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Full Name of Author — Nom complet de l'auteur

JAN MICHAEL BEDNARSKI

Date of Birth — Date de naissance	Country of Birth — Lieu de naissance
MAY 4, 1953	CANADA

Permanent Address — Résidence fixe

12803-121 AVE.
EDMONTON, ALBERTA
T5L 2S6

Title of Thesis — Titre de la thèse

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GLACIER FLUCTUATIONS AND SEA LEVEL HISTORY OF CLEMENTS MARKHAM INLET, NORTHERN ELLESMERE ISLAND

by

(C) JAN BEDNARSKI

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

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(Signed) .. J Bednarski .....

PERMANENT ADDRESS:

..... 12803 - 121 Ave. ..

.... Edmonton, Alberta ..

.... T5L 2S6 ....

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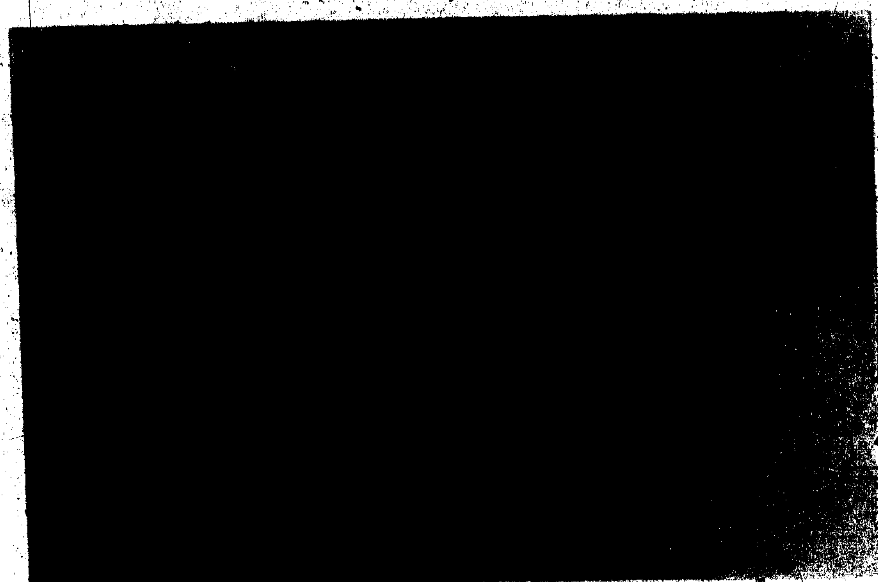

NOAA-6 satellite image showing northern Ellesmere Island and Greenland. Ice-covered highlands are readily distinguished from currently unglaciated terrain. The left central part of the image shows Axel Hieberg Island.

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FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled "Glacier Fluctuations and Sea Level History of Clements Markham Inlet, Northern Ellesmere Island." submitted by Jan Bednarski in partial fulfilment of the requirements for the degree of Doctor of Philosophy

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Supervisor

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Date ..March 21, 1984.....

This thesis is dedicated to my father.

ABSTRACT

Clements Markham Inlet is a major re-entrant on the northernmost coast of Ellesmere Island which cuts into the Grant Land Mountains. The head of the Inlet is bounded on three sides by mountain ice caps. However, a 30 km belt of ice-free uplands occurs along the outer part of the Inlet. Lowlands at the head of the Inlet are also ice-free and are characterized by extensive raised marine deposits. Fieldwork and mapping provides a scenario of former glacier behavior, especially the pattern of ice retreat from confluent positions at the head of the Inlet. Over 40 radiocarbon dates on driftwood and marine shells are used to develop a chronology and to reconstruct the history of sea level.

High-level ice marginal channels and mountain summit erratics indicate that old glaciations inundated the whole of Clements Markham Inlet. At least one of these undated glaciations flowed unconstrained by the local topography and inundated the entire region. In contrast, the most recent glaciation involved confluent trunk glaciers which terminated near the head of the Inlet. Beyond this major terminus, smaller glaciers along the sides of the Inlet debouched into a high full-glacial sea. Initial retreat from the last glaciation is well documented by moraines, kame terraces, and ice contact-deltas.

The oldest date on the terminus of the last glaciation is 9845 BP. After this time, slow retreat was in progress so that some

of the ice margins were within 7 km of their current position by ca. 9.7 ka BP. The mouths of the confluent valleys at the head of the Inlet did not become ice-free until ca. 8 ka BP. However, after ca. 8 ka BP glacial retreat accelerated greatly so that the entire lowland became ice-free within ca. 400 years.

Relative sea level curves concur with the ice load changes suggested by the pattern of recession from the maximum limit of the last glaciation. Distal to the ice limit the full glacial sea attained 124 m a.s.l.. The strandlines indicate that slow emergence occurred from 11-10.5 ka BP to ca. 8 ka BP (0.72 m 100 yr^{-1}). This period was followed suddenly by 'normal' rapid postglacial emergence which decelerated to the present.

The characteristic stratigraphy varies between the proximal and distal sides of the 10 ka BP ice margin. Up-ice of this margin, at the head of the Inlet, the sections commonly show that a marine transgression immediately followed the retreat of a grounded glacier. Conversely, distal to the 10 ka BP margin, along the sides of the Inlet, the stratigraphy shows complex intercalations of marine and glacigenic sediments indicating proximal ice front conditions throughout the whole of the sedimentary record where small valley glaciers contacted the full glacial sea.

The marine limit of the full glacial sea rises from 92 m a.s.l., at the outer coast, to 124 m a.s.l., near the last ice limit at the head of the Inlet. Emergence from this marine limit, on the distal side of the ice limit, occurred simultaneously along the sides of the Inlet. Conversely, the marine limit on the proximal side of the ice limit descends in the up-ice direction. In this area the age of the marine limit is directly controlled by the retreat of the ice front and is therefore progressively younger in the up-ice direction.

Individual strandlines tilt up in a southwesterly direction, towards the central Grant Land Mountains, suggesting a center of isostatic uplift in that area. Regional isobases on the 10 ka and 8 ka BP shorelines show this discrete center which is separated from the Greenland center by a cell of low emergence over the Lake Hazen area. The regional isobases indicate that by 6 ka BP the tilting of shorelines was dominated by the former Greenland ice load.

The history of emergence on northernmost Ellesmere Island is similar to that on northeastern Ellesmere Island and Greenland, but initial emergence occurs two thousand years sooner in Clements Markham Inlet. This may be due to: a dissimilar glacioclimatic regime influenced by the proximity of the Arctic Ocean; or due to a difference in the response times between the Greenland Ice Sheet and the Ellesmere Island ice caps.

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I would like to thank John England, my intrepid supervisor, for introducing me to the arctic and providing support throughout all these years. I will always remember my adventurous first year in 1979, the year of the 'windswept desert'; hours of discussion under the midnight sun with Tom Stewart; and long jaunts during the midday sun with Don Lemmen. The 1980 field season saw us back in full force with 'Big Eric' and 'Hula Danger', tackling the treacherous sea ice and waters of Clements Markham Inlet and fighting off fearsome sea creatures. Little did we know the winged lizzards would descend upon us. By the end of the 1981 season I traversed most of the Inlet with the indefatigable help of Doug Calvert.

I have engaged many other people in discussion concerning my thesis. I would especially like to thank Art Dyke for numerous sessions. Dave Fisher provided information on glaciology and ice-cores of the region, and Hans Trettin provided maps of the bedrock geology. Carolyn King ran the till fabrics, and Anita Moore spent many diligent hours descrambling the manuscript on a word processor. Lastly, I would like to thank my supervisory committee for their constructive criticisms.

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Chapter I

Introduction

1.1 Nature of the Study

Quaternary studies on northern Ellesmere Island were initiated in the early 1950's with reconnaissance work by Prest (1952), Blackadar (1954), Gadbois and Laverdiere (1954), Hattersley-Smith _et al._, (1955), and Christie (1957). Much of this early work was associated with studies on the dynamics of ice shelves and detached ice islands on the northernmost coast (Fig. 1.1; Koenig _et al._, 1952; Marshall, 1955; Crary, 1956, 1958, 1960; Hattersley-Smith, 1957). During the International Geophysical Year 1957-58, geological, glaciological, and geomorphological studies were carried out under the auspices of the Defense Research Board and the Geological Survey of Canada (Christie, 1958, 1959, 1960; Deane, 1958, 1959; Hattersley-Smith, 1958, 1960; Smith, 1959, 1960; Brochu, 1959). Subsequent studies on glacier fluctuations and climatic change (e.g. Hattersley-Smith, 1969, 1971; Hattersley-Smith and Long, 1967; Lyons _et al._, 1972; Lyons and Mielke, 1973) continue to the present (e.g. England, 1974a, 1976a, 1976b, 1978, 1982, 1983; England and Bradley, 1978; England _et al._, 1981; Stewart, 1981; Stewart and England, 1983). Similar reconnaissance studies on the glacial history of adjacent NW Greenland have also been undertaken (Koch, 1928; Davies, 1972; Weidick, 1972, 1976). A detailed summary of previous research on northern Ellesmere Island can be found in England _et al._, (1981b).

This study is an extension of the latter investigations on the Quaternary history of northern Ellesmere Island. However, the majority of previous and contemporary studies were undertaken on NE Ellesmere Island, S of the Grant Land Mountains (Fig. 1.1). Consequently, the purpose of this investigation was to extend the Quaternary data base to the northernmost coast of Ellesmere Island. Clements Markham Inlet was chosen for this work since it is a major re-entrant of the Lincoln Sea, Arctic Ocean extending well into the N flank of the ice-covered Grant Land Mountains. Moreover, previous reconnaissance work by Christie (1967) reported widespread Quaternary deposits in this area which warranted further investigation.

1.2 Objectives of the Field Work

Prior to this study, detailed field work on the glacial history of northernmost Ellesmere Island has not been undertaken despite the area's major ice fields and fiord systems and its direct proximity to the Arctic Ocean (Fig. 1.1). The primary objective of the thesis, therefore, is to expand the glacioisostatic data base for Ellesmere Island and to relate it to the late glacial stratigraphy. This information would be particularly relevant to: the controversy regarding the extent of ice in northern high latitudes during the last glaciation (Blake, 1970, 1975; England, 1976 a and b; England et al., 1981; Hughes et al., 1977); the chronological and paleoenvironmental interpretations of High Arctic ice cores (Dansgaard et al., 1971; England, 1974b; Paterson, 1977; Koerner et al., 1979); the importance of the Arctic Ocean as a moisture source for the growth of northern

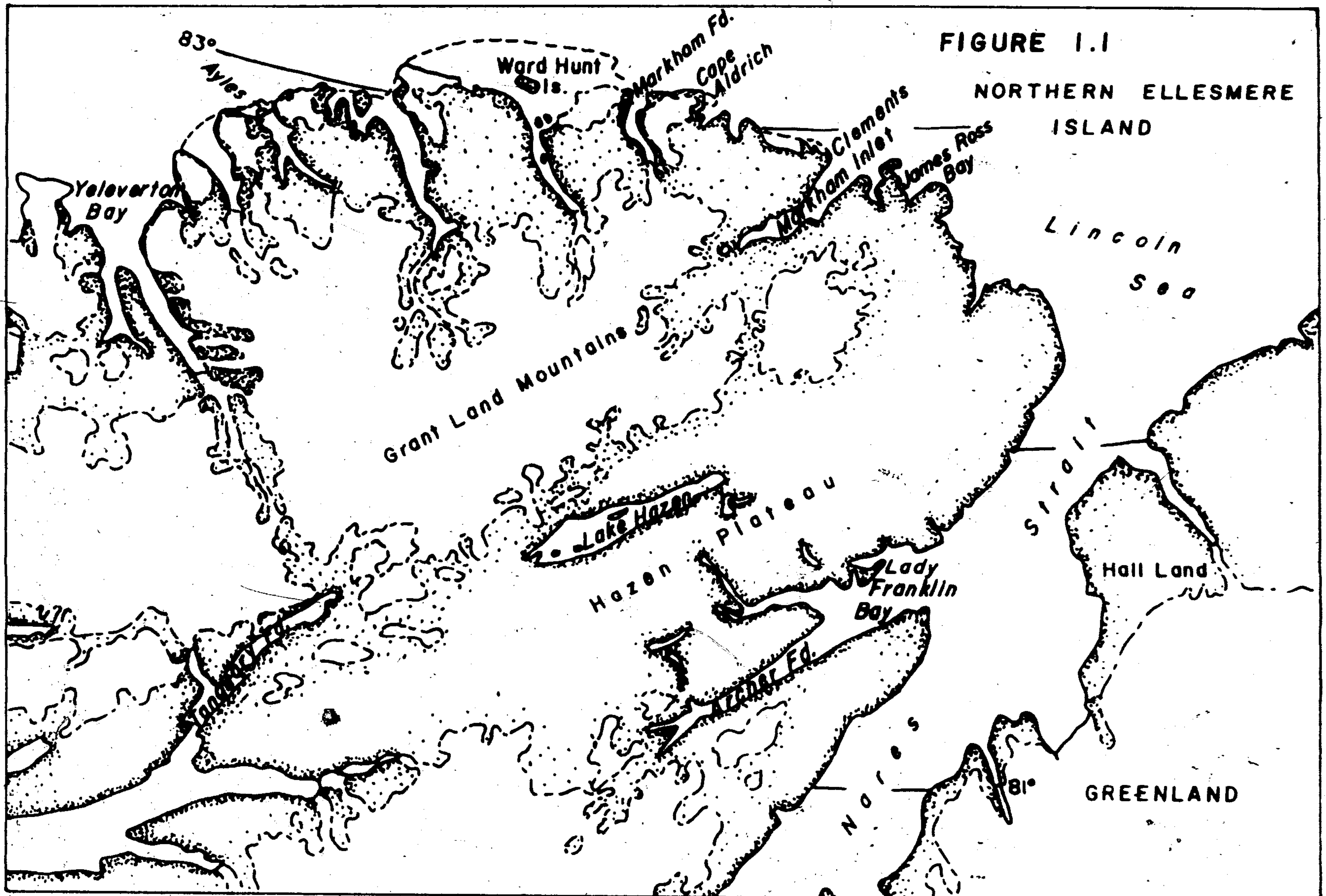

FIGURE 1.1

NORTHERN ELLESMERE ISLAND

high latitude ice sheets (Ewing and Donn, 1956; Hunkins et al., 1971; Herman and Hopkins, 1980); and finally the hypothesis for biological refugia in this area (Brassard, 1971; Leech, 1966).

Specifically, the thesis is concerned with the extent and chronological relationships of past glaciations in Clements Markham Inlet (Fig. 1.1). A second related objective was to determine the nature and pattern of postglacial emergence in order to improve on the provisional postglacial isobases originally drawn for NE Ellesmere Island (cf. England 1976b, 1982).

Field work in Clements Markham Inlet was conducted during the summers of 1979, 1980 and 1981. As a working hypothesis, the author approached the study with a few a priori assumptions. These assumptions were based on the model of glaciation derived from NE Ellesmere Island (i.e. England, 1978). This model suggests that northernmost Ellesmere Island was subjected to at least two discernable glaciations, besides the earliest, limited incursion of the Greenland Ice Sheet onto NE Ellesmere Island. The last glaciation, of Ellesmere Island origin, was less extensive than the preceding one. Moreover, the extent of the last glaciation could be marked by Hazen Moraine equivalents of ca. 8 ka BP or younger age. These hypotheses, however, were taken with caution since there was no reason to believe that the last glaciation, N of the Grant Land Mountains, should necessarily parallel the glacial history to the S, especially because the Arctic Ocean could have provided an important moisture source in the Clements Markham Inlet area. Moreover, the current glacioclimatic regimes of the two areas are dissimilar as the Hazen Plateau is largely ice-free, whereas the Grant Land Mountains are generally covered by ice caps. Hence, field work in this area would help to test the degree of similarity between the N and S flanks of the Grant Land Mountain ice caps during the last glacial cycle. Lastly, since the postglacial emergence on NE Ellesmere Island was dominated in timing and magnitude by the Greenland Ice Sheet (England, 1976a, 1982), it was thought that Clements Markham Inlet, being well removed from the latter ice load, would provide an independent record of postglacial emergence controlled solely by the ice load over the Grant Land Mountains.

1.3 General Physiography of Northern Ellesmere Island

Northern Ellesmere Island contains two major physiographic regions: the Grant Land Mountains in the north, and the Hazen Plateau to the south (Fig. 1.1). The mountains form part of the Innuitian Orogenic System, having predominant NE-SW structural trends, and are principally comprised of clastic and carbonate sequences of eugeosynclinal rocks (Trettin, 1972). Mountain summits range from ca. 1000 m a.s.l. along the rugged northernmost coast, to greater than 2500 m a.s.l. SW of the Inlet where peaks project through ice fields as spectacular nunataks. These thin, topographically controlled ice fields are generally found above 1100 m a.s.l. and attain maximum thicknesses ca. 900 m (Hattersley-Smith et al., 1969). Major trunk glaciers, flowing from these ice fields, feed the north coast fiords forming floating ice tongues as they debouch into the sea.

Nonetheless, the heads of Clements Markham Inlet and Ayles Fiord are presently ice-free, although several valley glaciers in Clements Markham Inlet terminate within ca. 8 km of the present coast.

To the S, the Grant Land Mountains are abruptly bounded by the "Lake Hazen Fault Zone" which forms the NE boundary of the Hazen Plateau (Christie, 1964; Trettin, 1969, 1972). Here the mountain flanks contain several southward flowing valley glaciers which presently terminate along the Lake Hazen Fault Zone. The elevation of the Hazen Plateau is ca. 300 m a.s.l. in this area and rises to ca. 1300 m a.s.l. along its S boundary formed by Archer Fiord and Lady Franklin Bay. Judge Daly Promontory, on the S shore of Lady Franklin Bay/Archer Fiord, forms the northern extension of the Victoria and Albert Mountains (SW of the Hazen Plateau, Fig. 1.1).

In addition to the ice-covered uplands, the Hazen Plateau may have been an important source area for past ice accumulation because much of it is just below the present equilibrium line altitudes. Moreover, several small plateau ice caps already exist along its uplifted S rim (England and Bradley, 1978). The Hazen Plateau is connected to the head of Clements Markham Inlet via a deep ice-free trench, crosscutting the general structural trend of the Grant Land Mountains. The trench, Piper Pass, is ca. 3 km wide and 900 m deep. All drainage from Piper Pass is northward to Clements Markham Inlet, however it is presently blocked by three prominent valley glaciers which in turn dam three proglacial lakes at the S end of the pass.

1.4 Development of the Regional Glacial History of Northern Ellesmere Island

The overall morphology of Ellesmere Island led many workers to conclude that large scale glacial modification of the landscape had taken place. Pelletier (1966) studied the channel bathymetry within the Queen Elizabeth Islands and deduced that they were eroded by large outlet glaciers (i.e. Nares Strait, Fig. 1.1). Certainly, some fiords on the NW coast of Ellesmere Island and elsewhere are overdeepened. For example, Disraeli Fiord (Fig. 1.1) is more than 300 m deep and a submarine canyon, reaching 850 m in depth, trends NW from the fiord mouth (Crary, 1956). However, deep, interisland channels may just as easily reflect the tectonic evolution of the arctic islands. That major outlet glaciers may have occupied fiords may be indicated by shoals ca. 15-20 km off the N coast, in the vicinity of Ward Hunt Island, which Crary (1956) interpreted as terminal moraines. However, again, there is no other evidence to support this speculation.

Similarly, Taylor (1956) reasoned that the well-developed fiords on the W coast of Ellesmere Island could not have been eroded by the diminutive Ellesmere Island ice caps, and therefore, he assumed that the Greenland Ice Sheet had once inundated Ellesmere Island. Smith (1961) also noted the absence of fiords between 78° and 82° N on the Island's E coast but attributed this to a protective ice cover which occurred when the Greenland and Ellesmere Island ice sheets merged. However, in an earlier paper, Smith (1959) rejected the hypothesis that Greenland ice had completely inundated Ellesmere

Island, since only Ellesmere Island erratics (chert conglomerates and quartzites) were found beneath mountain summits in the southern Grant Land Mountains. A former N to S ice flow was also recognized on the Hazen Plateau, reinforcing this view. Moreover, Hattersley-Smith (1961) suggested that Ellesmere Island ice had formed a protective cover over the topographically-uniform Hazen Plateau, whereas preglacial fluvial valleys on the W coast of Ellesmere Island had been extensively modified by major trunk glaciers flowing from local highland ice caps. However, none of the authors suggest specific glaciological conditions which would lead to such differential, large scale glacial erosion (e.g. warm versus cold-based ice). Moreover, it is now clear that the Grant Land Mountains and the Hazen Plateau have had distinct tectonic histories and are separated by major faults *(Trettin, 1972).

Definite evidence that Ellesmere Island was previously inundated by a substantial thickness of ice comes from the distribution of glacial erratics. On the northernmost coast Hattersley-Smith et al., (1955) reported a gabbro erratic at 335 m a.s.l. on Ward Hunt Island, whereas, Disraeli Fiord, to the S, was considered glaciated to 760 m a.s.l.. Lyons and Mielke (1973) thought erratics, up to 600 m a.s.l. in this area, represented a substantial thickness of ice during the last glaciation. Moreover, Fyles and Craig (1965) suggest that the erratic distribution of NW Ellesmere Island shows an ice flow across the general trend of ridges and valleys (ca. 900 m of relief). Christie (1967) notes erratics, from 880 to 1370 m a.s.l., below the summits of the United States Range, Grant Land Mountains. England (1978) found Ellesmere Island erratics up to 830 m a.s.l. at the head of Archer Fiord indicating at least a 1000 m thickness of ice. Similarly, Hattersley-Smith (1969) also estimated a maximum thickness of 1000 m for the former glacier which occupied the head of Tanquary Fiord, 150 km to the W (Fig. 1.1).

Evidence from erratics also indicates a limited incursion of the Greenland Ice Sheet onto Ellesmere Island. Christie (1967) mapped granite and gneiss erratics up to 760 m a.s.l., and 40 km inland, from the NE coast of Ellesmere Island. The profile of uppermost Greenland erratics has been subsequently mapped along 70 km NE Ellesmere Island, where they occur only 5-15 km onto Judge Daly Promontory, and descend rapidly to the south away from the presumed ice source in Petermann Fiord (England and Bradley, 1978). This advance, however, predates the maximum Ellesmere Island advance in the same area (England, 1978).

From the evidence cited above, it is clear that a number of authors agree that at some time in the past a penultimate glaciation occurred, however, controversy arises as to the exact timing of this event. Inspired by the "Flintian" view of maximum late Wisconsin ice (cf. Flint, 1971) a number of workers assumed that Ellesmere Island's last glaciation[1] was essentially all-pervasive (e.g. Blake, 1970;

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[1]The term "last glaciation" or last "glacial cycle" is used in this thesis to mean the latest major glacial episode in this area. This episode culminated during the late Wisconsin stage, but terminated in the earliest Holocene, and may probably be regarded as the approximate equivalent of the late Wisconsin stade of the Laurentide Ice Sheet to the S (cf. Dyke, 1983).

Hughes *et al.*, 1977). Conversely, others have found evidence supporting a limited ice cover during the last glaciation (e.g. England, 1974, 1976a, 1976b, 1978; England *et al.*, 1978, 1981a; Paterson, 1977). This dichotomy is examined in the following section.

1.5 The Last Glaciation

The previous section noted that problems exist in delimiting the limits of the last glaciation in the area. Clearly, one of the complications is the polycyclic nature of Quaternary environments. For example, the region may have been subject to any number of glacial cycles in the past, each of which contributed to the overall evolution of the contemporary landscape (e.g. fiord erosion, erratic distribution). To date, two major lines of evidence have been used to overcome this problem and to determine the extent of ice during the last glaciation on Ellesmere Island. One is the inverse method of noting the amount and pattern of glacioisostatic emergence; the second is using direct glacial geologic evidence.

The amount of postglacial emergence is commonly thought to reflect the magnitude of a former ice load. Blake (1970) found a ridge of greatest emergence running from Bathurst Island to Eureka Sound. From this he concluded that an "Innuitian Ice Sheet" existed and that it inundated the Arctic Islands north of Lancaster and Viscount Melville Sounds and coalesced with the Greenland Ice Sheet in Nares Strait during the last glaciation. Walcott (1972) also contoured a ridge of maximum emergence, which he named the Innuition

Uplift, extending from near Bathurst Island, across Ellesmere Island, to NW Greenland.

