock | b blanket | e eroded
C colluvial | f fan | k kettled
F fluvial | h hummocky | w washed
F glaciofluvial | p plain |
L lacustrine | r ridged |
L glaciolacustrine | t terraced |
M morainal | v veneer |
W marine |
URFICIAL DEPOSITS

Morphological Modifier | Process Modifier
--- | ---
b | blanket | e | eroded
f | fan | k | kettled
h | hummocky | w | washed
p | plain
r | ridged
t | terraced
v | veneer
Process Modifier

- e eroded
- k kettled
- w washed
SUFFICIAL DEPOSITS OF THE

BAFFIN ISLA
FIGURE 3

DEPOSITS OF THE BURTON BAY
BAFFIN ISLAND
FIGURE 3

CIAL DEPOSITS OF THE BURTON
BAFFIN ISLAND
THE BURTON BAY AREA, I ISLAND