England (1972a, 1982), on the other hand, interpreted the emergence data in a different manner. He suggested that uplift over the eastern Queen Elizabeth Islands had a gradient and orientation reflecting the dominance of the Greenland Ice Sheet to the E. He argued that since the lithosphere is rigid, an ice load will depress the land beyond its margin (Brotchie and Silvester, 1969; Walcott 1970), and therefore Greenland and Ellesmere Island ice did not necessarily have to coalesce to produce this emergence pattern. Using a theoretical geophysical model (Brotchie and Silvester, 1969), England (1974) reproduced the postglacial isobases (isopleths of emergence) by assuming an advance of 100 km by the Greenland Ice Sheet and ca. 60 km by the Ellesmere Island ice. However, this model may not be entirely realistic as it assumes complete isostatic equilibrium and no restrained rebound before deglaciation. Conversely, Weidick (1976), took the same emergence data and contoured it to indicate maximum uplift from Greenland towards Ellesmere Island. Paterson (1977) examined the conclusions of the above authors, and noted that different authors provide dissimilar interpretations of the same emergence data. He also questioned whether isostatic equilibrium was attained within each successive glaciation. For this reason primary geological evidence must be given precedence over uplift data in order to determine the extent and timing of different glaciations (e.g. Boulton, 1979). Moreover, the emergence data can only have meaning if considered under the constraints of the direct glacial record.

Delimiting the margins of the last glaciation from glacial geologic evidence comes primarily from datable marine and terrestrial deposits, which can be related to specific ice margins and simple morphostratigraphic relationships. The most elementary method is determining the relative extent of surface weathering and attributing it to duration of exposure. For example, Hattersley-Smith (1969) noted a number of areas on northernmost Ellesmere Island with well developed dendritic drainage systems and fluvially eroded 'V' shaped valleys. He attributed the lack of glacial modification to a prolonged period of fluvial erosion since the area was last occupied by ice (possibly during pre-Wisconsin time). In addition, he described till of "two different ages" on the bases of morphological preservation and lichen cover in Tanquary Fiord. The lower, less extensive till cover was considered to be from the last glaciation. Dates on the oldest postglacial shells and initial peat growth indicate that deglaciation of the fiord head had taken place prior to 6.8-6.5 ka BP (Hattersley-Smith and Long, 1967). Moreover, Hattersley-Smith (1969) recognized that many of the larger scale glacial features could date from an earlier period than the Wisconsin.

Surfaces with deep weathering, usually containing tors and blockfields, are often cited as evidence of nonglaciation during at least late Wisconsin time (cf. Dyke, 1976; Ives, 1978). England and Bradley (1978) described such extreme surface weathering within the zone of Greenland erratics on NE Ellesmere Island. This, they note, is similar to the advanced weathering described for the outermost glacial deposits on Inglefield Land, NW Greenland (Tedrow, 1970).

Furthermore, England (1978) dated fossiliferous Greenland till within this weathered zone >80 ka BP (based on amino acid age estimates). Stratigraphically, this represents the latest advance of the Greenland Ice Sheet onto Ellesmere Island and from the distribution and profile of Greenland erratics England and Bradley (1978) concluded that the Greenland Ice Sheet's maximum extent did not exceed ca. 100 km beyond its present margin. Moreover, England and Bradley (1978) also describe a subsequent Ellesmere Island ice advance, also quite weathered, which cross-cuts the lower elevations of the Greenland erratic zone. This advance was characterized by thin topographically-controlled ice lobes which terminated in Kennedy Channel, producing small ice shelves when relative sea level was ca. 162-175 m a.s.l. (England _et al._, 1978, 1981a). Marine terraces bordering the former ice shelves provided minimum radiocarbon dates of >30 ka BP, whereas, amino acid age estimates are >35 ka BP (England _et al._, 1981). Since these deposits show no evidence of being overridden, England and Bradley (1978) conclude that this coastline was entirely ice-free during the last glaciation. Earlier, England (1974a) identified morainal belts near the heads of fiords and bays, termed the Hazen Moraines, which he suggested were the limits of the last glaciation on NE Ellesmere Island.

As a reason for limited ice advance in the High Arctic, England (1976), and England and Bradley (1978), suggested that advection of moist air from the S was inhibited by the Laurentide Ice Sheet during the last glaciation. They noted that the formation of the Hazen Moraines on NE Ellesmere Island probably did not occur until ca. 8 ka BP, possibly in response to the Laurentide Ice Sheet breaking up in the S. However, recent evidence, particularly from the raised marine record, indicates that the ice reached the Hazen Moraines before ca. 11 ka BP and maintained this position until ca. 8 ka BP. A period of slow recession followed and was succeeded by rapid deglaciation 6.2 ka BP (England, 1983). Moreover, an isostatically depressed, ice-free corridor existed between the Greenland and Ellesmere Island Ice Sheets which emerged synchronously during initial deglaciation (England, 1976, 1978). By establishing emergence curves for NE Ellesmere Island, England (1983) is able to distinguish between areas that were ice-covered or ice free during the last glaciation.

Evidence supporting ice-free zones during the last glacial cycle is also found on the Greenland side along northern Nares Strait. Tedrow (1970) studied three successive glacial advances in Inglefield Land and found the most recent one extended only 15 km beyond the present ice margin. Moreover, he obtained a date of ca. 20.8 ka BP from organic debris in a terrace, 50 km from the present margin. Davies (1972) also recognized three glacial advances in Hall Land, NW Greenland. The last, which he believed to be late Wisconsin in age, extends only 12 km beyond the present ice margins. Weidick (1976) also found no evidence of glacial advances over marine sediments dated ca. 19 ka BP, 50 km beyond the present ice in Orlik Fiord, NW Greenland. Lastly, Christie (1975) identified three moraine systems in Peary Land. The oldest moraine represents expansion of the Inland Ice from the S, whereas, a 'local' moraine was derived from a subsequent advance of the Nordkronen ice cap to the N. The youngest moraine may represent either a late-glacial readvance or a stillstand during retreat of the Nordkronen ice cap, however, Christie (1975) gives no ages for these advances.

The above studies suggest that during the last glaciation there was a restricted ice advance bordering northern Nares Strait, and hence, the coalescence of the Greenland and Innuitian Ice Sheets is not presently supported by field evidence. Furthermore, both botanical and zoogeographical evidence suggest the likelihood of refugia on northern Ellesmere Island during at least the last glaciation (Leech, 1966; Brassard, 1971).

Nonetheless, proponents of the maximum late Wisconsin model remain. In addition to the uplift data discussed, the main arguments for this model are three-fold. The first is the presence of glacial striae. Blake (1977a) found fresh striae on surfaces at Pim Island and Cape Herschel, east-central Ellesmere Island, and he suggested that they were probably produced by southward flowing Greenland ice entering Kane Basin during the last glaciation. In addition, Blake (1977a) noted that, in general, freshly scoured bedrock is confined to low-lying areas, whereas highly weathered rock mantles the uplands. He explained this by invoking an ice cover with cold-based conditions in the uplands and pressure melting in the lowlands. Freshly preserved striae were also found on the Carey Islands in NW Baffin Bay, off the Greenland coast (Blake, 1977b), suggesting extensive advance during the last glaciation. Surface striae are generally found only on fresh, recently deglaciated surfaces, since resistance to subaerial exposure is poor. However, Boulton (1979) points out that the presence of striae does not imply that they were produced during the last glaciation. Ambiguity can be introduced if 'fresh' striae, of unknown age, are long protected by a mantle of overburden, to be exposed at a later date. Moreover, postglacial high sea levels may contribute to exhuming striae at lower elevations. In addition, England *et al.*, (1981) noted that some of the higher-level striations observed by Blake (1977) could have been produced by local glaciers, rather than by a greatly expanded Greenland Ice Sheet. A number of radiometric dates on *in situ* shells also suggest the Carey Islands were deglaciated >32 ka BP (Davies *et al.*, 1963; Bendix-Almgreen *et al.*, 1967), and therefore escaped the last glaciation.

The second argument by proponents of the maximum ice cover hypothesis is that numerous radiocarbon dates in the High Arctic, particularly those in the 20-12 ka BP range which define past ice limits, are anomalous. The general assumption is made that these dates are due to mixing of 'old' material with Holocene material (Blake, 1974, 1976). Although this may be the case for samples of shell fragments or accumulated organic material, it would be unlikely for *in situ* shells in marine deposits. Moreover, numerous 'old' deposits occur in the High Arctic which show no signs of being overridden by subsequent glacial advances (England, 1976b). Given this evidence, it would seem logical to expect materials which would also produce intermediate dates between early Holocene and infinite, particularly if the areas were ice-free during the last glaciation.

Lastly, the third argument, related to the previous two, deals with the actual location of ice margins at the height of the last glaciation (ca. 18 ka BP along the southern Laurentide Ice Sheet; Dreimanis and Karrow, 1972). The proponents of the maximum model maintain that such margins were generally offshore and consequently unmappable. Furthermore, they maintain that the numerous well established early Holocene ice margins were merely stillstands during deglaciation (cf. Prest, 1970; Andrews, 1973; Blake, 1977). Although this hypothesis provides a reason for the missing 18 ka BP ice margins, the consequence must be that large areas of extensively weathered terrain and undisturbed 'old' deposits must have been preserved by 'non erosive' cold-based ice. Although small, localized areas may survive glaciation with little alteration, no modern day analogs occur where large, unaltered areas have been recently exposed by retreating cold-based ice. Furthermore, areas which are thought to have been covered by cold-based ice during the last glaciation on Somerset Island left many clues of previous occupation (Dyke, 1983). A final point is that if the 18 ka BP margins were less extensive than subsequent early Holocene margins, they would be equally as difficult to find.

Dichotomy concerning the extent of the last glaciation is not only confined to the terrestrial and marine records. Reynaud and Lorius (1973) suggested that the gas content in ice dating from the last glaciation from the Camp Century core, NW Greenland, indicated that it originated 1300 m above the present ice divide. Consequently, it was suggested that it originated from an earlier ice divide which may have formed over Kane Basin (Robin, 1973). However, Paterson (1977) provided an alternative explanation by suggesting that the ice may have originated from the main Greenland ice divide to the E, and not the present subsidiary one. Koerner (1977) stated that a restricted ice cover in the Arctic Archipelago during the last glaciation agrees with the isotopic record from the Devon and Ellesmere Island ice cores, in that major thickening of the ice caps was not recorded.

In summary, the above review emphasizes a number of problem areas concerning the glacial history of northern Ellesmere Island. One problem is largely a product of the scarcity of data from large areas - for example, the uneven distribution of accurately measured and dated strandlines between Greenland and Ellesmere Island (>100 points exist on Ellesmere Island, compared to ca. 5 on the adjacent Greenland coast). Consequently, any number of isobase patterns can be drawn for the same data base. Secondly, indirect evidence of former ice loads based on postglacial isobases must be classified by direct morphostratigraphic and stratigraphic evidence delimiting these former ice margins. This, in turn, can be demonstrated only by dated moraines and ice-contact deltas (the glacial limit) beyond which an ice marginal depression may be identified. Moreover, because very few ice margins are actually dated in the High Arctic, other than the Hazen moraines on NE Ellesmere Island and west-central Ellesmere (Hodgson, in press), the relation between actual ice coverage and rebound in this area remains speculative. Hence, it is clear that the controversy with respect to the extent of the last glaciation can only be resolved by field work.

1.6 Post Last Glaciation

The glacioclimatic history of northern Ellesmere Island, following major deglaciation in the early Holocene, is less controversial. It is generally recognized that a mid-Holocene climatic deterioration occurred. This is indicated by glaciers which advanced into V-shaped valleys and over raised beaches, and by piedmont glaciers which advanced into main valleys causing divergence of drainage. In the Tanquary Fiord area, dates on buried organic material suggest this occurred ca. 4 ka BP (Hattersley-Smith, 1969). Moreover, organics underlying till adjacent to the Eugene Glacier dated 4.9 ka BP (Lowdon and Blake, 1979). Ice shelves also began to form or at least re-exposed on the N coast during this interval (<4 ka BP; Crary, 1960; Lyons and Mielke, 1973). Furthermore, dates on organic deposits near the snouts of the Gilman and Air Force glaciers indicate that during the last 900 years there has been little change in their terminal position (Trautman, 1963; Lowdon et al., 1967). Currently, however, numerous larger glaciers appear to be advancing while several smaller glaciers have receded from moraines during the last 40 years (Hattersley-Smith, 1969).

Chapter II

Physical Characteristics of the Study Area

2.1 Physiography

Clements Markham Inlet lies along the northernmost coast of Ellesmere Island, Northwest Territories (82°42'N, 67°30'W; Fig. 1.1). The Inlet forms a SW trending reentrant which extends from the Lincoln Sea into the Grant Land Mountains, and is ca. 50 km long and averages 8 km in width. Its origin appears to be the result of Paleozoic faulting (Blackadar, 1954; Christie, 1964) rather than glacial erosion. Nevertheless, subsequent glacial erosion most certainly occurred as glacial polish and erratics have been observed in this area to 600 m a.s.l. (Hattersley-Smith _et al._, 1955; Christie, 1967; this study).

The head of Clements Markham Inlet comprises a lowland (ca. 200 km^2) which is occupied by the contemporary sandar of the Clements Markham River and Piper Pass valleys. Raised marine silt plains which are extensively gullied are also found in this area. The Grant Land Mountains, surrounding the head of the Inlet, reach 1500 to 1000 m a.s.l. and progressively decrease in elevation to ca. 600 m a.s.l. at the outermost coast. Outlet glaciers emanating from ice caps on the Grant Land Mountains terminate 8 to 30 km from the Inlet shore (Plates 2.1, 4.2). Because the mountains decrease in elevation toward the

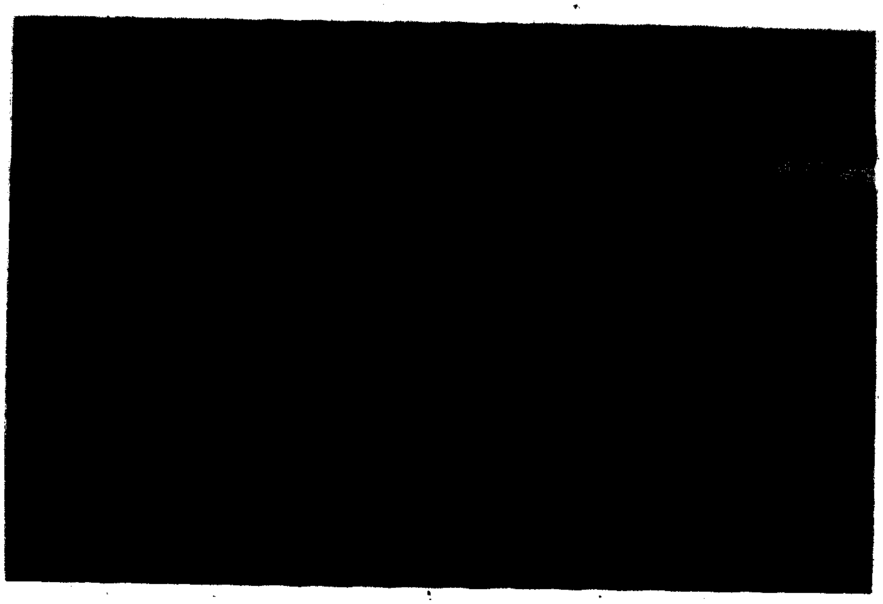


Plate 2.1 Outlet glaciers in the Upper Clements Markham River valley looking south. The Barrier Glacier is seen as it flows into the valley from the east. The terminus of this glacier lies ca. 30 km upvalley from the present shore of the Inlet. The Clements Markham Glacier lies further upvalley, immediately to the right and is not visible.

outermost coast only a few isolated snowfields and minor cirque glaciers exist in sheltered localities. In the relatively ice-free outer 30 km, the landscape exhibits a predominantly fluvial character with dendritic drainage patterns and 'V' shaped valleys dissecting the alpine topography. Another noteworthy feature is that the relative size of the deltas and raised marine deposits also decreases from the head of the Inlet to the outer coast. This aspect will be discussed in Chapters III and IV. Place names used throughout this thesis (official and unofficial) are provided in Figure 2.1.

2.2 Geology

Northern Ellesmere Island falls within the Innuitian Orogenic System (Fortier et al., 1954) which has a complex history of deposition and tectonic activity spanning late Proterozoic to late Tertiary time (Trettin, 1972). Proterozoic, intrusive-metamorphic terrane on the NW tip of Ellesmere Island, comprising part of the Pearya Geanticline, is bounded to the SW by several geosynclinal belts, associated with the early Paleozoic Franklinian Geosyncline and the late Paleozoic to mid Tertiary Sverdrup Basin, which have a predominant SW-NE structural trend. These two periods of sedimentation were terminated by uplift during the Ellesmerian and Eurekan Orogenies respectively. Figure 2.2 depicts the geology of the Clements Markham Inlet area.

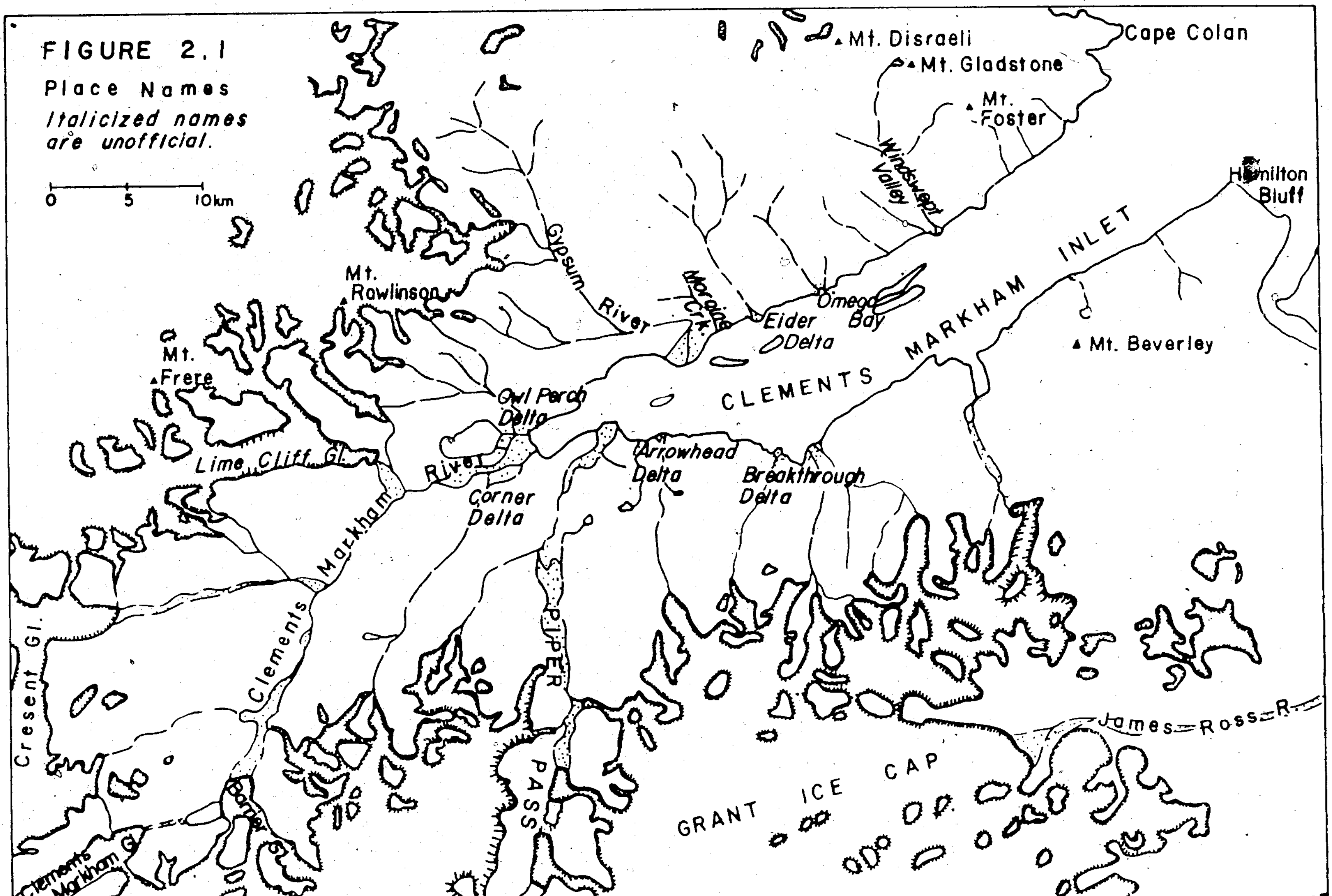


FIGURE 2.1
Place Names
Italicized names are unofficial.

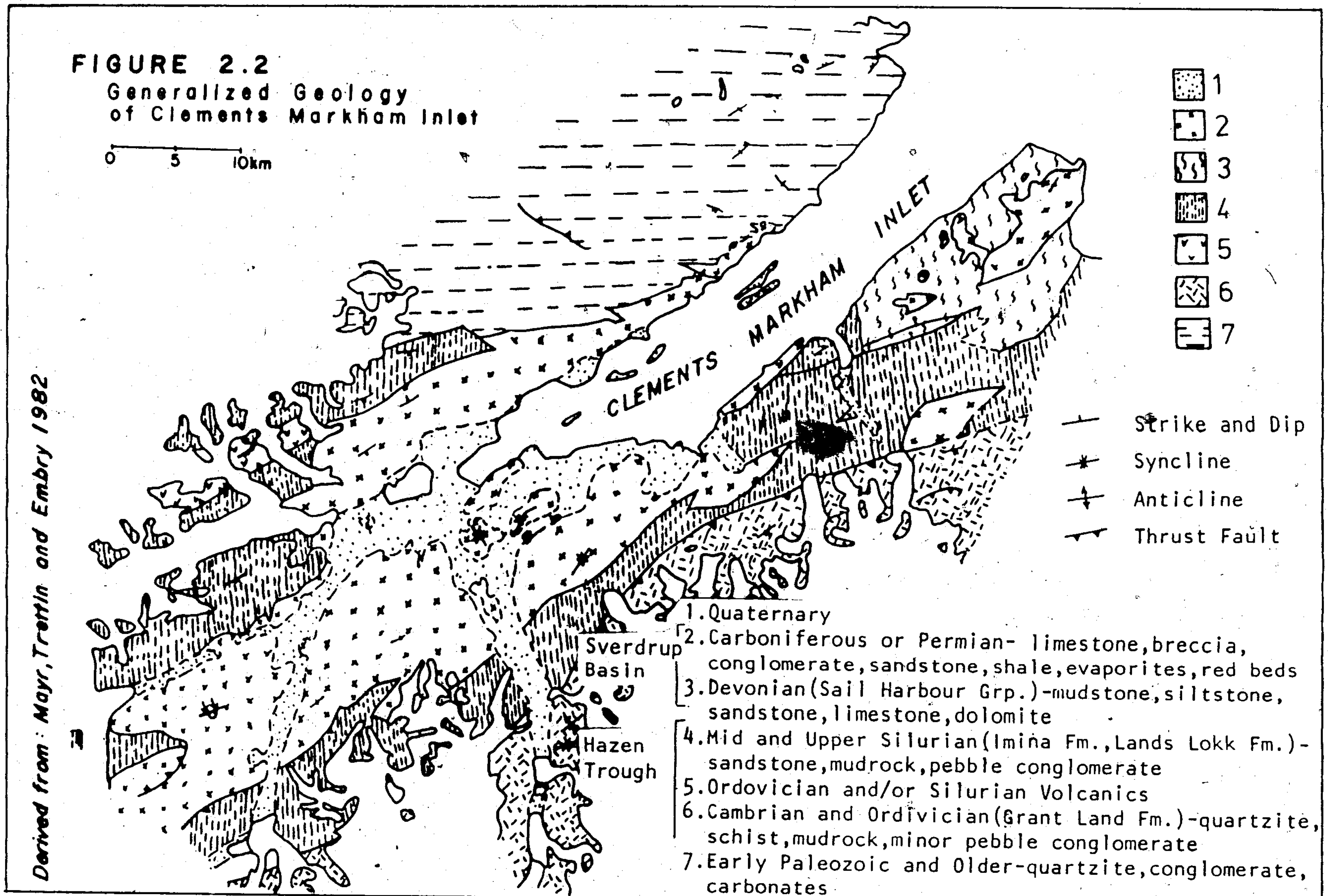


FIGURE 2.2
Generalized Geology of Clements Markham Inlet

2.2.1 Pearya Geanticline

The intrusive-metamorphic terrane of the northernmost coast, between Cape Aldrich and Markham Fiord (Fig. 1.1) is known as the Cape Columbia Complex (Trettin, 1969). It is comprised of ultramafic intrusions, granitic and pegmatitic dykes, biotite and granitic gneisses, and schists (Trettin, 1969; Frish, 1974). The intrusions have a wide metamorphic range and span a considerable time range, dating from late Precambrian to early Devonian (Frish, 1974; Sinha and Frish, 1976; Mayr et al., 1982). The Cape Columbia Complex formed part of the Pearya Geanticline which supplied clastic sediments to the Franklinian Eugeosyncline to the SE (Trettin, 1971). The Cape Columbia Complex is important to this study because it probably provided the ice-rafted granite and gneiss erratics found below the marine limit in Clements Markham Inlet. However, an alternate source may be the Greenland Shield because lithologies from these two areas are indistinguishable (Trettin, pers. comm., 1981). Nevertheless, eastward flowing surface currents along the N coast of Ellesmere Island make Cape Columbia a more likely source.

2.2.2 Franklinian Geosyncline

Lower Paleozoic strata in the study area were deposited during several phases of the Franklinian Eugeosyncline and Hazen Trough, a transitional region between typical miogeosynclinal and eugeosynclinal deposits (Trettin, 1969). Lithologic suites range from limestones and dolomites to quartzites, sandstones and conglomerates.

The principal units found in the study area are shown in Figure 2.2.

The lower Paleozoic rock units were tightly folded during the mid-Paleozoic Ellesmerian Orogeny. This period is characterized by two main phases of uplift (middle to late Devonian and Mississippian) during which the area occupied by the southern Grant Land Mountains was uplifted relative to the regions SE and NW (Grant Land Uplift; Trettin, 1971). Tight NE trending folds, originating from the uplift, are exposed along the S margin of the Grant Land Mountains. Trettin (1971) also suggested that additional movement in this area occurred along a number of elongate grabens. Furthermore, trends in the fold axes suggest that major thrust faulting at the site of the Clements Markham Inlet trough occurred during the Ellesmerian Orogeny (Blackadar, 1954; Christie, 1964).

2.2.3 Sverdrup Basin

A large part of the study area is comprised of upper Paleozoic rocks. These units were deposited into the Sverdrup Basin, a successor basin to the Franklinian Geosyncline whose location was controlled by the latter's "deep seated crustal structure" (Trettin, 1972). These rocks comprise the central part of the Grant Land Mountains (Fig. 2.2). Lithologies of nunataks in this area include limestone, sandstone, and chert pebble conglomerate (Christie, 1964). The chert pebble conglomerate, with its maroon sandstone matrix, produces distinct glacial erratics on the Hazen Plateau.

The mid-Tertiary Eurekan Orogeny terminated the Sverdrup Basin and represents the rejuvenation of the Grant Land Uplift (Trettin, 1973). The main movements during this time occurred along the Lake Hazen Fault Zone, a belt of thrust faults bounding the United States Range on its SE perimeter (Trettin, 1971). The Porter Bay Fault Zone has a subparallel trend to Clements Markham Inlet, and can be traced from Porter Bay, on the Lincoln Sea, to ca. 16 km SW of the Barrier Glacier, within Piper Pass (Fig. 2.2). This fault zone separates the Permo-Carboniferous units of the Sverdrup Basin from the lower Paleozoic units (Trettin, 1971). The oldest known beds which were undisturbed by the Eurekan Orogeny are Pleistocene (Trettin, 1973), whereas syntectonic sedimentation deposited Oligocene conglomerates in the Lake Hazen Basin (Miall, 1979).

2.2.4 Erratics as Indicators of Glacial Flow

As noted, chert pebble conglomerates from the Grant Land Mountains form distinct glacial erratics in this region, however, specific flow patterns cannot be reconstructed from these erratics due to their large area of provenance. Dioritic, diabasic and basaltic dykes and sills of Proterozoic to Mesozoic age also occur in abundance in northern Ellesmere Island (Christie, 1964), and hence are not generally useful indicators of glacial flow. However, characteristic lower Paleozoic (Ordovician?) volcanics outcrop in a small area S of Mt. Rawlinson and NE of Mt. Frere and in Proterozoic terranes near the Gypsum River, and are useful flow indicators (Fig. 2.2; Trettin, pers. comm. 1981). These erratics are further discussed in Chapter IV.

2.2.5 Deep Structure and Geophysical Characteristics

The lithospheric characteristics of northern Ellesmere Island are of particular relevance to this study because any interpretations of glacioisostatic adjustments are dependent upon 'normal' lithospheric conditions (cf. Walcott, 1970; England, 1982). These general conditions are discerned by crustal thickness, gravity and geomagnetic anomalies, and seismicity.

Trettin (1972) found that most of the Queen Elizabeth Islands have an average crustal thickness of ca. 35 km, which is normal. The Hazen Plateau near Alert has an estimated crustal thickness of ca. 23 km (Utsu, 1966) whereas the United States Range has roots ca. 6 km deeper (Trettin, 1972). Approximately 100 km eastward, on Hall Land, NW Greenland, there is also a normal crustal thickness of 32 km, which gradually thins to 18 km beneath the Lincoln Sea (Fig. 1.1; Sobczak and Stevens, 1973). Moreover, Niblett and Whitham (1970) also found normal levels of heat flow on northern Ellesmere Island, indicating a normal lithospheric thickness.

Bouguer gravity anomalies indicate a major depression (-120 mgal) centered over the United States Range, and a high (+90 mgal) over the N Lincoln Sea where water depths exceed 2000 m (Trettin, 1972). Trettin (1972) noted that the negative anomalies correlated with the high relief of the United States Range and consequently he concluded that NE Ellesmere Island was in a state of isostatic equilibrium. Furthermore, the trend in the gravity field conforms with the predominant structural trend and does not appear to be interrupted by Nares Strait, a prominent submarine rift valley (Kerr, 1967). A large geomagnetic anomaly occurs along a narrow zone from Alert to Eureka (Alert Geomagnetic Anomaly) and is thought to be caused by a long conducting body in the lower crust (Praus et al., 1971). The anomaly coincides with the axis of the former Hazen Trough and may be a relic geosynclinal or orogenic feature (Niblett and Whitham, 1970; Trettin, 1972), but its effect on glacioisostasy is not known.

Lastly, although the lithosphere appears 'normal', postglacial faulting could also produce anomalies in the glacioisostatic record by displacing sea level indicators. Numerous faults crosscut the Phanerozoic sediments (Christie, 1964; Trettin, 1971) and may have been reactivated during glacioisostatic unloading. Stein et al., (1979), have estimated that postglacial crustal flexure may have attained stresses which were high enough to reactivate pre-existing faults in eastern Canada. Contemporary seismicity in parts of Arctic Canada has also been attributed to triggering by glacioisostatic unloading (e.g. Basham et al., 1977; Wetmiller and Forsyth, 1978), however, northern Ellesmere Island is currently aseismic (Wetmiller and Forsyth, 1978). Although abundant evidence of glacioisostatically-induced faulting occurs in northern Sweden (Lundquist and Lagerbäck, 1976), and to some extent in southern Canada (cf. Adams, 1981), England (1982) has found that postglacial isobases are spatially consistent throughout NE Ellesmere Island, particularly across Nares Strait rift valley. In conclusion, no clear evidence currently exists for postglacial faulting on northern Ellesmere Island, although the possibility cannot be ruled out entirely, particularly during rapid deglaciation.

2.2.6 Summary

The geology of Clements Markham Inlet is characterized by late Proterozoic to late Tertiary deposition of sedimentary rock into two successive geosynclinal basins, each terminating with an orogeny which deformed the pre-existing rocks. The Inlet lies parallel with the general SE-NW structural trend and was the locus of mid-Paleozoic tectonism which initiated its formation. Distinct volcanic outcrops in the area provide erratics which record former directions of glacier flow. Lastly, geophysical studies indicate that the lithosphere and crust are 'normal' on northern Ellesmere Island, and imply a predictable glacioisostatic response. Postglacial faulting may produce an anomalous response, but no evidence of faulting currently exists in the Pleistocene deposits of the area.

2.3 Climate

Northern Ellesmere Island has a continental climate which reaches its extreme degree in the interior, and is only slightly less so along the margins of the sea ice covered Arctic Ocean. The area can be classified as a polar desert due to its low precipitation and a net annual water balance close to zero (Bovis and Barry, 1974). Due to its high latitude northern Ellesmere Island remains in total darkness for several months of the year, and hence, winter temperatures are severe. Well developed periglacial features are common in the area. Englacial temperatures remain well below 0°C throughout the year and consequently the glaciers and ice caps display polar and subpolar morphologies (cf. Ahlmann, 1948).

Nonetheless, a striking characteristic of northern Ellesmere Island's climate is its high spatial and temporal variability. Temperatures vary widely depending on the proximity of the moderating ocean waters and the seasonal extent of sea ice cover. For example, Alert (ca. 100 km E of Clements Markham Inlet) has a mean annual temperature of -17.8°C, but this coastal site averaged ca. 5°C warmer during the 1957-1958 winter than did the inland site at Lake Hazen, only 120 km SW of Alert (Fig. 1.1). Conversely, the mean maximum summer temperature for Alert is only 6.9°C, whereas it is 10°C at Lake Hazen (Meteorological Branch, 1970).

Annual precipitation also varies from as low as 25 mm on the Hazen Plateau (Jackson, 1959) to ca. 170 mm at Alert (Bradley and England, 1978). The northernmost coastal fringe bordering the Arctic Ocean receives greater amounts; on the Ward Hunt ice shelf 178 mm fell as rain during the summer of 1954 (Christie, 1954). The mean annual snowfall at Alert is 1.2 m (Meteorological Branch, 1970).

Clements Markham Inlet can be subdivided into four climatic zones indicated by the relative amount of permanent snow cover, equilibrium line elevations, and glaciation levels. These zones are:

(1) the high-elevation ice caps and glaciers of the Grant Land Mountains surrounding the head of the Inlet; (2) the ice-free lowlands at the head of the Inlet; (3) the ice-free uplands of the outermost 30 km of the Inlet; and (4) the zone of low-elevation glaciers, snowfields, and ice shelves along the northernmost coast.

2.3.1 Ice Caps and Glaciers on the Grant Land Mountains

Most of the land above ca. 1100 m on northern Ellesmere Island supports an ice cover (Hattersley-Smith et al., 1969). Because the Grant Land Mountains and their ice caps rise above 2000 m a.s.l., colder temperatures may be expected there than in adjacent lowland areas. South of Clements Markham Inlet, Hattersley-Smith (1960) recorded a mean annual temperature of ca. -24°C at 1805 m a.s.l. (cf. Alert mean annual temperature; -17.8°C). At the same site snow pit studies indicated an annual accumulation of 12.8 g cm^{-2} which is ca. 2.5 times greater than the precipitation recorded at Hazen Camp on the N shore of Lake Hazen (Fig. 1.1). However, mean annual precipitation similar to the Hazen area was reported at lower elevations on the nearby Gilman Glacier (Lotz and Sagar, 1960). Hattersley-Smith (1960) estimated a mean annual precipitation of 150 mm above 1400 m a.s.l. on an ice cap, and 147 mm was recorded at Alert by the Meteorological Branch (1970). Further discussion on the glaciers and ice caps will follow in Section 2.3.5.

2.3.2 Clements Markham Inlet Lowlands

The ice-free lowlands at the head of the Inlet are characterized by a continental climate despite their coastal location. This is due to the combined effects of: summer insolation against the surrounding mountain flanks; the lack of open water within the Inlet throughout most summers due to stable sea ice cover; and lastly, the occurrence of intermittent chinook conditions that cause strong summer ablation (cf. Courtin and Labine, 1977). Nonetheless, cool air advected up the Inlet from the Arctic Ocean is common throughout the summer and moderates maximum temperatures. For example, during the 1980 season (June 16–August 6) the head of Clements Markham Inlet experienced a mean maximum temperature of +1.8°C and a mean minimum temperature of -0.5°C. On the other hand, the warmest temperature recorded was 13.5°C on July 25 when the head of the Inlet received strong chinook winds from Piper Pass (Fig. 1.1). Barry and Jackson (1969) also found that chinook conditions contributed to unusually high summer temperatures in Tanquary Fiord (80 km to the SW), the warmest of all high arctic weather stations (Bradley and England, 1978).

2.3.3 Ice-Free Uplands in the Outermost 30 km

Little is known about the climate of the ice-free uplands on the northernmost coast, however, Hattersley-Smith (1969) postulated that glaciers were absent because of low precipitation and the lack of large surface areas high enough for snow accumulation. Generally, this area can be regarded as a transition zone, too low in elevation to be glacierized, but too far inland to be affected by the increase in moisture associated with the Arctic Ocean.

2.3.4 The Northernmost Coast

The outermost coast of Clements Markham Inlet is characterized by low-elevation glaciers, snowfields, landfast sea ice, and ice shelves. Advection of cold air, coupled with increased precipitation from the Arctic Ocean, causes lower glaciation levels and equilibrium line elevations along the northern coast. For example, Marshall (1955) reported positive net annual accumulation on the surface of the Ward Hunt Ice Shelf during the mid 1950's. An important factor that reduces ablation in these areas is the high incidence of coastal fog and low stratus cloud. Coastal fog and cloud generated by the Arctic Ocean have also been considered responsible for reduced ablation and positive mass balance on the ice cap on Meighen Island (Paterson, 1969; Alt, 1979). Another factor which favours low-elevation glaciers in this area is the accumulation of wind-drifted snow along the outer coastline from the Arctic Ocean sea ice.

2.3.5 Height of Glaciation Level

An isopleth map of the glaciation levels (isoglacihypes) and equilibrium line altitudes for the Canadian High Arctic was drawn by

Miller *et al.*, (1973). Several regional papers have already discussed the importance of glaciation levels in the Arctic (e.g. Andrews, 1975; Andrews and Miller, 1972; Miller *et al.*, 1973) and therefore only a brief discussion will follow on the local conditions in Clements Markham Inlet.

Generally, the glaciation limit or the glacihype (GL) is thought to represent the long-term balance of winter accumulation and summer ablation modulated by topographic conditions. The limit indicates the threshold elevation required to sustain glaciers. Mountain summits below the GL are commonly ice-free, whereas summits above it support glaciers. The equilibrium line altitude (ELA), on the other hand, is the elevation on a glacier where the net annual mass balance equals zero, and may not coincide with the GL. Miller *et al.*, (1973) proposed that the difference in elevation between the ELA and GL is a function of the difference between local topoclimate and regional macroclimate respectively.

Miller *et al.*, (1973) show the broad trend of the GL and the ELA at a scale of $1:10^6$, and as may be expected, the lowest values are found along the coastlines and gradually rise toward the interior of landmasses as solid precipitation and cloudiness decrease (increasing ablation). Due to the sparse data points used by Miller *et al.*, (1973; commonly one per 2500 km^2), a GL map was constructed specifically for the Clements Markham Inlet area (Fig. 2.3). Construction followed the 'summit' method described by Andrews and Miller (1972).

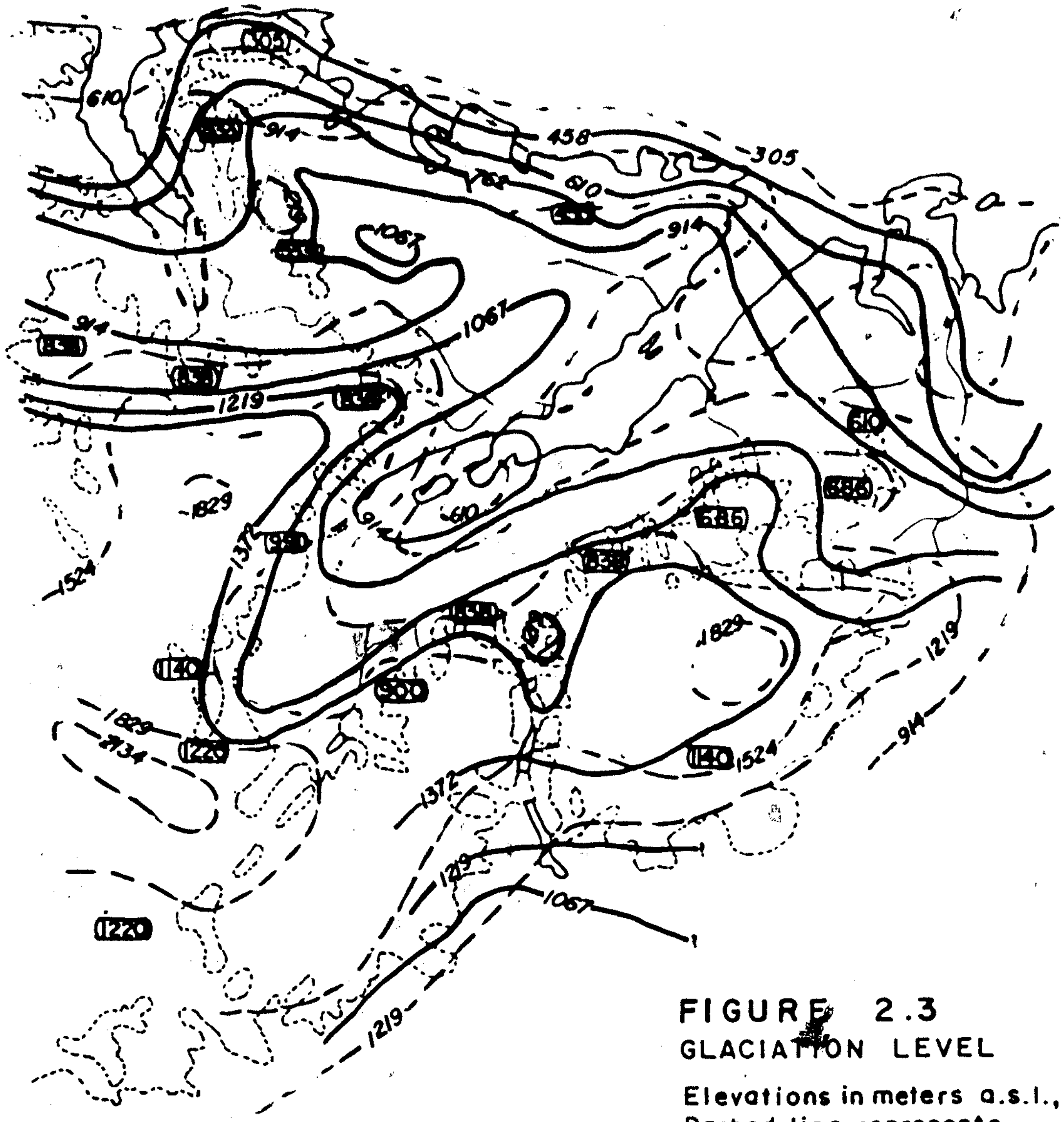


FIGURE 2.3

GLACIATION LEVEL

Elevations in meters a.s.l.,
Dashed line represents
generalized topography,
E.L.A.'s in brackets

The isoglacihypes for the Clements Markham Inlet region show a general correspondence with Miller's <u>et al.</u>, (1973) map. The lowest values (ca. 450 m a.s.l.) occur along the northernmost coast and, from there, they increase inland to >1300 m a.s.l. over the central Grant Land Mountains. Although the plot presented here was hampered by the lack of suitable data points in some areas, significant departures from Miller's <u>et al.</u>, (1973) map are noted. For example, noticeable inflections of the isoglacihypes, which are absent in Miller <u>et al.</u>, (1973), occur in each of the major re-entrants along the N coast (Markham Fiord, Clements Markham Inlet, and James Ross Bay). Furthermore, there is a depression of the GL centered over the head of Clements Markham Inlet. Decreasing values also occur on the S flank of the Grant Land Mountains (Fig. 2.3) which may indicate the effect of orographic precipitation in this area.

On the other hand, the reduction in the height of the GL in the major re-entrants is probably due to local topographic funnelling of moisture-bearing winds from the sea. Moreover, extensive low-lying areas, open to the sea, are subject to increased winter snow accumulation by wind drifting, and to decreased summer ablation due to coastal fogs. During the summers of 1979, 1980 and 1981, fog would commonly blow into Clements Markham Inlet, especially later in the season (August) when sea ice cover was most reduced.

As GL's are partially a function of topographic controls, isopleths of maximum elevation per 10 km^2 were also plotted in Figure 2.3. A rough correspondence between the GL surface and elevation seems to bear this out. The height of the GL in relation to local topography may be an important indication of the susceptability of an area to glacierization. Much of the outermost ice-free 30 km in the Clements Markham Inlet area is ca. 150 m below the GL. Nevertheless, it is important to note that the GL is time-transgressive and may not be in phase with the current climate (Miller et al., 1973). For example, a number of ice bodies in the outermost ice-free zone appear to be remnant glacier snouts. These ice bodies are generally found on valley floors immediately downvalley from vacant cirques. Hence, if these features originated from upvalley cirque glaciers, the former glaciers must have ablated within the cirques while their snouts survived in sheltered locations at lower elevations, perhaps with the benefit of snowdrifting and protecting fogs. Although these ice bodies probably could not have formed at their present locations, they are likely in quasi-equilibrium with the current environment and affect the pattern of the GL surface by lowering it. Furthermore, response lags of present glaciers will also make the GL surface deviate from present climatic conditions. For example, recent thinning of the ice caps in this area has led some workers to believe they are not in equilibrium with the present environment and are in a state of adjustment. However, the snouts of main outlet glaciers have remained stationary in recent times (Smith, 1961).

A number of ELA's given in Figure 2.3 were estimated using an accumulation area ratio of 0.65 or by the contour inflection method (Andrews, 1975). The values obtained agree with those presented by Miller et al., (1973) which are ca. 300 m a.s.l. over the northernmost coast and rise to ca. 1220 m a.s.l. in the Grant Land Mountains. The values estimated by the above methods give only long-term ELA trends. For example, although Hattersley-Smith (1960) measured the ELA on the Gilman Glacier near 1220 m a.s.l., which is in agreement with the above estimates, Marshall (1955) measured positive mass balance on the Ward Hunt Ice Shelf, which would mean the ELA was near sea level in those years and not at 300 m a.s.l. as indicated in Figure 2.3. Bradley and England (1978) reported that after 1963 the height of the July freezing level and ELA's in the High Arctic fell by 250-540 m. Furthermore, this study has shown that there must have been at least an equal, long-term depression of ELA's and GL's in Clements Markham Inlet during the last glaciation because the currently ice-free cirques at the outer coast were occupied at this time (Ch. IV).

2.3.6 Summary

The climate of Clements Markham Inlet can be characterized as cold, arid, and continental, although variability occurs depending on elevation and proximity to the Arctic Ocean. Cold polar conditions, with essentially no summer melting, prevail on the interior ice caps of the Grant Land Mountains, whereas a dry ice-free zone is found at the head of the Inlet where substantial ablation occurs. Conversely, the sea modulates temperature extremes at coastal sites which also receive more precipitation and wind-drifting. Moreover, less melting also occurs at these sites due to increased summer fog. The effectiveness of ablation in each of these zones appears to be the primary control for the trends in GL and ELA (Fig. 2.3), although local accumulation must also be important.

Chapter III

Methodology and Theory of Glaciomarine Stratigraphy

3.1 Introduction

The purpose of this chapter is twofold: to describe the field techniques used; and to review the geological and geomorphological features used for paleoenvironmental reconstructions in Clements Markham Inlet. The latter part includes a general facies model of glaciomarine sedimentation in order that basic stratigraphic interpretations may be made.

This study was concerned in part with reconstructing the sea level history in Clements Markham Inlet by tracing and dating former water planes. The former water planes (strandlines) are represented by beaches and delta terraces which occur at various elevations up to the maximum height of the marine inundation (marine limit). In reviewing the previous field work on northern Ellesmere Island (section 1.4) the relationship between glaciated coastlines and the history of relative sea level was introduced. Where former ice margins contacted the sea, marine transgression behind the margins could only occur following glacial retreat. Hence, the initial entry of the sea directly records the late glacial history of that site. Moreover, determining the history of postglacial transgressions and the nature of a landscape's subsequent emergence sheds light on the style of glacial recession and isostatic unloading.

3.1.1 Strandlines

Geologically, the formation of individual strandlines represents a short period of time, assuming normal postglacial emergence. Andrews (1970) noted that on average their formation takes from 50 to 300 years in the eastern Canadian Arctic. Elson (1967) found a mean age of 80 years for the development of strandlines on Glacial Lake Agassiz. Hence, on a time scale of several thousands of years, strandlines such as these provide good chronological markers.

An important consideration is the actual cause of strandline development. Generally strandlines originate when land and sea are stationary relative to one another. Nevertheless, once isostatic uplift is initiated it is normally monotonic, such that unless sea level is also rising rapidly, other causes for distinct strandline formation must be invoked. Several workers in Arctic areas have found that strandlines are usually associated with glacial stillstands (Løken, 1962; Sissons, 1963, 1967; Andrews, 1970). The evidence commonly cited is the termination of strandlines at moraines (cf. Andrews, 1970). In this respect the formation of certain strandlines is analogous to the formation of fluvial outwash terraces (valley trains) in more temperate environments. These terraces are normally considered to be the result of climatically-induced glacial advances, coupled with increased proglacial sedimentation. However, increasing evidence from polar desert environments, particularly Antarctica, suggests that significant proglacial sedimentation can only occur during glacial retreat, not advance (Rains et al., 1981). Furthermore, in the Canadian High Arctic, Bradley and England (1978) have shown that increased snowfall is essential for major glacial advances and, given the aridity of this environment, it is doubtful that such an advance could also be simultaneously coupled with increased runoff. Moreover, during the last glaciation on NE Ellesmere Island, England (1983) found that little sedimentation occurred in the full glacial sea during a long interval of a stable ice load, whereas during initial unloading (deglaciation) abundant sediments prograded into the sea.

Nevertheless, if a strandline forms due to a glacial response, be it an advance, a stationary position, or a subsequent retreat; the strandline becomes an important chronological marker. For example, if a strandline can be dated (by means of marine shells, driftwood, whale bones, etc...) and then traced to a given moraine, that moraine will be approximately contemporaneous with the strandline. Furthermore, in many arctic areas strandlines that mark the marine limit and terminate at an ice margin (i.e., an ice-contact delta) correspond with initial emergence, and hence, deglaciation of that particular site (Andrews, 1968). In addition, the elevation of the marine limit will also give some indication of the former ice load (assuming glacioisostatic equilibrium).

3.1.1.1 Marine Limits

To reiterate, the marine limit marks the "highest level reached by the sea on glacioisostatically depressed coasts" (Andrews, 1975, p. 181). In Clements Markham Inlet two types of marine limit were identified based on their areal extent and position. The first is the regional marine limit which marks the highest level attained by the sea along the whole Inlet. However, this overall marine limit did not register in many parts of the Inlet because of glaciers occupying the land which excluded the highest sea level stand (e.g. in valley mouths). For this reason a local marine limit was recognized in each of these areas which is lower and younger than the regional marine limit.

Several geomorphological criteria have been used to identify the marine limit in the field. These are: (1) the highest glaciomarine delta; (2) the uppermost beach ridge or beach shingle; (3) the highest occurrence of marine fossils or driftwood; (4) the uppermost wave-cut bench or notch; and (5) the lower limit of unaltered glacial debris, such as perched boulders, representing a washing limit. Of course, subsequent, lower water planes can also be identified with a number of the above criteria.

3.1.2 Peripheral Depression

As noted in the previous section ice-contact deltas often form during deglaciation and mark the marine limit. However, significant depression of the land also occurs beyond the margin of an ice sheet because the lithosphere is rigid. Such a peripheral depression has already been described in the ice-free corridor between NE Ellesmere Island and NW Greenland during the last glaciation (England, 1976a, 1981), and it also has been described or inferred in other arctic areas such as Baffin Island (Dyke, 1979) and Banks Island (Vincent, 1980). In such peripheral depressions (which may be up to 240 km wide; Dyke, 1979), the marine limit may not necessarily record deglaciation of the immediate area, but rather the sea level history may be more complex. Because the peripheral depression remained ice-free, the crustal loading, which was caused by the growth of the ice sheets to their limits, should be recorded in the peripheral zone by transgressive sediments. Furthermore, if isostatic equilibrium was achieved during that glaciation, the marine limit sea level would be maintained for a considerable length of time so that associated fossiliferous sediments would provide a wide range of ages. During glacial loading the transgression to the marine limit in the peripheral depression occurs at a much slower rate than it does inside the ice limit following glacial retreat. Consequently, transgressive littoral deposits, which are absent due to rapid inundation inside the ice limit, may be present in the zone of peripheral depression. However, there is the problem that sedimentation rates are likely to be very low in the arid, unglaciated areas of the peripheral depression (England, 1982). With such minimal sedimentation significant progradation may not occur, leaving a poor stratigraphic record of the initial transgression. In addition, the peripheral depression should also provide the total record of emergence initiated by the first thinning of the ice (cf. England, 1983). That is, it should provide not only the rebound since glacial recession, but also the "restrained" rebound component which occurs when the ice sheet initially thins before actual retreat occurs (cf. Andrews, 1970).

Lastly, geophysical modelling suggests that the bending of the lithosphere produces a 'forebulge' which lies along the edge of the peripheral depression (Walcott, 1970; Clark, et al., 1978). The forebulge is due to positive displacement of the lithosphere and is characterized by low postglacial emergence, or even submergence, as it collapses upon deglaciation. Understandably, coastlines in the vicinity of the forebulge would exhibit a more complex sea level history than those described above.

3.1.3 Relative Sea Level Curve

The primary method of depicting the sea level history at a particular site is by plotting a time/elevation graph known as the relative sea level curve. Inside the last ice limit, many researchers have shown that glacioisostatic emergence or uplift has a predictable logarithmic response to unloading (cf. Andrews, 1968; Walcott, 1970). In such cases 'emergence' curves are drawn during which initial emergence is very rapid (3-10 m/yrs.) but quickly decelerates towards the present (Andrews, 1968, 1970; Blake, 1975). Because this postglacial emergence represents a monotonic response through time, the construction of emergence curves has been used to provide ages for intervening emergent shorelines where datable material is lacking. Moreover, Andrews (1968) showed that 'predicted' emergence curves could be derived from only a single sea level point given its age and elevation. Such curves could, in turn, be used to estimate the age of any other undated sea level indicators in the area, including the marine limit. However, the widespread usefulness of such predicted emergence curves appears questionable given the greater variability of sea level history recently shown in Boothia Peninsula (Dyke, 1979) and NE Ellesmere Island (England, 1983).

3.1.3.1 Relative Sea Level Curve Construction

Constructing any relative sea level curve involves proper identification of the sea level indicators, accurate measurement of their elevation, and correctly dating them.

Identification of former water planes based on morphostratigraphic grounds is a problem encountered in the field. On some of the shoreline features mentioned in section 3.1.1 one cannot pinpoint a former sea level and therefore the selected elevation can be somewhat subjective. It is important to bracket sea level accurately. For example, it used to be common to place sea level right at, or somewhere above, a given shell sample elevation. However, it is apparent that marine bivalves could have lived anywhere from the intertidal zone to 165 m below sea level (i.e. the species range for Hiatella arctica and Mya truncata (Wagner, 1969)). This stratigraphic problem is well illustrated by the anomalously steep emergence curves of Hattersley-Smith and Long (1966), and Lyons and Mielke (1973), and discussed by England (1976a).

A better method of pinpointing former sea levels is based on dating specific delta terraces. In such cases the stratigraphic interpretation is critical, depending upon where the datable material was collected. Ideally, a sample can be obtained from Gilbert-type foreset beds which can be traced upwards to contemporaneous topset beds. This point of intersection is commonly interpreted as the mean elevation of local sea level at that time. Nevertheless, the stratigraphy of the delta is not always straightforward because well developed topset, foreset, and bottomset sequences may be absent, making sea level interpretations difficult. Moreover, it is common to take the outermost edge of a delta terrace as the sea level to which progradation occurred, however, subsequent backwasting of the scarp can give erroneously high estimates. The delta lip is also usually graded to low tide, whereas the measuring datum is usually high tide. Furthermore, wave erosion or fluvial abrasion may cut into pre-existing marine deposits and the resulting dates will relate to higher and older sea levels. Solifluction terraces, fluvial terraces, or kame terraces may also be misinterpreted as marine delta terraces. In summary, a large delta system may exhibit all of the above factors such that considerable stratigraphic work must be done before datable samples can be obtained and related to specific former sea levels. The general model of glaciomarine facies described in section 3.2 further aids such interpretations.

Driftwood embedded in beach gravels also provides reliable control points in the construction of an emergence curve (cf. Blake, 1975). However, several factors may introduce errors. For example, driftwood can always be redeposited from its original point of stranding to lower elevations, particularly where beach-forming processes are weak. Conversely, errors may also be introduced when sea ice pushed ridges or storm beaches deposit driftwood several meters above mean high tide. In addition to the positioning errors, the radiometric age of the wood can also be older than the sea level involved, depending on its history of residence in the sea. However, this is probably not a major problem because the mean residence time of driftwood in the Arctic Ocean is generally 5-20 years (Häggblom, 1982; Stewart and England, 1983).

Once a sea level indicator is identified it is necessary to measure it accurately relative to present sea level. Generally an altimeter is used for reconnaissance work, however, because altimeter readings are pressure and temperature dependent, the accuracy of any given measurement is often limited to 5% in elevation. However, when multiple readings can be made during periods of stable atmospheric pressure (common in the High Arctic) the error in the altimetry is often much less than 5% when checked with levelled elevations. Nonetheless, a surveying level provides the most confident results with an accuracy of ca. 1% error on transects of >1 km. A second source of error occurs in the choice of a common datum from which to measure. Generally mean high tide is used, but sea ice pushed ridges or land fast sea ice may interfere with determining its exact location. Hence, when open water is unavailable, the datum must be the surface of the sea ice, which is actually above sea level due to the freeboard. Nonetheless, because the tidal range in Clements Markham Inlet is ca. 30 cm, the error due to tides is likely to be no more than ca. 1 m.

The final factor for obtaining a reliable sea level curve concerns the actual radiocarbon dates on the samples. Although well preserved samples, with little chance of contamination, are sought they are often not possible to obtain in the field. Furthermore, a problem over which the field worker has little control is the 'laboratory dating error' on a given sample. For example, radiocarbon dates always have a standard error which is usually in the order of a few hundred years for a sample of early Holocene age. Although this imprecision can be graphically portrayed for each sample, this age range must always remain a source eof imprecision.

In summary, it is evident that delimiting former sea levels and obtaining datable materials can be problematic. Therefore, several lines of evidence are required to pinpoint a former sea level. Moreover, because postglacial relative sea level curves show rapid emergence following deglaciation it is apparent that higher relative sea levels formed during the early period must be accurately dated, whereas, lower relative sea levels must be pinpointed in elevation (as emergence is less rapid). It is therefore important to date a marine limit as accurately as possible, otherwise large temporal errors would be introduced when projecting an emergence curve back in time from dated, lower relative sea levels (cf. Andrews, 1970).

3.1.4 Equidistant Diagram

Equidistant diagrams provide a second method of depicting former sea levels, but on a regional, versus a local scale. In this case the elevations of different strandlines are plotted against distance so that their spatial distribution may be compared. By this method individual strandline remnants within Clements Markham Inlet may be correlated. Normally, it is rare for any shoreline to be traced for more than 1-2 km, hence the equidistant diagram contains isolated remnants of former water planes which are connected on the basis of their common age. The remnant shorelines are best correlated through the use of site-specific, radiocarbon dated, relative sea level curves. However, because curves are not available for all sites, a less reliable method of 'fitting' water planes to clusters of points can be employed. In a fiord situation the least squares method has been commonly used (cf. Andrews, 1970). Of course, any fitting of waterplanes, assumes that emergence was continuous and predictable and that postglacial faulting did not differentially displace sea level indicators (see Chapter II).

Equidistant diagrams should be oriented orthogonally to the pattern of the isobases (to be discussed in the next section). However, if the pattern is not initially known, as in Clements Markham Inlet, the orientation of the abscissa must be adjusted so that minimum variance between strandline features occurs (Andrews, 1970).

Once the proper orientation of the diagram is established, its contribution to analyzing crustal deformation and the pattern of isobases is significant. Equidistant diagrams may also indicate whether or not a particular strandline is incised into the face of an o[illegible] marine deposit (e.g., a regraded delta terrace).

Nonetheless, Andrews (1970) has pointed out several problems pertaining to the use of equidistant diagrams. The first is: what is the actual shape of the former strandline, tilted or warped? Although Broecker (1966) has suggested an exponential warping based on geophysical parameters, on a scale of ca. 100 km the curvature is so small that is is probably not unrealistic to assume a rectilinear form given the observational errors in the field. A second problem involves the circular input required in regression analysis because only those points assumed to fall on a particular water plane are used. This problem was not encountered in this study because the strandlines were fitted by age and not by regression. A third problem discussed by Andrews (1970) is that the orientation of the projection plane is dependent on the spatial distribution of the points themselves (i.e. points usually parallel a coastline).

3.1.5 Isobases

Isobases are isopleths which join points of similar emergence over the same period of time. The derived emergence curves and equidistant diagrams portray the pattern of shoreline displacement upon which the isobases are drawn. These, in turn, serve to spatially integrate differential emergence over a larger area thereby indicating areas of maximum uplift.

In the case of Clements Markham Inlet, and the northern coast of Ellesmere Island in general, it is reasonable to assume that the isobase pattern should reflect a dominant ice load in the Grant Land Mountains to the SW. Furthermore, because Clements Markham Inlet is greater than 200 km beyond the limits of the present Greenland Ice Sheet, this pattern should be independent of the load imposed by the Greenland Ice Sheet which has strongly influenced NE Ellesmere Island, SW of the Grant Land Mountains (England, 1976a, 1982). It therefore follows that the timing of initial emergence on the northernmost coast should be entirely controlled by the history of the local Ellesmere Island ice load, and not by the Greenland Ice Sheet which was close enough to control the initiation of emergence on NE Ellesmere Island (England, 1976a, 1983).

3.1.6 Glacier Profiles

In order to document the ice loads responsible for the glacioisostatic adjustments in Clements Markham Inlet, field work also involved delimiting former ice margins. The objective was to connect former glacier profiles in the Inlet to discrete periods of glacioisostatic adjustments.

In addition to identifying end moraines, three other sources of evidence can be used to reconstruct glacier margins. First, because subpolar glaciers have englacial temperatures well below $0^{\circ}C$, they have well developed marginal and submarginal drainage systems. Such meltwater channels found along valley sides indicate stillstands of the glacier formerly occupying the valley. This method was used successfully to produce a pattern of ice retreat in Clements Markham Inlet (see Chapter IV). The second method uses the distribution of erratics to delimit the former extent of ice. However, due to the lack of distinct erratics in Clements Markham Inlet, this was limited to only a few isolated cases (discussed in Chapter IV). A further complication, which also pertains to the first method, is that no age differentiation is apparent between one or more glacial advances. The third method attempts to overcome this temporal constraint by defining various weathering zones within the Inlet. Presumably each successive glaciation rejuvenates the landscape such that the least subaerially weathered surface represents the most recent glaciation, whereas the most weathered surface indicates the longest interval of subaerial exposure, hence the oldest glaciation. Moreover, some of the most weathered surfaces on NE Ellesmere Island do not show any evidence of ever having been glaciated (England <u>et al.</u>, 1981). However, no rejuvenation may occur if the ice was cold-based, except that meltwater channels would be cut (cf. Dyke, 1983).

Unfortunately, most observations on relative surface weathering in Clements Markham Inlet have not indicated any obvious weathering breaks, other than a few restricted cases (see Chapter IV). Sedimentary rocks do not present any practical measure of weathering (Birkeland, pers. comm., 1980), and if differential weathering was not obvious in the field, little is gained from time-consuming transects in order to find a difference (Dyke, pers. comm., 1980). Nevertheless the fluvial landscape which occurs on the outer coast of Clements Markham Inlet may indicate a long interval since deglaciation. Furthermore, England et al., (1981) identified three distinct weathering zones along western Kennedy Channel, NE Ellesmere Island which they attributed to multiple glaciations.

3.2 Glaciomarine Sedimentation in an Arctic Fiord Basin

3.2.1 Introduction

This section outlines the possible factors affecting glaciomarine sedimentary sequences in a High Arctic fiord basin during a single glacial cycle of ice advance and retreat. This model is used to anticipate the kinds of stratigraphic sequences which may be found along the coastline of Clements Markham Inlet.

In order to understand the character and sequence of the sedimentary facies to be found, the main controls must be determined. Two main controls are identified. The first control is the type and volume of sediment supplied to the coastline. When considering a glacial cycle the majority of the sediment input would be glacigenic, and therefore, it would depend on not only the type of material available for glacial erosion and transport, but also the character and extent of the sediment-supplying glaciers. The proximity of the glaciers to the sea would, in turn, control the mode of deposition of that material. For example, if the ice margin is inland from the coast, its sediment could reach the Inlet only by fluvial or eolian transport, whereas, if the glacier is grounded below sea level its sediments could be deposited directly into the water. The second major control on the facies sequences would be the relative movements of the land and sea. The rates and amplitudes of these fluctuations, coupled with their interaction with sediment input during the glacial cycle, will have important effects on the stratigraphy and sedimentology. In addition, the character of postglacial eustatic and isostatic movements will not only determine the facies, but also the amount of sediment which will be exposed above present sea level. Only the sediments exposed above present sea level can be easily studied by the field worker. These sediments, in turn, may be discontinuous and may contain numerous unconformities. An examination of the interaction between the two controls and how they may result in various stratigraphic sequences will be attempted here.

3.2.2 Sediment Control–Glaciomarine Sedimentation

As noted above, the type of sediment input and mode of deposition at the fiord head interacts with the relative movements of land and sea which will determine the stratigraphy. Because the major sediment input in Clements Markham Inlet was from former glaciers in contact with the sea, emphasis is placed on glaciomarine sedimentation.

Several authors have developed models of glaciomarine sedimentation. These models fall into two general categories: those concerned with large scale conditions beneath Antarctic ice shelves (e.g. Cary and Ahmad, 1961; Reading and Walker, 1966; Drewry and Cooper, 1981); and secondly, those concerned with fiord glaciers in the Northern Hemisphere (e.g. Boulton, unpubl.; Nelson, 1978, 1981; Powell, 1981). The former models recognize three major depositional zones: an ice-contact zone below and adjacent to a grounded glacier; a floating ice shelf zone; and lastly, an iceberg zone. Differential deposition between cold and warm-based ice conditions is also discussed. On the other hand, the latter fiord models involve a smaller scale and temperate ice conditions. For example, the model proposed by Nelson (1978, 1981) for eastern Baffin Island is based in part on Boulton's (unpubl.) model based on Spitsbergen fiord glaciers. These glaciers have extremely high sedimentation rates and no ice shelves. Powell's (1981) model is also based on very active maritime glaciers on the Alaskan coast. Although the glaciers on northern Ellesmere Island are subpolar and may be associated with ice shelves, it is thought that the model proposed by Nelson (1978) may be generally applicable to the glaciomarine facies within Clements Markham Inlet. However, it is apparent that the sedimentation rates for subpolar glaciers would be much lower.

Nelson (1978) recognized three depositional zones that grade from one another with increasing distance from the ice margin. The zone adjacent to the ice margin is termed the Proximal Glaciomarine Zone and is characterized by very high sedimentation rates and actively calving ice. The Proximal Zone is further subdivided into the inner and outer subzones. The inner-proximal subzone is characterized by both tills and fan deposits. Where the ice is grounded, normal terrestrial lodgement and meltout tills are deposited. Adjacent to the margin, ice-contact fans of moderately, to well sorted sands and gravels are deposited from issuing meltwater, particularly near the sides of the fiord. These fans may be submarine or partially above sea level along the sides of the fiord. With time the fans may coalesce to form gravelly bars. Moreover, the fans can be interbedded with poorly sorted, subaqueous flow tills (cf. Evenson et al., 1976), or other material rolling and slumping down the fans. The characteristic sedimentary sequences in the inner-proximal subzone would be cyclic deposits of all grain sizes which would be impossible to correlate, even over short distances.

The outer-proximal subzone is characterized by massive deposits of silts and fine sands with relatively few clasts. High current velocities carry most of the silt and clay away from the ice margin to this zone. However, icebergs are generally carried beyond the outer-proximal subzone before they begin to deposit their debris. Nonetheless, shoals may ground some icebergs, and consequently, they may become mantled by iceberg-derived diamictons. The icebergs may also scour and deform the underlying sediments. The outer-proximal subzone is also subject to turbidites and strong currents originating from the inner-proximal zone which winnow, channel, and grade the deposits. Generally, microfauna are very sparse in this zone because of high sedimentation rates and fast-changing environments.

The Intermediate Zone is distal to the Proximal Zone and occurs approximately 1-2 km from the ice margin. This zone is characterized by diamicton deposition from melting icebergs and by the settling out from suspension of fine silts and clays. Distal turbidite deposits are also found in this zone, but unstratified to poorly-stratified mixtures of all grain sizes are also common. Generally this zone contains more uniform sediment bodies with more gradation between contacts, and microfauna are common but not abundant.

Lastly, the Distal Zone, beginning 4-6 km from the ice margin, is characterized by massive, shelf type, sandy silts and clays of uniform composition and thickness. However, occasional clasts may still be found from ice rafting. In Spitsbergen, the sedimentation rate is approximately 1000 times less in the Distal Zone than it is in the Proximal Zone, and consequently microfauna are abundant.

A glaciomarine model such as that of Nelson (1978), effectively describes the possible sequences of facies found in a glaciomarine environment. However, such a model is essentially static and does not describe the sedimentary sequence as a fiord glacier advances or retreats, or as sea level lowers or rises. Do the depositional zones merely migrate down or up the fiord as the glacier advances or retreats, or is the process more complicated? In order to have a better understanding of this, the related glacioisostatic and eustatic fluctuations which balance the input of glaciomarine sedimentation must be considered.

3.2.3 Eustatic and Glacioisostatic Control

Clark et al., (1978) estimated the magnitude and character of eustatic and isostatic movements which occurred in Arctic areas following deglaciation at ca. 16 ka BP. Clements Markham Inlet may fall into two possible zones proposed by Clark et al., (1978). The existence of these zones depends on the extent of the last glaciation and hence the magnitude of the ice load which depressed this area. If the last glaciation extended well out into Clements Markham Inlet, the fiord head would fall into Zone I. Zone I is characterized by continuous and ongoing postglacial emergence because glacioisostatic unloading exceeded the postglacial sea level rise. If the last glaciation did not inundate the fiord, the fiord would fall into the Zone I/II transition. This zone is characterized by initial postglacial emergence, followed by submergence caused by the collapse of the forebulge. Although, this geophysical model (Clark et al., 1978) is theoretical and based on the instantaneous uniform melting of the ice sheet, it does give an approximation of the types of responses which will likely occur given different glacial histories. The results of this model are also paralleled by recent geophysical solutions for different glacial histories in Atlantic Canada (i.e. Quinlan and Beaumont, 1981). It is concluded here that the relative movements of land and sea will depend on the extent of ice during the last glaciation and this, in turn, will dictate the depositional sequence.

3.2.4 Analysis of the Stratigraphic Sequences

As noted, facies changes are largely controlled by the relative movements of land and sea coupled with the influx of sediment. Several workers have investigated these relationships by looking at onlap–offlap sequences found in older sediments, especially the Cretaceous marine transgressions in the interior of North America (e.g. Lane, 1963; Curray, 1964; Franks, 1980). Andrews (1978) suggests that, although the origins of these transgressions and regressions is different, these basic stratigraphic models can be applied to glacially-induced, Arctic sea level fluctuations. Such an application will be attempted here.

Curray (1964) sought to classify ancient marine transgressions and regressions by analyzing the balance between sediment input versus sea level rise or fall. His model (Fig. 3.1) illustrates the possible conditions leading to transgressions and regressions, and is adapted here in order to describe the different sediments that may accompany sea level changes caused by a glacial cycle.

Table 3.1 indicates the possible sequence which Curray (1964) classifications may follow during a glacial cycle. In this case, the cycle begins with ice occupying the fiord. It is assumed that any preceding glaciomarine sediments are overlain by till deposited by this advance. As the ice retreats, submergence occurs within Zone I of Clark et al., (1978). Because the land is depressed well below sea

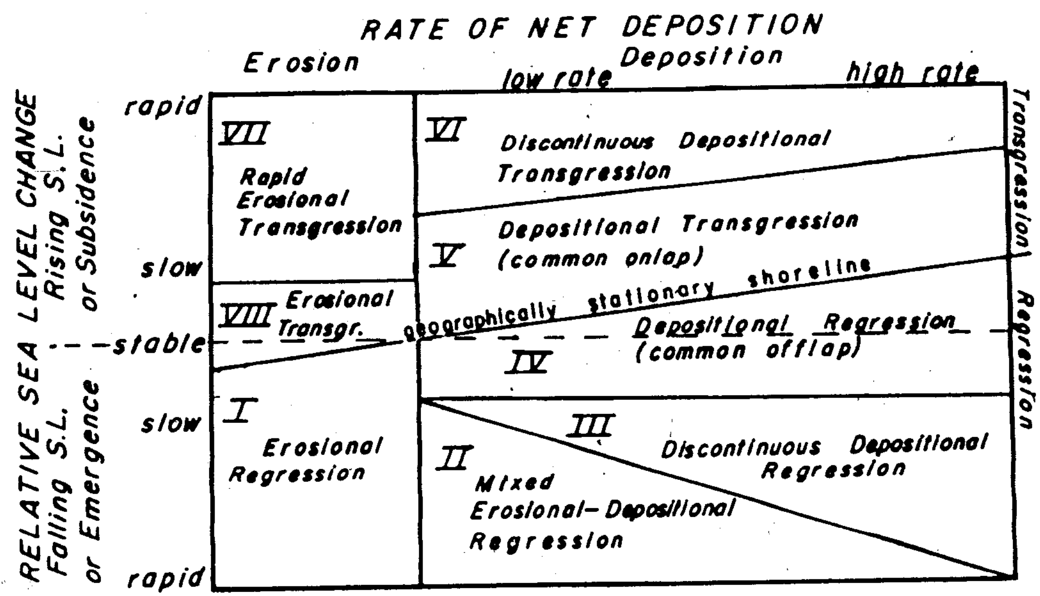


FIGURE 3.1 A classification of transgressions and regressions (after Curray, 1964).

TABLE 3.1

Classifying a transgression and regression that occurs during a glacial cycle.

	Classification (Curray, 1964)	Glacial Cycle	Conditions	Results
	-	Glacial	Ice occupying fiord	Glacigenic deposition
	VII	Deglaciation (depression below) sea level	Transgression >> Sedimentation	overstep marine over glacial deposits
	VI*		Transgression > Sedimentation	thin veneer of littoral sands, overlain by discontinuous marine sediments
		decreasing ice load		
	V* (approaching marine limit)		Transgression $\geq$ sediment supply	marine onlap
	VIII* (or short duration possible)	Uplift = Eustatic sea level rise	geographically stable coast	marine limit reached
	I	sea level rise < uplift	Emergence >> Sedimentation	no beaches, dissection of uplifted marine sediments
Isostatic Zone I	II*		Emergence > Sedimentation	mostly wave cut beaches, regressive strandline
	III*			
		decreasing rebound		
	IV*		Emergence $\geq$ Sedimentation	marine off lap, regressive beaches, delta building
Isostatic zone I/II	VIII		net erosion from wave action, local subsidence from compaction	
	V	Collapse of the forebulge	Transgression	marine on lap

**shorelines may be preserved

level a rapid transgression takes place simultaneously with the retreating ice margin. If the rate of transgression greatly exceeds the rate of sedimentation, normal facies changes associated with marine onlap will not occur (case VII, Fig. 3.1). Rather, the glacigenic deposits (i.e. till) will be overlain immediately by deepwater marine deposits, and lateral facies changes will be negligible. The unconformity formed between the deeper water sediments and the underlying till is termed "overstep" (Dunbar and Rodgers, 1957).

In many fiords and valleys the above condition existed until the glaciers receded above the marine limit. However, in proglacial areas along valley sides where the sea was in contact with the land, the rate of sedimentation may increase relative to the rate of transgression (case VI to V, Fig. 3.1). If both of these rates remain high, the material supplied could have been redistributed parallel to the shoreline via waves and currents. In Clements Markham Inlet, however, open water is limited by landfast sea ice, and thus, wave action is weak and limited to the short summer season. Given some littoral processes, the coarsest material would be found near shore, fining seaward. As this type of transgression proceeds the littoral facies would shift landward resulting in a "common marine onlap" sequence recorded in the sediments. Nonetheless, the onlap sequence may be complicated if sediment is temporarily impounded within sandar and deltas during the transgression and therefore unavailable for littoral processes. In this case the water may rise over the terrestrial deposits, without the graded deposition associated with common marine onlap (Mathews, 1974).

During the establishment of the marine limit the rate of transgression decreases until it is balanced, and then overcome, by isostatic rebound due to the removal of the ice burden. During this time, there may be a brief interval when the relative sea level is stable at the marine limit (case VIII, Fig. 3.1). Moreover, a brief transgression may follow due to net erosion by wave action or local subsidence due to compaction of sediments. However, it has been demonstrated that postglacial emergence in the Arctic, within the glacial limits, is initially very rapid and then logarithmically decreases to the present (cf. Andrews, 1968; Blake, 1975). These conditions of initial emergence would be represented by case I (Fig. 3.1). Here, the initial rate of emergence would be very rapid relative to the rate of sedimentation. Under these conditions the expected seaward shift of coarse terrigenous facies would be discontinuous. Following emergence, areas at the head of the fiord would be mantled by fine marine sediments which lack a cover of the coarser offlap facies. These exposed marine sediments would be subject to gullying and deflation, moreover, no littoral facies marking past strandlines would be present (e.g. Blake, 1975). A cover of littoral sediments may also be absent in special circumstances where there is no source of coarse matereial (eg., till) for littoral processes to act on.

As the rate of emergence decreases through time, the rate of sedimentation will eventually balance the rate of emergence.

Initially only wave-cut beaches would form (case II, Fig. 3.1), however, they would soon be succeeded by depositional beaches as a normal seaward progression of coarse terrigenous facies occurred (marine offlap, cases III to IV, Fig. 3.1). Furthermore, if the supply of sediment remained high a depositional regression would occur because of rapid progradation by deltas. Although the most pronounced phases of delta building would be expected during times when the rate of emergence approaches zero, in arctic areas these periods are generally coupled with greatly reduced sediment supply due to diminished glaciers. At the head of Clements Markham Inlet case IV represents the contemporary stage in the postglacial marine sequence.

If the zone I/II transition of Clark et al. (1978) is encountered within the fiord, a condition of limited submergence may occur (case VIII). Although case VIII submergence in Curray's model is due to the compaction of sediments or net wave erosion, this is not the case in Clark et al.'s, (1978) model where transgressive, marine onlap sequences may also occur (case V) if the forebulge collapses, as predicted by the geophysical model. As noted earlier, the Zone I/II transition may have occurred within Clements Markham Inlet. If this was the case transgressive sequences would be expected in the outermost areas of the Inlet. However, no evidence of this was found (see Chapter IV and V).

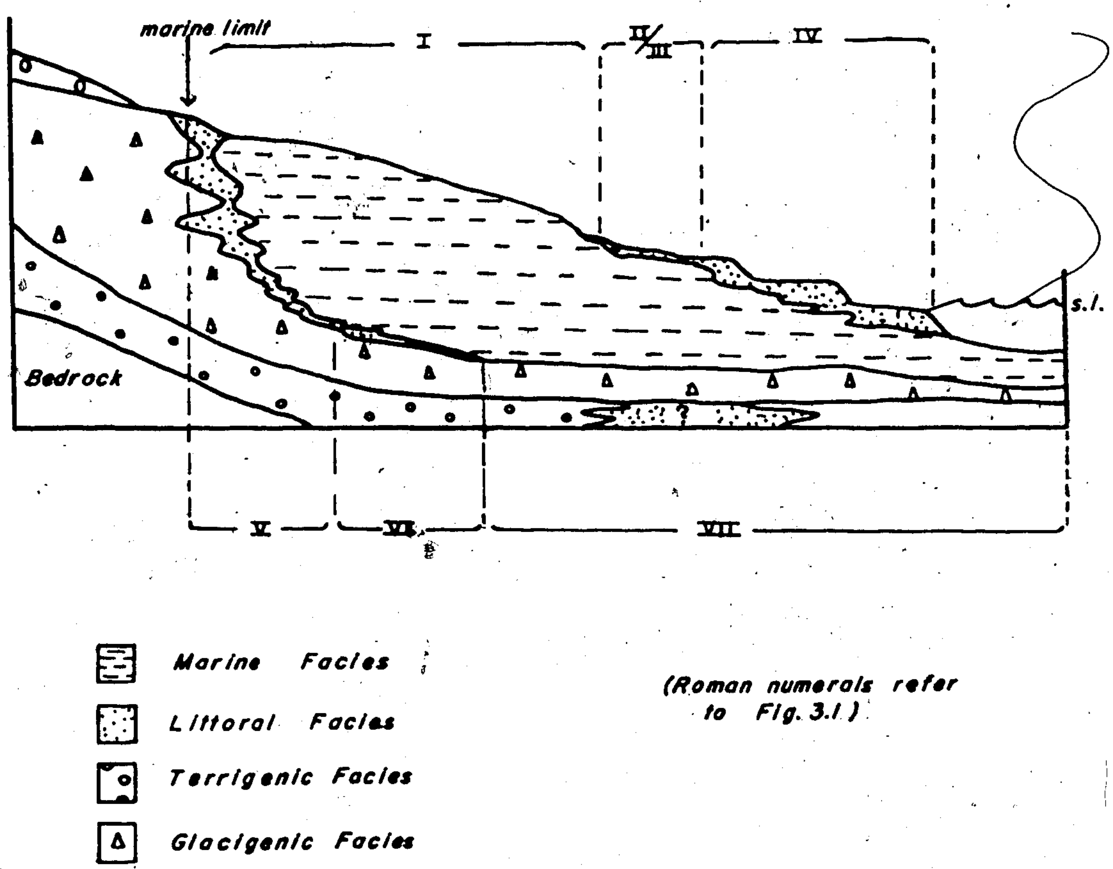
3.2.5 Discussion and Conclusions

The stratigraphic sequences expected in an Arctic fiord during a glacial cycle are shown in Figure 3.2. This diagram attempts to describe the succession of deposits when considering the fiord as a whole. However, it is evident that some deposits may contain complex assemblages of grain size, sorting, and sedimentary structures that cannot be readily accommodated by such a large scale model. In general, Figure 3.2 disregards the variability in sediment supply from one area to the next, as well as temporal fluctuations in sediment output from the glaciers. Another limitation to the model is that it only considers a single glacial cycle. Should there be a more complex glacial history (i.e. multiple glaciations), the isostatic and eustatic responses would also be more complex.

Lastly, as pointed out by Andrews (1978), there is an intrinsic problem of coarsening or fining-upward sequences in glacial fiords. For example, the nature of the sediment will change not only with the depth of water, but also with its proximity to a glacier. At least three conditions can produce a coarsening-upward sequence: (1) an approaching glacier, supplying increasingly coarse sediments; (2) a falling sea level; or (3) a rising sea level with very high sediment input (Mathews, 1974). However, the last case must be rare in a glacial environment.

Despite the above limitations, it is hoped that by outlining the major factors controlling fiord sedimentation during a glacial cycle some insights can be provided on the past depositional environments of Clements Markham Inlet. An important benefit of this model is that it anticipates complications which may be encountered in

FIGURE 3.2 A schematic section along an arctic fiord following a glacial cycle.



the field as well as providing an explanation of the spatial distribution of glaciomarine facies. For example, it was noted that glaciomarine sediments can be carried long distances out the fiord as well as be deposited considerable distances upvalley during glacial retreat. Therefore, recognizing specific facies becomes important (e.g. esker sands versus littoral sands). In addition, lateral gradation of facies and interlayering facies may be expected in specific areas of the fiord. For instance, one would not expect to find evidence of former strandlines in areas of extensively gullied marine silts. The surface silts themselves indicate an extremely rapid regression. Furthermore, if certain stratigraphic relationships can be recognized, such as an overstep sequence, the model can provide constraints on the conditions that prevailed (e.g. fast or slow glacial retreat), and aid further stratigraphic interpretations of the fiord deposits.

Chapter IV

Geomorphic Data

4.1 Introduction

This chapter is divided into three principal sections. The first section (4.2), deals with a map of the surficial deposits in the study area. The spatial distribution of the deposits is described and briefly analysed. The second section (4.3), describes the geomorphology and stratigraphy of specific sectors in Clements Markham Inlet (Fig. 4.1). These areas are delimited by major watersheds and commonly contain deltas on their seaward margins. All pertinent radiocarbon dates are included in this section. The final section (4.4), deals with past glacier configurations within the study area. Ice limits and ice retreat patterns are provided based on information from the two preceding sections.

4.2 Surficial Materials Map

As a base for studying the geomorphology of the area, a surficial materials map was compiled from airphotos and ground-truthing (Fig. 4.2, in the back cover). The legend (Table 4.1, back cover) describes the materials identified, giving their symbols and an explanation of geomorphic descriptors used.

The area mapped covers the terrain between the existing

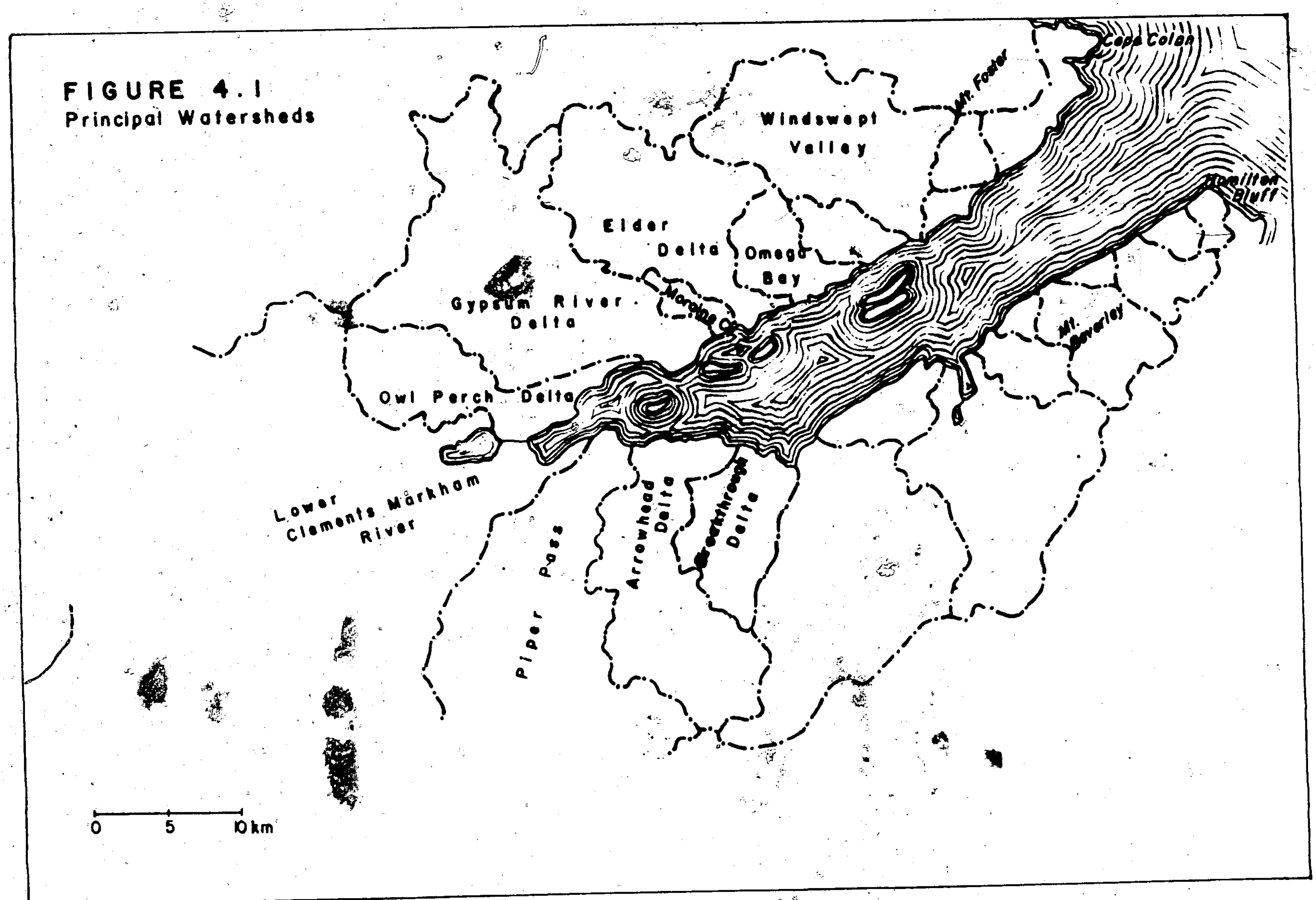


FIGURE 4.1
Principal Watersheds

glaciers and the head of the Inlet, as well as the ice-free uplands down the Inlet. Full airphoto interpretation was hampered in some narrow valleys where strong shadows occurred. The ice-free uplands were also difficult to map because of snow cover. This was particularly so when trying to distinguish till veneers from rock rubble, hence questionable areas were left blank. The following is a description of the surficial materials.

4.2.1 Bedrock

The predominant surface in the area is bedrock of various ages composed primarily of sedimentary lithologies (see Ch. II). The bedrock is found in various stages of decomposition from extensively frost shattered rubble, mantling broad summits and hill slopes; to freshly exhumed surfaces.

Although localized weathering breaks were found on similar rock types, no spatially coherent patterns could be mapped. Several factors govern the extent of rock weathering including: the length of exposure, susceptibility, and microenvironment. The variability presented by the latter two factors is thought to have obscured recognizable glacigenic weathering zones.

4.2.2 Till

Non-sorted glacial debris is widespread in the map area. It is commonly found as a thin, discontinuous veneer overlying the bedrock. Slightly thicker veneers (still <1 m thick) are found in valley bottoms and cirque floors. Till surfaces are characterized by varied clast sizes within a predominantly sandy matrix. For the most part the material is locally derived, but an abundance of erratic debris distinguishes it from _in situ_ weathered bedrock. Nonetheless, even predominantly weathered bedrock surfaces contain isolated erratics. Furthermore, it is often impossible to distinguish till from frost shattered bedrock, especially in outermost Clements Markham Inlet where recent glaciation was very localized, limiting erratic dispersal. Till blankets (>1 m) rarely occur and likely represent concentrations near former glacier margins.

4.2.3 Glaciofluvial Sediments

The major source of surface runoff in the study area is from glacier melt, hence most of the waterlain sediments are glaciofluvial. Glaciofluvial sediments on the surficial map cover a wide range of genetic types. Generally, these sediments are gravel and sand deposited beneath, and in front of, the marginal zone of a glacier, forming valley trains (sandar) which terminate at the coast line.

Identifying sandar within valleys is straightforward, nonetheless several major qualifications are necessary. First, because many of the sandar terminate at the sea, they form the coarse topset sediments of prograding delta systems. Because postglacial emergence characterizes this area large lateral shifts in facies often occur; for example, remnant, deltaic deposits may be found within contemporary sandar environments far inland. Hence, on airphotos, confusion may result when trying to separate coarse, marine delta deposits from sandur deposits, particularly, where the terraces are found far inland, ergo the two units are lumped together. A second important qualification is that the contemporary glaciers in this area are subpolar (see Ch. II) and drainage channels along the ice margins are common, furthermore, the existence of many old, ice-marginal meltwater channels downvalley suggests that their thermal regime was similar in the past. Hence, many outwash terraces are kame terraces delimiting former ice margins.

The origins of the various types of glaciofluvial terraces are crucial to the interpretation of relative sea levels. Sandar remnants depict former surfaces which grade to specific relative sea levels and they commonly terminate at delta terraces constructed into that sea level. Clearly, kame terraces, whose downslope sides were buttressed by glaciers, bear no such relationship to sea level. Differentiation is most important when dealing with ice contact deltas, where kame and delta terraces may be found in close proximity, especially where parts of the glacier terminus contacted the sea.

4.2.4 Glaciomarine Sediments

Fine grained, raised marine deposits form the most widespread surficial unit in the area other than rock. These deposits are predominantly horizontally stratified silt and fine sand, and comprise the bottomset beds of several delta systems at the head of the Inlet.

Although numerous broad valley bottoms are in fact below the marine limit for considerable distances inland, they lack fine marine deposits due to subsequent river erosion. Nonetheless, some flat-lying areas within a few kilometres of the Inlet are mantled by thick silts which are extensively gullied. The silts contain varying amounts of dropstones which form a surface lag in some areas.

4.2.5 Minor Materials

A number of surficial units mapped are very local in extent. The most common of these are alluvial fan and colluvial deposits mantling the valley slopes. Solifluction and debris flows appear to be the dominant processes transporting the material from weathered rock upslope. As noted, the tills in the area are often difficult to distinguish from the weathered rock, hence the glacigenic contribution to these deposits cannot be discerned. Furthermore, local talus production itself contributes substantially to lateral moraine production in alpine areas. Further modification may also take place should the valley-side material become rock glacierized. A number of areas designated on the map are thought to be rock glacierized lateral moraines (cf. England, 1978).

4.2.6 Discussion

The Surficial Map shows a complex association of moraines, and marine and glaciofluvial deposits within the valleys and along the shores of Clements Markham Inlet. Because most main valleys experienced glaciation, followed by a marine transgression and post-glacial emergence, the spatial organization of deposits within individual valleys follows a similar pattern. The glaciers provided a major source of runoff and sediment, hence large delta complexes are found at the mouths of formerly glaciated valleys. Furthermore the mouths of the valleys usually contain kames, moraines, and flat-topped glaciofluvial terraces suggesting glacial stillstands during these higher sea levels. The marine limit is usually marked by uppermost terraces which prograded from these ice-contact deposits. Each delta also usually contains a series of descending delta terraces which lie seaward of the marine limit terrace. These delta terraces may have corresponding sandar terraces upvalley. Occasional foreset bedding depicts the progradation of the terrace over deeper-water sediments, however, the terraces are also frequently inset into the underlying material with a cut and fill relationship.

Deep water, bottomset sediments often lie in the outer perimeter of each delta. These deposits are composed of horizontally stratified silts and fine sands, often 5-10 m thick. The sediments are locally fossiliferous and may contain a preponderance of ice-rafted material.

Upvalley from the delta systems the surficial cover is predominantly till veneers or colluvium, otherwise the surface is comprised of bedrock (Fig. 4.2). The till deposits are not extensive enough to describe a coherent pattern of former ice cover. However, remnants of ice marginal drainage channels on the mountain slopes depict former glacier occupation. These features were used extensively to construct the ice retreat map (Sec. 4.4).

4.3 Site Analyses

This section deals with the detailed geomorphology of specific areas within Clements Markham Inlet. Although deposits and glacial features were mapped in the highlands of upper Clements Markham River and its major tributaries, detailed field work concentrated on deposits found along the coastline where exposures and fossiliferous sediments are common. The following site descriptions begin at the head of Clements Markham Inlet and first follow along the NW coast for 55 km to Cape Colan, bordering the Lincoln Sea. The next set of descriptions follow its SE coastline from the mouth of Piper Pass to Hamilton Bluff bordering Parker Bay (Fig. 4.1). Each site is discussed in the following order: general description; geomorphology and stratigraphy; and interpretation. Official and unofficial place names are given in Figure 2.1.

4.3.1 The Lower Clements Markham River

4.3.1.1 General Description

The area of lower Clements Markham River is a broad plain which is occupied by the modern sandar, a large lake (ca. 8 km^2), and extensively gullied, raised marine silts (Fig. 4.2). The Crescent Glacier and the Barrier Glacier lie ca. 25 km upvalley (Figs. 2.1 and

4.2), whereas the Lime Cliff Glacier marks the western boundary (ca. 4 km WSW of the lake, Fig. 4.3, Plate 4.1). Glaciofluvial and delta terraces, rising above the silts, form the N and S boundaries of this plain and occur along the sides of the main NW-SE oriented valley.

The majority of the thick silt deposits are found in two bodies on either side of the Clements Markham River. One deposit (ca. 7 km by 1.5 km) lies immediately S and SW of the lake, and the other deposit (ca. 3 km by 1.5 km) borders the southern side of the river valley. These silt bodies are considered to form a single unit, which once occupied the entire lowland area, prior to being separated by subsequent fluvial erosion. The contemporary Clements Markham River sandur (ca. 2 km wide) now separates these silt bodies. River erosion into the northern silt body exposed a diamicton layer (ca. 3 m thick) underlying the silts. Christie (1967) referred to this area as Till Island (Fig. 4.3).

The southern silt body is bounded by a large delta system to the S (Corner Delta). The Corner Delta emanates from the mouth of a valley which lies parallel to, and midway between, the mouths of the Clements Markham River and Piper Pass. Two glaciers lie at the head of this valley ca. 13 km from the apex of the delta (Figs. 2.1 and 4.2).

4.3.1.2 Geomorphology and Stratigraphy

Three principal areas are described: Till Island, Corner

Figure 4.3 Lower Clements Markham River.

Note the large areas of extensively gullied, fine grained marine deposits. The two main silt bodies discussed in the text are found on either side of the river. The Corner Delta formed adjacent to a glacier which flowed out of the valley from the bottom center of the photograph. The arrow indicates the location of the section at Till Island discussed in the text. Several terrace elevations are given in meters a.s.l... Wind fluted silts are found at X. Radiocarbon dates 1 to 3 are described in Table 5.1; scale 1:60,000. (National Airphoto Library, A-16603-111)


CLIFF

Plate 4.1 Glaciers near the head of Clements Markham Inlet, looking northwest. The snout of the Lime Cliff Glacier is clearly seen as it overlies an outlier of ice-cored outwash. These glaciers are ca. 8 km from the present shore of the Inlet. Note the gullied marine deposits in the foreground, as well as the contemporary sandar associated with these glaciers. The Corner Delta is directly beneath the photographer (Fig. 4.3).

Plate 4.2 The exposure along the southeast side of Till Island (Fig. 4.3). Polished gypsum and dolomitic breccia is overlain by till which is, in turn, conformably overlain by stratified marine silts and fine sands.

faint bands of silt and fine sand form discontinuous beds within the upper parts of the till. Above this, the marine sediments form undisturbed, horizontal beds of silt and fine sand. The sediments are unfossiliferous and contain few drop stones. The drop stones form a lag of angular rock debris on the surface of Till Island.

Two fabrics about 500 m apart were measured on the till within the above exposure (Fig 4.3, Plate 4.2). Both sites contained numerous striated and faceted limestone, quartzite, and sandstone clasts, but Site 2 contained more pea sized and rounded material. The fabric obtained at Site 1 shows a relatively strong alignment with the long axis of the valley, and hence, the inferred glacial flow (NNW to NW). The Site 2 fabric, on the other hand, has a strong bimodal distribution with a secondary orientation orthogonal to the valley. A similar sequence of bedrock, till, and marine deposits is found within the second major silt body to the S, and near the Corner Delta.

The <u>Corner Delta</u> is comprised of a series of descending terraces overlying the stratified silts (Plate 4.3, Fig. 4.3). Major terraces have lip elevations at 101, 84, 70, 46 and 37 m, respectively. The terraces are comprised of unconsolidated sands and gravels which have indistinct, steeply dipping bedding. The uppermost terrace grades rapidly upvalley to an arcuate zone of hummocky, boulder gravel lying across the valley (110 m a.s.l.). Several large depressions dominate this area. As with the Till Island deposits, no <u>in situ</u> marine bivalves were found in the stratified silts underlying the Corner Delta complex. However, whole valves of the marine bivalve

Plate 4.3 The Corner Delta. This delta clearly shows the terminal position of ice which occupied the mouth of the valley during the time the uppermost delta terrace was being constructed into a 101 m relative sea level (Fig. 4.3). This is indicated by the arcuate zone of kettled outwash across the mouth of the valley in the center of the photograph. Note the proximity of the glaciers upvalley. Also note the dissected silts exposed on the right flank of the delta. The uppermost silts contained shells which dated 7635 ± 80 BP (SI-4761).

Hiatella arctica were found enclosed in a lobe of colluviated silts at an elevation of 82 m (site 1). The colluvial deposit is ca. 30 cm thick and overlies undisturbed silts; in turn, the 101 m terrace lies immediately upslope from this deposit. The radiocarbon age of the shells (7635 ± 80 BP, SI-4761) provides a minimum age estimate for the formation of the upper terrace.

The Lime Cliff Glacier lies 7 km to the NW of the Corner Delta, on the opposite side of the Clements Markham River valley (Fig. 4.3). Th[redacted] of this large valley glacier produces a major sandur which [redacted] into the Cl[redacted]m River. Christie (1967), Stewart (1981), and Stewart and England (1983) reported on abutting, ice-cored, outwash terraces which are currently being overthrust by the advancing glacier. These terraces were once part of a contiguous surface which has been displaced to form three distinct levels. Moreover, the terrace fragments are polygonal in plan view suggesting that decollement took place along a pre-existing network of ice wedge polygons. Similar glaciotectonic features have been described on Axel Heiberg Island (Kälin, 1971) and Bylot Island (Klassen, 1982). In addition, the ice core, exposed along a stream cut, lies 2 m below the surface and contains dirty and bubbly bands with pronounced downvalley dips. Christie (1967) suggested that the core may be glacial ice dating from a late Wisconsin or later Holocene advance. Of interest to this study is that these terraces may be an outwash surface which graded into a former sea level. The upper surface of the thrusted terrace is at 121 to 123 m, whereas a relatively undisturbed surface occurs at ca. 117 m. One km to the NE of the Lime Cliff Glacier, another large, flat-topped sand and gravel terrace occurs on the valley side at 104-106 m, and it may represent the same graded surface to a local marine limit of ca. 104 m (Fig. 4.3).

Similarly, a third large terrace is found 4.5 km to the NE of the Lime Cliff Glacier. This massive terrace rises steeply, 100 m above the lake, to 102 m (Fig. 4.3). It appears to be composed entirely of loose sand and gravel lying at the angle of repose, although minor silt outliers are found on its flank to ca. 30 m. Although no buried ice was seen beneath this terrace, a very large depression separates it from the adjacent valley side. A circuitous channel running through a spectacular limestone gorge drains the depression into the lake.

The last prominent terrace found in this area lies ca. 3 km S of the Lime Cliff Glacier and ca. 11 km from the sea. This terrace is also comprised of sands and gravels and it has a large depression occupied by a lake on the upvalley side. Its flat topped surface lies 101 m and may correspond with sandur remnants found farther up the Clements Markham River valley (Fig. 4.3).

4.3.1.3 Interpretation

A discussion of the sediments found at Till Island will be presented first. These deposits are significant because they are centrally located in the lower Clements Markham River valley, and consequently they record the late Quaternary events in the main valley system.

The principal concern here, as well as at other sites described in later sections, deals with the origin of the diamictons. Is the diamicton a till deposited by a grounded glacier, or is it a glaciomarine unit released from an overlying ice shelf? As noted in previous chapters, contemporary ice shelves occur extensively in the fiords of northernmost Ellesmere Island and they were probably more extensive during past glacial cycles. However, given the preferred orientation and characteristics of shear preserved in the unit, the diamicton layer found throughout the lower Clements Markham River area is interpreted to be an orthic till, deposited up-ice from the grounding line at the base of a glacier.

The stratigraphic sequence of Till Island is interpreted as representing a glacial advance-retreat cycle accompanied by a marine transgression. The initial advance scoured the valley floor, polishing bedrock surfaces and trimming off protuberances. A downvalley direction of shear stress is evident where resulting rock debris is incorporated into the lower part of the till. Subsequent deposition of till and retreat of the glacier was coupled with marine inundation resulting in horizontally stratified silts lying conformably over the till. In some parts, the highly deformable nature of the water-saturated till resulted in the mixing of the two units. Generally, however, the contact is sharp and no intervening glaciofluvial or littoral sediments occur. This overstep relation means that glacial retreat was rapid (see Ch. III). Conversely, coarse sediments overlying the marine silts are also absent. The lack of an overlying regressive facies suggests that postglacial emergence after the deep water phase was also very rapid. Large areas of silts form plains exposed at the surface where they are subject to deflation and intensive gullying. Only where coarse sediment input was high, such as at valley mouths, are the silts capped with gravels.

The preceding analysis indicates that most of the fine grained sediment in this area was deposited in a short interval between rapid glacial recession and subsequent rapid emergence. This implies that sedimentation rates during this time were extremely high in order for thick bodies of silt to be deposited. This may also be indicated by the absence of any marine fauna within these silts. High sedimentation rates, coupled with brackish water conditions caused by rapidly melting ice, may have prevented invasion by marine fauna. Moreover, marine shells were only found in beds at higher stratigraphic positions, indicating that environmental conditions for marine fauna improved only in late phases of deglaciation. The above sedimentary history is dissimilar to paleoenvironments farther down the Inlet which will be discussed in later sections.

After the main glacier retreated from the lower Clements Markham River valley the marine limit terrace of the Corner Delta prograded into the high sea at 101 m. Consequently, the sediments of the Corner Delta represent a later stage in deglaciation when the upland valleys were still occupied by ice. The arcuate zone of glaciofluvial ridges and kettles at the apex of this delta complex indicate that the terminus of the tributary glacier was immediately upvalley at this time and ample coarse sediment was supplied for progradation (Plate 4.3). The radiocarbon date suggests that this was happening prior to ca. 7.6 ka BP.

Whereas the origin of the Corner Delta terraces is clear, the large gravel terraces bordering the Lime Cliff Glacier are more difficult to interpret (problems of interpreting terrace genesis were discussed in Sec. 4.2.3). These terraces are not part of a continuous delta system, but rather they are found as isolated bodies abutting the valley sides. Consequently, they may have been formed along the margins of glaciers as indicated by their kettled nature. The terrace contacting the Lime Cliff Glacier, which presently contains a glacial ice core, may provide an analog for such a glaciofluvial terrace. Nonetheless, the similarity in elevation between the flat tops of the glaciofluvial terraces (104 m) and the marine limit at the Corner Delta (101 m), suggests that they prograded, or were regraded, to a similar relative sea level.

4.3.2 Owl Perch Delta

4.3.2.1 General Description

The Owl Perch Delta lies on the NW shore at the head of Clements Markham Inlet, immediately to the NNE of the prominent lake (Figs. 4.1, 4.4). Sediment and meltwater are currently supplied to the watershed by small glaciers on the flank of Mt. Rawlinson, about 6

Figure 4.4 Owl Perch Delta.

Note the descending delta terraces from the marine limit at 93 m a.s.l.. The upper part of the delta is marked by numerous ice contact features including moraines (M) and glaciolacustrine terraces (L). Arrows indicate various sections discussed in the text. Also note: Till in section (T); Stratified sand in section (S); Bedrock ridge (R). Radiocarbon sample 8 found in the dissected raised marine silts is described in Table 5.1; scale 1:30,000. (National Airphoto Library, A-16603-113)


Lake

km upvalley (Fig. 2.1). This watershed (ca. 60 km^2) is comprised of several subparallel tributaries which merge above the apex of the delta located at the valley mouth. The main channel shows thick accumulations of sediment overlying bedrock. However some of the surface relief within the delta is a result of underlying bedrock ridges which trend obliquely across the main channel.

4.3.2.2 Geomorphology and Stratigraphy

Like the Corner Delta, the apex of the Owl Perch Delta is characterized by thick deposits of angular clasts in a sand and gravel matrix. Large depressions are found on its surface near the valley sides. Some of the depressions are bounded on the seaward side by descending, arcuate ridges having asymmetrical cross sections. The inward slopes of the ridges tend to be steeper than the seaward facing slopes, suggesting an ice-contact origin.

About 1 km upvalley from the delta apex, benches of horizontally stratified, indurated silt lie at the confluence of two main tributary valleys at 130 m and 190 m (site L, Fig. 4.2). These silts are unfossiliferous and contain abundant dropstones. Seaward of the delta apex a series of descending delta terraces occur at 93, 81, 69-70, 47, 30 and 21 m. The 70 and 93 m terraces are paired, whereas the lower terraces are non paired and concentrated on the W side. Dissected marine silts are exposed along the periphery of the delta terraces, particularly on the E side of the delta system (Fig. 4.4). Abundant marine pelecypods in growth position were found protruding from the surface of a silt plain at 60 m (site 8). A radiocarbon age of 7540 ± 95 BP (UQ-260) was obtained from massive Mya truncata at this site.

Local exposures show a variety of deposits within this delta complex. A river cut near the valley mouth exhibits an indurated diamicton 3 m thick overlying bedrock (site T, Fig. 4.4). The diamicton, which contains striated clasts, is in turn, overlain by unconsolidated gravel lying on the surface. Further downstream, the diamicton is absent, and current bedded sands comprise the base of the section. These sands contain sparse Portlandia arctica pelecypods and plant debris, and in turn, are overlain by an unfossiliferous, coarse sand and gravel delta terrace (ca. 70 m). Because bedrock is not exposed at this site, the relationship of these sands to the diamicton is unknown. In other sections within the delta, stratified silts abruptly overlie glacially polished bedrock.

On the NW side of the Owl Perch Delta, a continuous section through a delta terrace lying at 47 m (Fig. 4.4), shows a progressive shift in sediment character related to distance from the former delta apex. Although the surface of the terrace is comprised of loose sand and gravel, the exposure shows that the terrace is in fact underlain by fine stratified sediments. The distal end of the section shows silts and fine sands having a gentle seaward dip. About 50 m closer to the delta apex, these sediments contain an increasing amount of rock debris. Simultaneously, there is a gradual shift in the fines to more sand, and finally dominantly sand. After 100 m laterally these sediments are obscured by loose sand and gravel which comprises the delta terrace surface.

The last stratigraphic section to be discussed is a ridge which lies transversely to the delta, on the NE side of the river (site R, Fig. 4.4). The sediment exposed here is unlike any other in this area. The most striking characteristics of the sediment is that it is extremely indurated and has a distinctly salty taste. The material within the W end of the exposure is coarse grained and poorly sorted. Generally the base of the section contains a heterogeneous mixture of rounded clasts supported by a matrix of silty sand. Conversely, the upper parts of the section contain lenticular beds of sorted gravels and numerous scour and fill structures. Eastward, these structures are lost and the entire section is a diamicton. The flanks of the ridge are conformably mantled by marine silts that comprise the upper unit in the section.

4.3.2.3 Interpretation

The history of the Owl Perch Delta resembles that of the Corner Delta. Ice-contact, glaciofluvial ridges and kettles attest that a tributary glacier occupied the mouth of the Owl Perch valley during initial stages of delta building. Prior to this, more extensive ice deposited till further downvalley, and this event may represent a period when the tributary glacier merged with the main trunk glacier occupying the Clements Markham River valley.

A subsequent position of the glacier terminus may be masked by the transverse-lying ridge in the lower delta, however, genesis of this deposit is uncertain. The west end of the ridge appears to be comprised of bedload and channel fill deposits, whereas the eastern end of the exposure may be till, although striated clasts are absent and the transition between these sediments is gradational. Whether glacigenic or not, the overlying marine silts indicate that the ridge predates the postglacial transgression by the sea. Furthermore, it is probable this deposit is indurated because of the cementing by salts, but to what extent this indicates age is not known. This deposit is certainly more indurated than any of the other Quaternary deposits found in the area. Christie (1967) reported widespread caliche deposits which he speculated were deposited during upward movement and evaporation of water from thawing ground.

When tributary glaciers occupied the Owl Perch valley, proglacial lakes formed at high elevations because the ice impeded drainage in the lower valley. The silt benches at 130 m and 190 m are considered to be glaciolacustrine rather than marine because they are unfossiliferous, and they occur considerably above the local marine limit at 93 m. Furthermore, they are a pale brown colour (10 YR 6/3 d) and are more indurated, compared to the grey colour (10 YR 3/1 d) and friability of the fossiliferous marine silts at lower elevations. This difference probably reflects their different sedimentary environments. Finally, proglacial lakes are common in the present environment in similar settings.

During deglaciation ice in the vicinity of the delta apex provided sediment which prograded into the sea at 93 m. The minimum age estimate for this event (7540 ± 95 BP, UQ-260) is provided by shells from the bottomset silts exposed in the distal parts of the delta complex (site 8). Insufficient detrital organics or shells were retrieved from the more proximal current bedded sands bordering the river channel. However, these sediments may postdate the marine limit phase if they are part of the same prograding sequence which led to the formation of the overlying sand and gravel terrace at 70 m. The section described on the NW side of the Owl Perch Delta, which cuts through a 40 m delta terrace, clearly depicts such a facies shift laterally as one moves from distal to proximal locations.

4.3.3 Gypsum River Area

4.3.3.1 General Description

The Gypsum River area provides the most stratigraphic and geomorphic detail in the Clements Markham Inlet (Figs. 4.1, 4.5; Plate 4.4). The river lies in a broad valley which is bordered to the W by an ice cap that supplies meltwater along the 35 km of river course into Clements Markham Inlet (Fig. 4.2). Although the upper part of the river flows SE, the lower part swings NE as it flows along a major strike trough which parallels the Inlet. The intervening ridge to the S, which separates the Gypsum River from the Inlet, is flat-topped and forms a plateau whose SW end abuts a pyramidal mountain rising to ca. 750 m (Plate 4.4). To the NE, the plateau is crosscut by a narrow

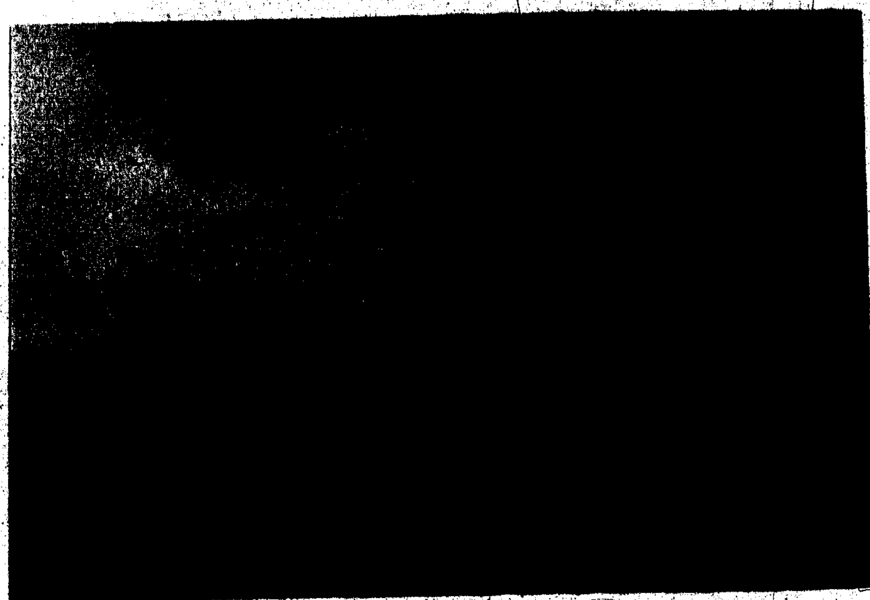


Plate 4.4 The bluffs of the Gypsum River area looking southwest, up the Inlet. Ice marginal channels on the seaward side of the pyramidal mountain in the background mark the approximate position of a major trunk glacier within Clements Markham Inlet ca. 10,000 BP. The bluffs form the seaward edge of a plateau over which lobes of ice from the right spilled into a high sea at 124 m a.s.l..

Figure 4.5 Gypsum River Delta.

Note the numerous ice contact ridges, kettles and abandoned meltwater channels on the plateau. Well developed outwash terraces are found on the landward side of the plateau at the junction of the Gypsum River and its two tributaries. Also note the marine limit terrace at 124 m a.s.l. and the well developed beach at 110 m a.s.l. Letters A to E indicate the location of the sections described in Figure 4.6. Radiocarbon samples 19 to 23 are described in Table 5.1; scale 1:30,000. (National Airphoto Library, A-16603-115)

gorge which sharply diverts the last part of the Gypsum River to the S. The mouth of the gorge forms the apex of the delta, 3 km from the shoreline (Fig. 4.5).

4.3.3.2 Geomorphology and Stratigraphy

The surficial map (Fig. 4.2) shows a variety of deposits in the Gypsum River area. The contorted nature of the valleys, the product of structural control, has led to a complex assemblage of landforms. The previously described plateau, S of the Gypsum River's prominent bend, deflected glaciers flowing down the Gypsum River valley. The SW part of the plateau has widespread glaciofluvial deposits and extensive ice-contact topography is found there, including moraines, kame terraces and kettled surfaces. The moraines (up to 8 m in relief) have a configuration that depicts at least three terminal positions of the ice which crossed the plateau obliquely (Fig. 4.5). Integrated networks of former meltwater channels commonly truncate these moraines and join many of the larger kettles. A number of these channels also terminate on the steep seaward (Inlet) side of the plateau. Stream cuts along the lower seaward slopes of the plateau exhibit fine grained sediments of marine origin at 13 m. A radiocarbon age on in situ shells of Mya truncata and Hiatella arctica from these sediments (site 19) yielded an age of 5720 ± 150 BP (UQ-258).

The marine limit in the area is clearly marked by the termination of a large flat terrace, ca. 0.5 km in width, that extends along the steep seaward side of the plateau. The seaward facing horizontal lip of this sand and gravel terrace lies at 124 m, whereas the adjacent sides of the terrace terminate at till covered, bedrock protuberances. Toward the NW, in an ice-proximal direction, the terrace slopes gently upward and clast sizes increase. Eventually the proximal end of the terrace surface terminates at small asymmetric moraine ridges which mark the edge of a kettle ca. 160 m (Fig. 4.5).

The landward side of the plateau is dominated by fluvial gravels and flights of ice-contact ridges and terraces which descend to the contemporary Gypsum River sandur. A complex sequence of base level changes is indicated by these successive terraces that must be the result of retreating glaciers and land emergence, coupled with gorge cutting by meltwaters.

In the eastern sections of the Gypsum River delta large volumes of outwash at high elevations are less common. However, a bedrock ridge, the structural continuation of the plateau W of the gorge, has a gravel bench running along its seaward side at 110 m (Fig. 4.5). This feature has a hummocky tread (± 0.5 m), nonetheless, it is level for a distance of ca. 750 m and it is interpreted as a beach marking a relative sea level at 110 m. An important aspect of this beach is that it was partially formed on the landward, ice-proximal side, of the bedrock ridge. Because this ridge also formed an abutment for any glaciers flowing down the Gypsum River valley, it follows that this area must have been ice-free before the time of the 110 m relative sea level (Fig. 4.5). A number of small moraines in the trough along the landward side of the ridge trace the retreat of the ice upvalley.

Smaller, fine grained delta terraces mark former sea levels at 84, 70 and 69 m, respectively. Like the SW sector of the Gypsum River delta, the lower seaward slopes of the NE sector are also mantled with fine grained, deep water, marine sediments which form very light-toned areas on Figure 4.5. A horizontally stratified silt outlier, ca. 35 m thick, forms a prominent ridge running parallel to the shore, and transversely across the delta for ca. 500 m (Fig. 4.5). Abundant, articulated shells protruding from the surface of the outlier (65 m, site 23) initially yielded the oldest date yet recorded on marine sediments within Clements Markham Inlet (10,690 ± 520 BP; S-2137). Nevertheless, a second date on the same sample produced an age of 7940 ± 130 (S-2483) suggesting that the first date is in error because of a problem within the radiocarbon laboratory.

The most distal sections of the Gypsum River delta are dominated by several very large terraces lying between 9 and 23 m. These more recent stages of delta building are centered on outflow from the mouth of the gorge and truncate the older marine deposits. Channel cuts into the contemporary delta surface expose well developed foreset-topset sequences suggesting that active progradation is still in progress.

Numerous gullies cut into the seaward slopes of the Gypsum River delta expose complex sequences of glacigenic and marine facies.

The general succession of facies throughout the area is similar. Figure 4.6 illustrates five representative sections within the Gypsum River delta, from the SW side to the NE.

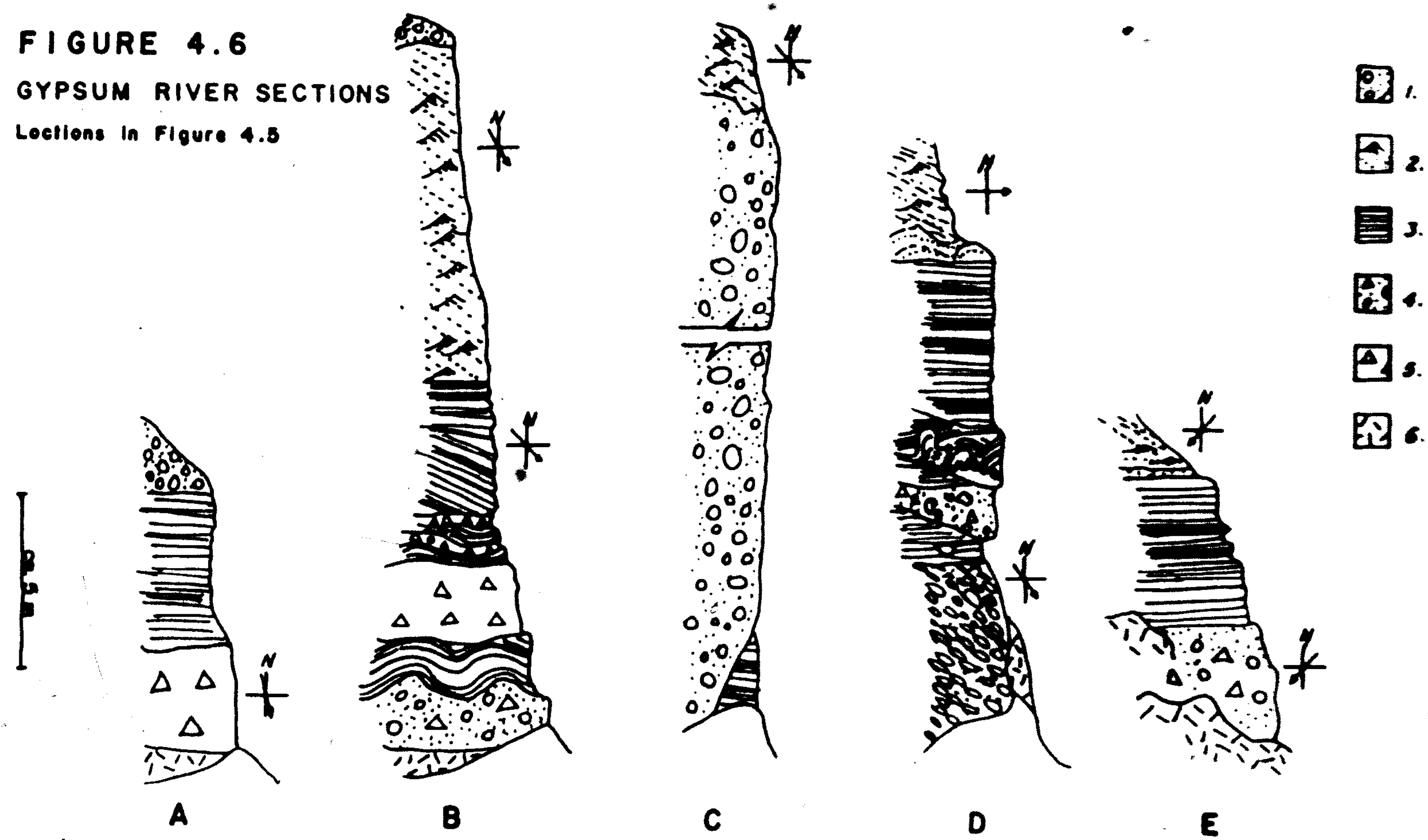
The first section (A) exhibits a well indurated, grey diamicton overlying bedrock. The diamicton is 3 m thick and contains striated and faceted clasts similar in appearance to the till described in the Owl Perch delta. Fabric analysis indicates a preferred orientation of 152° TN, and a 13° dip, in the seaward direction (Fig. 4.6). Therefore the diamicton is considered to be a basal till. The till is overlain along a sharp contact by 4 m of horizontally stratified silts. The silt, in turn, is overlain by unconsolidated gravels which comprise the upper delta terrace. The contact between the latter two units is not clear.

Section B, a short distance NE, shows a different diamicton overlying bedrock. This diamicton is darker, has a sandier matrix, and contains clasts that are more rounded and less faceted, than those in the till of section A. The upper contact of the diamicton is undulating ($\pm$ 50 cm) and bedded silts conformably overlie it. These silts, in turn, are overlain by a 2 m thick diamicton having similar characteristics to the till described at site A, containing numerous faceted and striated clasts. Silts interbedded with thin, steeply dipping diamicton bands overlie the diamicton layer. The silt beds show signs of disturbance, but undisturbed silt beds ca. 4 m thick and having a 6° SE dip, overlie this zone of intercalation. The thick, undisturbed silt unit is grey in colour and unfossilferous, however

FIGURE 4.6

GYPSUM RIVER SECTIONS

Loctions in Figure 4.5



1. gravel and sand 2. stratified sand 3. silt laminae 4. sandy diamicton 5. till 6. bedrock the uppermost beds change to a pale brown colour and coarsen into stratified sands. Occasional Portlandia arctica were seen in the uppermost silt beds.

The overlying stratified sands are at least 13 m thick and are widely exposed along the SW side of the Gypsum River delta. Numerous primary sedimentary structures associated with current deposition are present, including climbing ripples, large scale trough crossbedding, and plane beds. Occasional gravel bands and clay laminae are also found. These structures indicate a general paleocurrent direction of 144° TN, which is approximately the same direction as the 15-20° dip exhibited by the uppermost sand beds of this unit.

The upper part of the sand unit contains layers of detrital plant remains as well as in situ pelecypods. A number of paired valves were found at the top of vertical burrows cutting through the sand layers. These infilled burrows show the adaptation of the pelecypods to the rapid deposition of sand (cf. Reineck and Singh, 1980, p. 172). A radiocarbon age of 8660 ± 155 BP (S-2124) was obtained on shells collected from these sands (site 21). The sand unit, in turn, is overlain by sands and gravels of a 96 m delta terrace.

Section C, near the mouth of the gorge (Fig. 4.6) exposes a thick deposit (40 m) of sand and gravel, inset into stratified silts. The sands provide a matrix which supports larger clasts and bedding is absent. This unit is overlain by the stratified sands that extend from section B.

Section D, at the mouth of the Gypsum River gorge, has a more complicated sequence. The lowermost unit in the section is 5 m of imbricated gravels dipping 20° in a general SE direction. The gravels are well rounded with only occasional striated and faceted clasts. Colluvium obscures the base of the unit, whereas the top consistes of 30 cm of bedded silts containing a few dropstones. The silts, in turn, are overlain by 1.5 m of sandy gravel which truncates the silts. No bedding was evident within the gravel. The upper limit of the gravel unit is composed of intercalated bands of gravel and silt. The silts are usually folded, however, a number of rotated slump blocks of silt are also present. Gradually the zone of disturbance is overlain by 5 m of undisturbed silts. These silts, in turn, are overlain by current bedded sand recording a paleocurrent to the E.

In contrast to section B, no marine pelecypods were found within section D. Although the stratified sands are similar in appearance to those in sections B and C, they have a different paleoflow direction. Nonetheless, a single trunk of <u>Salix</u> sp. incorporated into these sands (site 22) dated 23,850 $\pm$ 850 BP (S-2140).

Section E is found on the NE periphery of the Gypsum River area. This section is one of several similar exposures produced by N-S trending gullies oriented toward the mouth of the next valley to the NE (Moraine Creek area). Undulating gypsum bedrock forms the base of the section. Generally a well indurated, sandy diamicton, up to 4 m thick, overlies the bedrock, but it is absent in some exposures. The diamicton contains many rounded clasts, particularly gypsum, and its matrix has a salty taste. Fabric analysis indicates strong clustering oriented 207° TN, with a 29° dip, which is from the direction of Moraine Creek valley (Figs. 4.1, 4.5).

The diamicton and/or bedrock is abruptly overlain by ca. 2 m of stratified grey silts that are unfossiliferous. These barren silts, in turn, are overlain by fossiliferous, pale-brown silts. The transition between the two silt units is marked by alternating strata of pale brown and grey silt. In situ shells collected from the upper pale brown silts at 45 m (site 24) dated 6865 ± 130 BP (S-2135). These silts gradually coarsen into sands whose beds dip gently to the SSW. The sands form the uppermost unit and are covered by a lag of pea gravel at the surface. A series of incipient beach ridges superimposed on this surface extend up to ca. 89 m.

4.3.3.3 Interpretation

The geomorphology and stratigraphy of the Gypsum River area exhibit the retreat of glaciers from positions along the shore of Clements Markham Inlet. The glaciers that flowed down the valley during their extensive stage had to flow upslope to surmount the plateau which separates the main valley from the Inlet. This would lead to a thinning of ice over the plateau. On the steep seaward side of the plateau, the ice then encountered a high sea at least 124 m, and consequently further advance was inhibited by calving along the edge of the plateau. This semistable ice front is documented by moraines trending across the plateau. Subsequent moraines also mark the retreat from this configuration (Fig. 4.5).

The general stratigraphic sequence suggests a single glacial cycle followed by a marine transgression and postglacial rebound. However, the type of glacigenic deposit attributed to the glacial advance varies between sections. Furthermore, the relationships between these sediments and the overlying marine silts also differ. Therefore the model of one glacial cycle recorded in these sediments may be simplistic. All the basal exposures along the SW side of the delta exhibit either a diamicton or coarse outwash over bedrock. Two distinct diamictons are present; one with faceted and striated clasts enclosed in a silty matrix, and the other with mainly rounded clasts in a sandy matrix. The former is similar to tills reported from nearby areas and it was probably deposited by a glacier. The origin of the latter diamicton is less obvious as clearly defined analogs are absent. The two diamictons are mutually exclusive in all but section B where the sandy diamicton is stratigraphically lower than the till.

Section A contains the diamicton that is unquestionably a till. The stratigraphic sequence here is similar to the one described at Till Island (Sec. 4.3.1). A grounded glacier apparently advanced across the bedrock surface depositing till whereas its subsequent retreat was followed by a marine transgression. Till fabrics indicate that the glacier flowed down from the plateau to this area, whereas the abrupt contact with the overlying silts indicates a rapid retreat and transgression. Unlike the other sections described, an absence of intercalating beds of marine and glacigenic deposits in section A, suggests that a quasi-stable ice front did not exist in this area during initial postglacial transgression. The ice flowing off the plateau was more extensive at this site. The reason may be that section A lies downslope from a shallow trough which crosses the plateau and ice-contact features indicate that it was occupied by a tongue of ice. Local thickening of ice flowing seaward, together with a more gradual seaward slope, allowed for a more extensive advance here compared to thinner, more restricted, ice elsewhere on the plateau. An alternate source of till being deposited here could be from main ice in the Inlet, but the fabric within the till suggests that flow was perpendicular to what would be expected.

The above till unit is also recognized in section B, but in this case the till is underlain by marine silts and the sandy diamicton. The sandy diamicton records an earlier event possibly unrelated to the glacial advance recorded in section A. However its absolute age and origin are presently unknown. Because the two units are so dissimilar the genesis of the sandy diamicton may be glaciomarine, or even colluvial, relating to a period of slope instability. This period of instability could be the result of high pore water pressures induced by the initial transgression early in the glacial cycle. The overlying silt unit may also relate to the transgression. Nonetheless, as the glacial cycle proceeded a glacier advanced onto the silts, depositing the upper till. The contact with the till indicates a complex alternation of coarse glacigenic and marine deposits indicating a proximal, ice marginal environment. Proglacial fans or debris flows, issuing from the ice front, episodically disturbed syndepositional marine silts in deeper water more distal to the ice.

Following the retreat of the ice front, the deposition of marine silt became dominant with only minor disturbance from ice-rafted clasts. The sequence of channels and moraines indicate the successive margins of an ice front retreating north-westward across the plateau. As end moraines were being produced at an elevation of ca. 160 m, an outwash terrace was built into a sea 124 m above present. Subsequent to the marine limit stage, the focus of meltwater outflow shifted to the NE, toward the present delta apex, where large terraces prograded into 107 m and 96 m relative seas. Eventually the environment became favourable for marine fauna sometime before 8.7 ka BP and the prograding delta deposited sands which eventually covered the area.

Where the massive sand and gravel deposit of section C (Fig. 4.6) fits into the above scenerio is not known. This 40 m unit is inset in the silts. If the silts relate to the last glacial cycle, the massive unit may indicate extreme sedimentation at a former center of meltwater outflow near the ice margin, perhaps in a subaqueous environment.

Section D, on the other hand, has no till but contains imbricate gravels of glaciofluvial origin which may be the facies equivalent of the tills deposited in sections A and B and the massive gravels of section C. Their deposition probably occurred in a submarine environment; perhaps fed by subglacial drainage issuing from the bedrock trough. Nonetheless, marine deposition followed for a short interval, which, in turn, was succeeded by a phase of fluctuating sedimentary conditions under which intercalated sequences of coarse gravels and silt beds were deposited. Coarse sedimentation from the edge of the ice front occurred intermittently, deforming the silt beds which were deposited during intervening periods of quiescence. The sedimentary environment is interpreted as a proximal, proglacial zone characterized by calving, ice-rafting, sediment flow, and seasonal meltwater production.

Section E, on the NE extremity of the delta, exhibits a sandy diamicton at its base. This diamicton has the same appearance as the diamictons of uncertain origin described in the other sections. Furthermore, the salt taste and induration are similar to the diamicton described at the Owl Perch Delta, perhaps indicating a comparable age. However, there may be no such relationship because of the ubiquitous gypsum in the area. Nonetheless, if the fabric of this unit is a result of glacial transport, the indicated flow is from a small valley and transverse to the flow from the main Gypsum River valley. Alternatively, if the deposit relates to a much older glacial episode, it may have been deposited by ice flowing down Inlet. Material identified as till is not found in the locality of Section E, however, it is found at the mouth of the small adjacent valley to the NE (Moraine Creek, Fig. 4.5).

In section E, grey marine silts abruptly overlie both the sandy diamicton and the bedrock, suggesting a rapid marine transgression. These barren silts in turn grade into the fossiliferous pale brown silts which dated 6.8 ka BP. Although the sedimentation between the two silt units appears continuous the date conflicts with older dates found upvalley, hence a hiatus in the section seems probable.

In summary, the stratigraphy of the Gypsum River area shows that glaciers, emanating from the valley, advanced into Clements Markham Inlet during the last glaciation. The extent of the advance in this area depended on the gradient of the submarine slope and the local variation in ice thickness. Generally, the ice remained in a quasi-stable position along the edge of the plateau, calving and dumping material into the sea. This glaciomarine environment was established sometime before 8.7 ka BP and it may have existed considerably earlier (> 10 ka BP), as will be proposed in a later chapter. Subsequent ice retreat and emergence from a marine limit at 124 m occurred before 8.7 ka BP, when major delta terraces were constructed.

4.3.4 Moraine Creek Area

4.3.4.1 General Description

Moraine Creek flows in a narrow valley which is ca. 3 km NE of the Gypsum River (Fig. 4.1, Plate 4.5). The valley is only 6 km in length and presently contains no glaciers. However, given the orientation of the valley, it must have comprised part of the expanded valley glacier network flowing down the Gypsum River valley during the last glaciation. The narrow mouth of the valley lies ca. 2 km from the Inlet shore, and a bedrock ridge 100 m in elevation lies along the shore, enclosing an intervening basin (Fig. 4.7).

4.3.4.2 Geomorphology and Stratigraphy

Generally the lower slopes in this area are covered by marine silts up to 100 m. Lateral moraines at the narrow mouth of the valley rise from 140 to ca. 200 m within 1 km upvalley. Very coarse and angular gravel terraces, at elevations from 130–180 m, are associated with these moraines (K, Fig. 4.7). Numerous kettles and small, sharp-crested, arcuate ridges within the terraces suggests that these are ice-contact features. Some distance downvalley, terraces comprised of finer sands and gravels are graded to a local marine limit of 110–111 m. At lower elevations the basin contains outliers of till and marine silts mantled by solifluction material. The bedrock ridge bordering the Inlet, on the other hand, is covered by *in situ* frost shattered bedrock, erratic rock debris, and marine silt. Shell

Plate 4.5 The northwest shore of the Inlet looking southwest from the marine limit terrace (114 m a.s.l.) above Omega Bay. The mouths of four valleys are visible in the distance; including Eider Delta, Moraine Creek, and finally Gypsum River where the marine limit rises to 124 m a.s.l. (Plate 4.4).

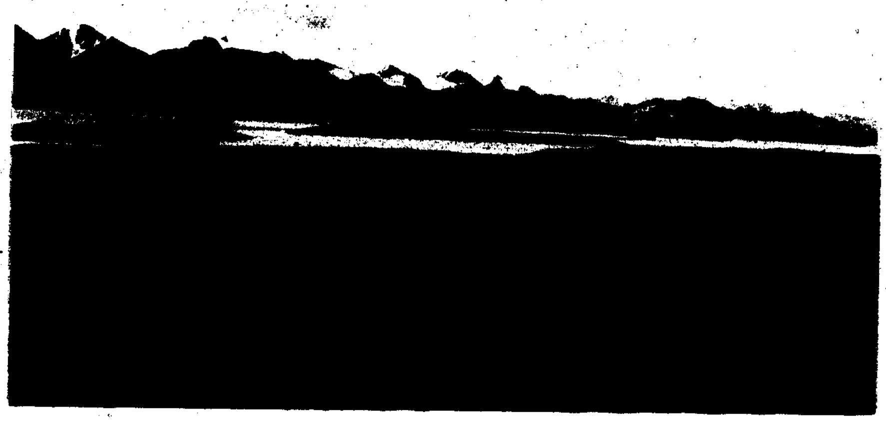
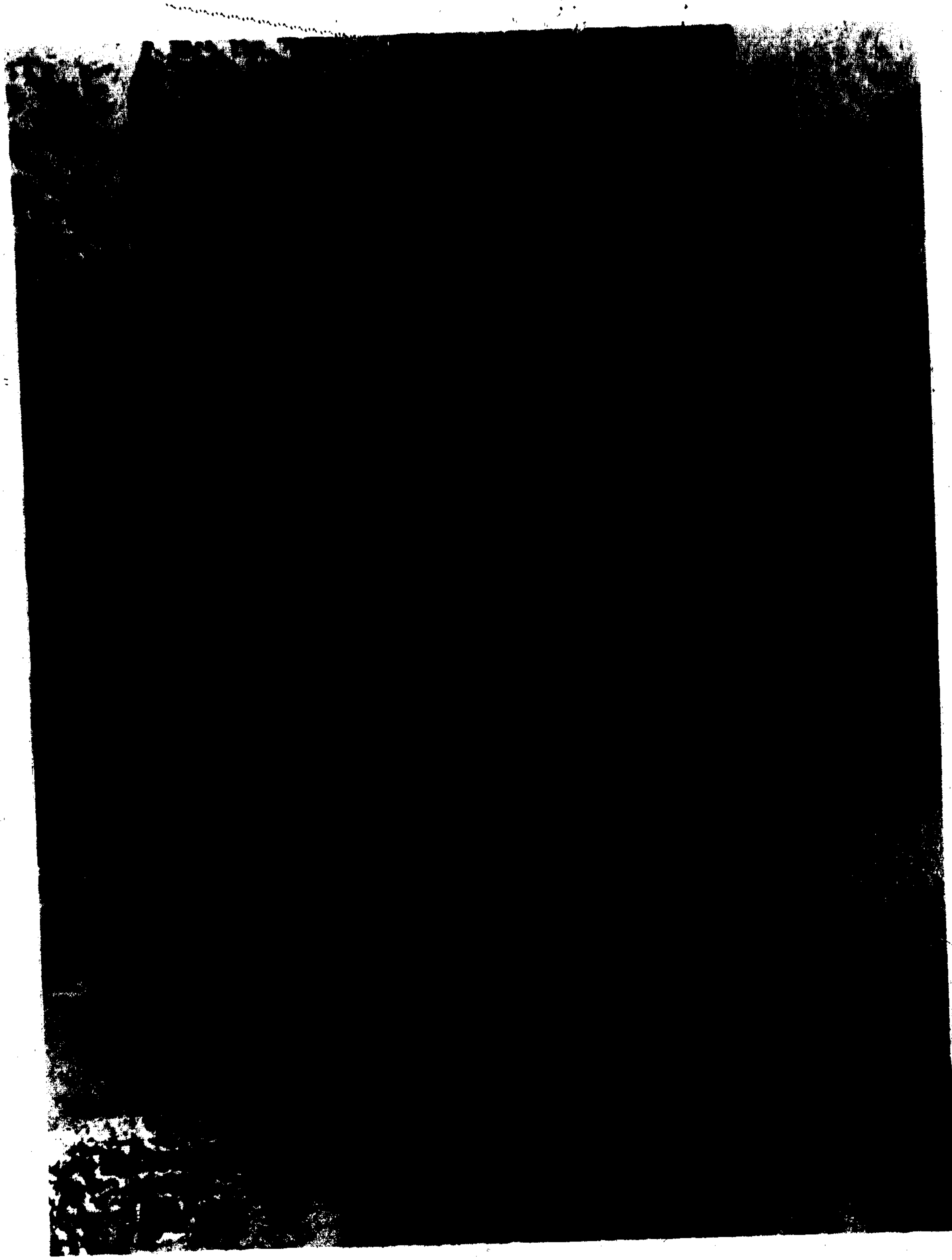


Plate 4.6 A view of the 114 m a.s.l. marine limit terrace above Omega Bay, looking south across the Inlet toward the Grant Ice Cap.

Figure 4.7 Moraine Creek and Eider Delta.

Note the prominant lateral moraines within Moraine Creek valley. Several kame terraces (K) associated with a more extensive ice front lie from 130 to 180 m a.s.l.. A prominant marine limit beach lies at 118 m a.s.l. at Eider Delta. Arrows indicate abandoned meltwater channels. Moraines at 'M' were formed by glaciers flowing from the Gypsum River valley. Radiocarbon dates 25 to 29 refer to Table 5.1; scale 1:30,000. (National Airphoto Library, A-16607-105)



fragments within this frost churned silt (site 27) dated 8605 ± 140 BP (S-2115).

At lower elevations in the basin, 500 m from the mouth of the valley, gullies expose a till containing many striated and faceted clasts. The till has similar characteristics to the other definite tills described previously. Elongated clasts within the till indicate a predominant fabric aligned with the mouth of the valley. The till overlies bedrock and, in turn, is abruptly overlain by horizontally stratified marine silts. The silts contain <u>in situ</u> shells (site 26) dated 9270 ± 265 BP (S-2105). Approximately 600 m seaward of this site, extensive silts at the same elevation contain shells (site 25) which dated 9845 ± 485 BP (S-2123). These silts occur above thick, barren silts overlying bedrock.

4.3.4.3 Interpretation

The Moraine Creek area stratigraphically records a marine transgression following the retreat of the ice front. Ice flowing from the Moraine Creek valley occupied the basin and as it subsequently underwent retreat, it exposed the peripheral areas to the sea by at least 9845 BP. Further inland, the sharp contact between the till and overlying silts suggests a rapid transgression following ice retreat. The minimum estimate for deglaciation here is 575 years (9270 BP) later than in the lower valley, and suggests a rate of retreat of < 1 myr^{-1}. However, given the large standard errors on the dates it is not possible to determine precise retreat rates. Numerous kame terraces at the valley mouth suggest a number of stillstands in this vicinity. The major one occurred during moraine formation after 9.2 ka BP.

The precise extent of the last glaciaton in this area is not known, however, if the sandy diamicton described on the NE side of the Gypsum River area is till (4.3.3.2, section B), it may have originated from Moraine Creek. Alternatively, the bedrock ridge, which formed a seaward buttress to the ice occupying the basin, may have remained ice-free. Till was not found in this area, rather only frost churned marine silt and local rock occur together with some ice-rafted debris.

4.3.5 Eider Delta Area

4.3.4.1 General Description

The Eider Delta lies at the mouth of a valley draining small cirque glaciers ca. 20 km upstream (Figs. 4.1, 4.7). The watershed lies entirely within the mountains of the northern coastal region which are free of ice caps. The narrow valley leading from the interior terminates ca. 2 km from the Inlet shore where initial delta building took place. Presently a deep gorge cut into limestone breccia connects the valley mouth with the contemporary apex of the delta. The upper seaward slopes contain a number of shallow troughs which trend E-W to the mouth of Moraine Creek.

4.3.5.2 Geomorphology and Stratigraphy

Like Moraine Creek, the mouth of the Eider valley contains several coarse glaciofluvial terraces recording glacial retreat and outwash production into a high sea level. Several former drainage networks, marking former ice margins, are found on the E side. The uplands are generally covered by a thin till veneer, whereas minor till ridges occur immediately above and below the marine limit. An extensive, well-washed, sand and gravel terrace marks the marine limit at 118 m (Plate 4.5). The slopes below the marine limit are mantled with silts which are in turn capped with angular rock debris. The uppermost fossiliferous silts occur on the E side of the river (site 29, Fig. 4.7) and dated 8535 ± 140 BP (S-2106). Shells collected from silts at 78 m on the W side of the river (site 28) dated 7640 ± 140 (S-2136). The shell samples were collected from the surface of the silts and both whole valves and fragments were used for dating. A series of sand and gravel delta terraces are inset into the silts forming terraces at 103, 41, 37, 30 and 9 m.

4.3.5.3 Interpretation

A glacier occupied the mouth of the valley and contacted a former sea level at 118 m a.s.l.. Because exposures in the gullies and river bank beyond the valley mouth contained only deltaic gravels, the extent of the ice is not known. Nonetheless, the large volume of surface lag gravels on the marine silts may indicate that a floating ice tongue or an actively calving ice front existed nearby.

Deglaciation of the eastern side of the Eider Delta and establishment of the marine limit (118 m) occurred sometime before 8.5 ka BP. However, the western side of the delta may not have become ice-free until ca. 900 years later. However, both dated shell samples were surface collections from sites that may still have been covered by ice during the establishment of the adjacent marine limit. Consequently these dates may only indicate that subsequent marine incursion was earlier on the eastern side of the Delta than on the western side, and no relationship exists with the marine limit terrace.

As the ice retreated from the valley mouth and emergence began, large volumes of meltwater issued from the valley and produced several shifting channels before the present gorge was excavated. Early outflow of meltwater was concentrated on the E side which left large volumes of sands and gravels deposited there (Fig. 4.2).

4.3.6 Omega Bay Area

4.3.6.1 General Description

The Omega Bay area contains two converging delta systems. The rivers which formed the deltas originate from presently vacant cirques in the nearby mountains. Similar to other areas described along this segment of the Inlet, an elevated plateau of 1-2 km in width lies along the foot of the mountains. The plateau contains a number of strike-controlled valleys and shallow troughs which trend

E-W, obliquely toward the Inlet shore. As the rivers flow into the Inlet, they truncate the plateau forming gorges (Fig. 4.8).

The area between the Eider Delta and Omega Bay is included in the discussion of Omega Bay. This area is comprised of a broad, hanging valley which has a central basin occupied by a lake. Present drainage from the lake is not directly seaward but connects to the Omega Bay drainage to the E (Fig. 4.8, Plate 4.5).

4.3.6.2 <u>Geomorphology and Stratigraphy</u>

The mouth of the broad, hanging valley between Eider Delta and Omega Bay is described first. A surface mosaic of bedrock and disturbed marine silts, mixed with rock debris and gravel, occurs on the seaward slopes up to 50 m. Outwash and till exhibiting numerous minor morainic ridges cover the surface at higher elevations. A large moraine dams the above described lake at ca. 160 m, above which lie flat terraces of angular gravel. Two major terraces on the W side of the lake occur at 163 and 170 m, and extend horizontally for ca. 500 m toward the sea, terminating abruptly on the slope (Fig. 4.8).

The local sandstone bedrock, exposed on the slopes above the terraces, exhibits a distinct weathering break at 320 m. The break is characterized by increased surface weathering above this elevation, where inclusions within the rock were inset and etched up to 3 cm in depth. Conversely, below this level, inclusions were flush with the relatively smooth bedrock surface. Unfortunately, this weathering

Figure 4.8 Omega Bay.

A moraine dammed lake is found in the upper left corner. Terraces occur above the lake, at 163 and 170 m a.s.l., which are probably kames, however a marine origin cannot be completely ruled out. Note the prominent marine limit beach at 114 m a.s.l. (left center) and the continuous beach at 110 m a.s.l.. Also note: several former meltwater channels (arrows); the basin containing numerous kame terraces (K); and the bedrock ridge (R) which rises above the marine limit. Radiocarbon dates 30 to 34 refer to Table 5.1; scale 1:40,000. (National Airphoto Library, A-16604-81)

break could only be discerned in this one locality because different lithologies elsewhere did not exhibit it. Moreover, no indication of a weathering break was found on till veneer covering a hill (ca. 306 m) which separates this area from the SW river at Omega Bay (Fig. 4.8).

Moving to the Omega Bay area proper, the apex of the delta system comprising the SW sector of Omega Bay contains a prominent sand and angular gravel terrace abutting a weathered bedrock protuberance. The terrace forms a horizontal surface at 114 m and marks the marine limit in the area (Plates 4.5, 4.6). To the N, just across the southernmost river, a subdued moraine occurs at about the same elevation. The moraine terminates at the steep mouth of the valley where it is obscured by colluvium. Downslope, the SW delta system also contains gravel terraces at 19, 36, 40, 46 and 56 m lip elevations (Fig. 4.8; due to the close spacing of these terraces, all are not shown).

The seaward slope of the intervening area between the SW and NE delta systems of Omega Bay has a long, well developed beach at 110 m. The beach is comprised of well rounded, pea sized gravels and sand, and is level for 200 m. Furthermore, this elevation coincides with a notch cut along the hillslope farther to the W. Conversely, the E end of the beach abruptly terminates at the edge of a shallow trough, a former channel of the NE river (Fig. 4.8). Conglomerate bedrock exposed upslope indicates that the well rounded nature of the gravels comprising the beach is an artifact of the conglomerate, and not due to recent water action. No dateable material could be found associated with this beach, and the highest level of marine silts lies downslope at 69 m. The silts are disturbed and overrun by solifluction lobes which contained shell fragments (site 31) dated 7275 ± 133 BP (S-2127).

The lower course of the NE river valley in Omega Bay is characterized by a winding bedrock canyon which cuts through the plateau. A broad strike trough crosscuts the river valley ca. 2 km from its mouth. The rim of the resulting basin contains numerous terraces of coarse angular gravel. The upper terrace, just upslope from the 110 m beach near the junction of the trough and the NE valley, lies at 119 m and it rises steeply up the trough (K, Fig. 4.8). The NE delta contains a series of lower terraces at 15, 22, 25, and 34 m which are comprised of foreset bedded sand and gravel.

The NE extremity of Omega Bay is characterized by flat uplands near the marine limit, truncated by strike troughs. Lower bedrock knobs are flat-topped and mantled with well rounded gravels. Although moraines are absent, a higher bedrock ridge is breached in a number of places by gravel floored channels, suggesting an ice-contact environment along its landward side. A horizontal gravel beach lies on the seaward flank of this ridge at 116 m. Seaward from this, in lowland areas, depressions are filled with horizontally stratified marine silts extending up to 74 m. A sample of a few <u>in situ</u> shells and fragments from these silts (site 34) dated 7850 ± 130 BP (S-2117). Farther to the NE a prominent sand and gravel terrace at 71 m marks a subsequent sea level on the Inlet side.

Exposures along the river of the SW delta exhibit crossbedded fine sands and silts overlying bedrock. This unit is overlain by horizontally bedded silts which exhibit reverse grading into foreset beds of silty sand which, in turn are overlain by gravel delta terraces. An important aspect of the delta stratigraphy is indicated along this section. The sections show that the uppermost foreset beds begin at a notch in the hillslope at 45 m with each successive, prograded bed running continuously to the valley bottom at an angle of 25°. From this, it is clear that the subsequently formed delta terraces at 40, 36, and 19 m were cut into the pre-existing delta. These terraces have no relationship to the foresets which comprise them. This stratigraphic evidence is critical when interpreting <u>in situ</u> shells obtained from the foreset beds (site 30) which dated 6860 ± 80 BP (S-4765). Clearly the minimum sea level estimate is 46 m, or higher.

Sections along cuts within the NE delta system display barren, horizontally bedded silts abruptly overlain by poorly stratified sand and gravel. The sands and gravels which comprise the delta terraces form an erosive boundary and are inset into the silts. Foreset bedding ~~is~~ evident with alternating bands of loose sand and gravel dipping seaward at 25°. A unit of coarse topset beds ca. 50 cm thick forms the treads of the terraces. Two <u>in situ</u> shell samples were obtained from the foreset beds of two terraces. The foreset beds of both could be traced to intersecting topset beds indicating relative sea levels at 40 and 15 m (sites 33 and 32). The ages obtained were 6195 ± 120 BP (S-2126), and 5255 ± 110 BP (S-2125) respectively. The latter shell date is clearly too old for the 15 m relative sea level (see below), indicating that the 15 m terrace level has been cut into an older, higher delta terrace. This similar condition was noted in the SW delta system (further shown in Ch. V).

4.3.6.3 Interpretation

This area is similar to Eider Delta in that numerous kame terraces and moraines in the upper parts of Omega Bay indicate that glaciers reached the lower valleys, however, their outermost extent cannot be discerned. Although a diamicton veneer mantles large areas below the marine limit, it is thoroughly mixed with silt and therefore it is probably glaciomarine in origin (i.e. formed from ice rafting, or an ice shelf). However, in the SW area, till veneer on the uplands and summits, above the preserved weathering break at 320 m a.s.l., indicates a more extensive glacial cover of unknown age. The weathering break itself indicates a stillstand or subsequent readvance during which the ice was ca. 200 m thick in the SW area. The third major ice extent in this area is indicated, downslope of the weathering break, by the kame terrace at 170 m. This terrace may indicate the ice extent during which time the marine limit formed but it cannot be traced directly to a marine limit indicator. Nonetheless, numerous lower kame terraces and moraines record the subsequent retreat from this last position. These retreat patterns will be illustrated and discussed in Section 4.4. An alternative hypothesis is that the 170 m terrace records a sea level which predates the last glaciation. If so, this would be the only example yet found in Clements Markham Inlet.

Like a number of other areas previously described, determining the marine limit is difficult where kame terraces, moraines and marine limits occur at approximately the same elevation. For example, if all these features coincide in elevation near the mouth of a valley, it suggests that the former ice thickness approached the depth of water. In this case the glacier may have been sufficiently thin that it was forced to float as it debouched into the sea. Seaward of the grounding line the surface gradient of the ice would be horizontal and lateral flow would produce horizontal ice-contact features (cf. England et al., 1978). Furthermore, should outwash or slope debris accumulate along these horizontal margins they would produce kame terraces or moraines at approximately the freeboard elevation of the glacier. Conversely, above the grounding line concentrations of marginal outwash or slope debris would have a gradient controlled by the glacier abutting the valley sides.

Given the above problems, the difficulty of discerning an ice-contact marine limit may be considerable. Nonetheless, two main criteria were used to differentiate the likely origin of terraces in such environments. A primary condition for defining a marine feature is its horizontality and its lateral continuity as indicated by other strandline indicators such as delta terraces, wave-cut notches, etc. (see Ch. III). Secondly, marine terraces tend to contain more rounded and finer sediment than kame terraces which, in turn, contain coarser angular material with a high incidence of striated clasts.

The Omega Bay area characterizes the hypothetical problems described above. For example, the 114 m marine limit terrace at the SW delta must have formed when the valley glacier occupied the valley mouth. Here moraines indicate that the water depth and the ice thickness were similar such that the terminus was probably buoyant beyond the mouth of the valley. Similarly, glaciers occupying the upper valleys leading to the NE delta system produced the upper kame terrace which grades down to 117 m at the valley mouth where it terminates. The long 110 m beach which lies just below this point was formed after the glacier vacated the valley mouth. Marine shells provide a minimum estimate for local deglaciation ca. 7850 BP and this event is marked by prograding outwash, forming distinct delta terraces. As previously described, substantial outwash production continued until relative sea level fell to ca. 56 m. Subsequently, lower terraces were cut into these sediments. The period of low sediment input after the ca. 56 m relative sea is thought to indicate the demise of the local glaciers within the Omega Bay watershed (ca. 6.5–7 ka BP).

4.3.7 Windswept Valley

4.3.7.1 General Description

Windswept Valley lies about two-thirds the way down Clements Markham Inlet and it comprises the last major catchment (126 km^2) draining into the Inlet's NW shore (Fig. 4.1). At present the southern part of this drainage basin is ice-free, however, the northern part contains a number of small glaciers and permanent snowbanks in cirques and cols on Mount Foster, Mount Gladstone, and Mount Disraeli (Figs. 2.1, 4.2). Windswept Valley is generally steep-sided and narrow, however, as with a number of valleys on this side of the Inlet, a bedrock sill runs across the valley near its mouth. The sill forms a dam, having a steep seaward slope and a gradual landward slope, which the river has truncated (Fig. 4.9).

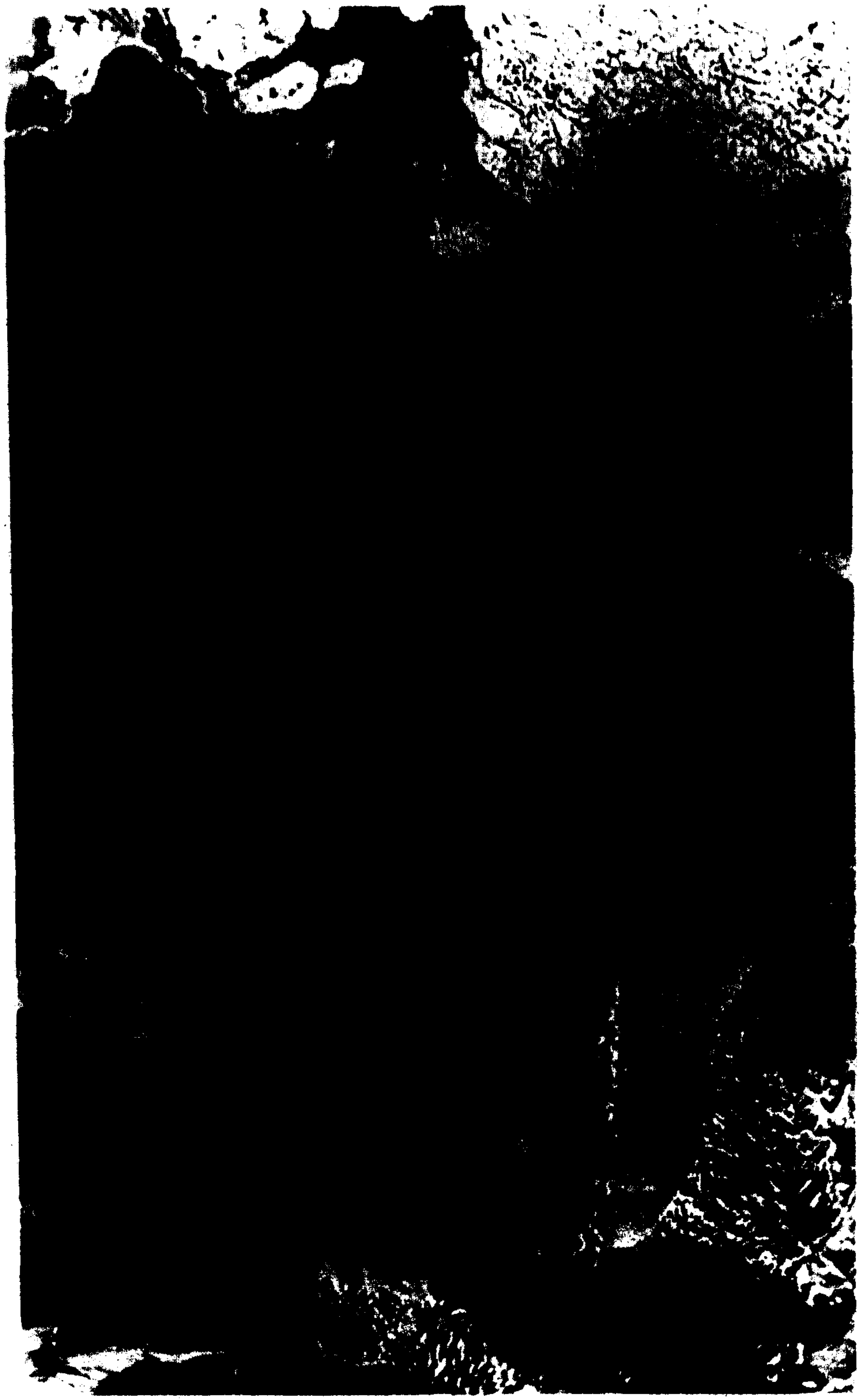
Three minor valleys, with steep gradients, lie adjacent to Windswept Valley and are also included in the description of this area. Two of the valleys, which lie to the NE are connected with the main valley by high passes (ca. 180 m, Fig. 4.9). The mouth of the third valley lies 1 km SW of the mouth of Windswept Valley.

4.3.7.2 Geomorphology and Stratigraphy

Benches cut in bedrock at elevations extending from just below the mountain summits to near the valley bottoms form prominent features along the valley sides. These benches may relate to scouring

Figure 4.9 Windswept Valley.

Note the river terraces within the valley (see Fig. 4.10). Arrows indicate some examples of numerous ice marginal channels along the valley sides. Also note: the location of the volcanic erratic (E); the alluvial fan base level at 114 m a.s.l. (X); and the section (S) described in the text. Radiocarbon dates 39 to 43 refer to Table 5.1; scale 1:40,000. (National Airphoto Library, A-16604-81)



by former ice marginal drainage systems which exploited the weak beds in sedimentary rock. Some support for this is given by isolated gravels, perched on the valley sides, and in particular, by a flat-topped lateral moraine nested on the steep valley side at 343 m (ca. 5 km upvalley). This moraine segment forms a continuum with adjacent bedrock benches, suggesting a common margin of a former valley glacier.

Evidence of greater ice cover in this area is indicated by an erratic of well rounded silicious tuff which was found near the summit of a minor mountain at 383 m, 6 km NW from the mouth of Windswept Valley (E, Fig. 4.9). Its provenance is resticted to small outcrops of lower Paleozoic volcanics S of Mount Rawlinson, and NE of Mount Frere (Fig. 2.2), ca. 40 km to the SW (Trettin; pers. comm., 1981). The implication is that at some time in the past a major ice sheet flowed in a NE direction parallel to the Inlet. Furthermore, several mountain barriers greater than 1070 m along this route must have been surmounted by the ice in order for the erratics to reach this site. The age of this advance will be discussed later.

The lower slopes of Windswept Valley are covered by colluvium, but less steep areas have till veneers. Thicker accumulations of till form minor moraine ridges, usually at the confluence of tributary and main valleys. Moraines with sharp crests and a fresh appearance also occur near the mouths of vacant cirques. Striated bedrock also occurs on the floor of Windswept Valley.

Windswept Valley contains a number of coarse gravel river terraces. The largest of these occurs at the junction with the tributary valley that divides Mount Gladstone and Mount Foster (Fig. 4.9). The uppermost terrace on the NE side of the Windswept river grades from 120 to 117 m, over a distance of 238 m. Farther downvalley, on the SW side, terrace remnants lie at 112 m and 107 m. Still farther downvalley a major remnant descends from 98 m to 86 m over a distance of 600 m. In the lowermost course small remnants also occur at 72 to 74 m and at 50 m. The 72 m terrace on the NE side of the river is composed of frost churned gravel and silty sand which contains Portlandia arctica, 2.5 km inland. Finally, the mouth of Windswept Valley contains a single delta terrace made of coarse sand and gravel with an outer lip at 25.5 m. A plot of the terraces described above is shown in Figure 4.10, and their discussion follows in a subsequent section.

The lower course of Windswept Valley also contains steeply dipping alluvial fans. A prominent break in slope occurs on the fan deposits at ca. 114 m a.s.l., 3 km from the sea. Gullies cut to bedrock exhibit thick accumulations of alluvium at this level, and two phases of gradation are evident; one to the upper level (114 m), and the second to the valley floor. In addition, lateral moraines descending from a tributary valley also terminate at approximately 114 m (Fig. 4.9) and this elevation could mark the marine limit within the valley.

As noted, the mouth of Windswept Valley is blocked by a

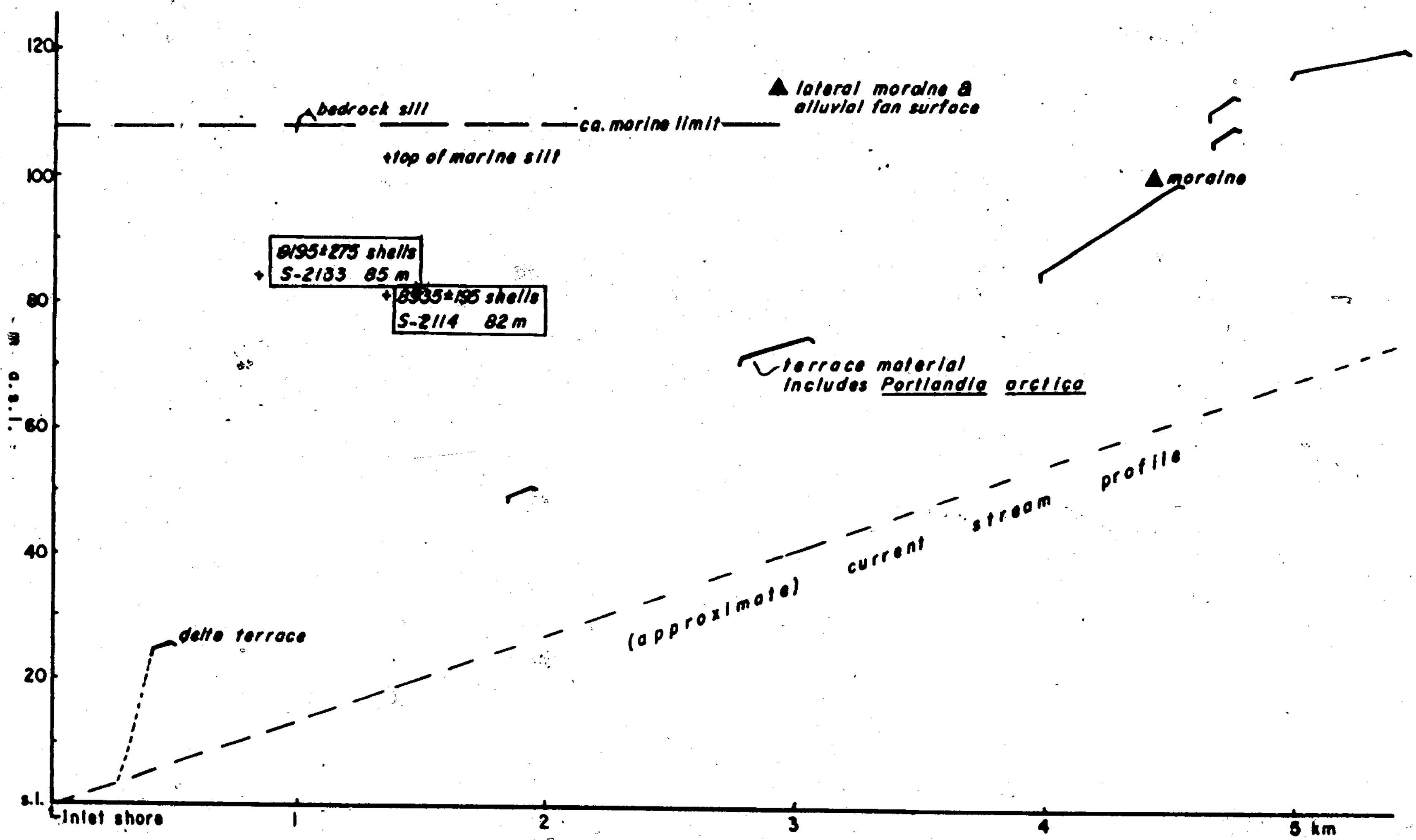


FIGURE 4.10 Terraces [illegible] indswept River

bedrock sill. The bedrock in this area is composed of extremely frost shattered shale and mudstone. A number of bedded silt outliers lie on the landward flank of this sill up to 102 m. Scattered shell fragments of H. arctica, collected on the surface at 81 m (site 40) dated 8935 ± 145 BP (S-2114). Further upslope, at ca. 110 m, the silts become thoroughly mixed with angular gravel and frost shattered bedrock. In some areas pockets of relatively undisturbed silts overlie outcrops of extremely frost shattered bedrock. Clearly, the bedrock was weathered before silt deposition occurred. No marine limit terrace was found in this area, but above ca. 110 m the silt matrix diminishes within the mixture of locally derived rock debris. On the seaward side of the sill, colluviated silts at 84 m (site 41) contain whole halves of H. arctica which dated 9195 ± 275 BP (S-2133). No shells were found within the single delta terrace that lies at 25.5 m, however surface shells from a gully cut into the delta dated 7250 ± 365 BP (S-2134).

Striae were found along the shore of the Inlet between the mouth of Windswept Valley and the minor valley 1 km SW of it. The bedrock exposures occurred where recent sea ice action removed the overburden. The orientation of the striae is with the long axis of the Inlet (SW-NE) and may be the product of the sea ice.

The adjoining valley, 1 km SW of Windswept Valley, contains lateral moraines with maroon conglomerate erratics high on the valley sides. The moraines descend to 130 m and 117 m at the seaward end of the valley (near site 39, Fig. 4.9). On the S side, the slope below the moraine has a series of descending mounds and swales made of angular gravel followed by a beach at 95 m. Whole valves of Hiatella arctica exposed by frost action in the beach gravels (site 39) dated 8605 ± 260 BP (S-2107), and these provide a minimum age for the beach. The highest marine silts in the area were found on the N side of the river at 43 m, and these are generally mixed with rock debris. A prominent delta terrace also lies at this elevation.

The two adjacent valleys, to the NE, connect to Windswept Valley via high passes (Fig. 4.9). One of the valleys contains striated and sculptured bedrock indicating an eastward ice flow direction which diverged from Windswept Valley. Weathered, maroon, conglomerate erratics were found near the summit of the pass, and also at the mouth of the valley, in a flat-topped moraine just above a steep scree slope. The top of the moraine lies at 121 m and surface clasts, including a few chert pebble conglomerates, appear weathered with pitted and etched surfaces.

The mouth of the second valley to the NE is flat-lying and contains gullied marine silts overlying conglomeratic sandstone bedrock at ca. 30 m. Where recently exposed, the bedrock is smooth and possibly ice-molded, but no striae were found. Outliers of silt could be traced upslope where they become admixed with the underlying shattered rock. Here the underlying bedrock and the outcrops are very weathered. A silt pocket at 85 m (site 43) contains abundant shell fragments of H. arctica and M. truncata which dated 8345 ± 140 BP (S-2113). Upslope an accumulation of diamicton forms a hummocky terrace at ca. 116-117 m. The mouth of the second valley is partially obstructed by a large, flat-topped ridge made entirely of gravel, lying at the angle of repose. The upper surface of this ridge occurs at 111 m, indicating an accumulation of ca. 80 m. A broad channel lies along the landward side of this ridge which, in turn, is bounded on the upslope side by a terrace comprised of angular debris (ca. 126 m), followed by frost shattered bedrock containing occasional conglomerate erratics (Fig. 4.9).

Owing to the steep and narrow nature of the valleys in this area, there are few exposures of anything other than colluvium. Exposures along upper Windswept Valley exhibit unconsolidated rock debris of various sizes, and till is not apparent. The major river terraces are comprised of more rounded, coarse sediment overlying bedrock. Further downvalley, the marine silts mantle very weathered bedrock, although at higher elevations the silts are thoroughly mixed with rock debris.

A large section was found in the minor valley 1 km SW of Windswept Valley. The exposure extends ca. 75 m and is ca. 25 m thick (S, Fig. 4.9). The base of the section exhibits bedrock overlain by silty colluvium, containing many blocky clasts, which in turn, is overlain by stratified silts containing occasional dropstones. The silts are overlain by a 1 m thick bed of very fine sands with massive bedding, and these, in turn, are overlain by silts containing a large volume of angular rock debris. Finally, the sandy gravels comprising the 43 m delta terrace form the uppermost unit in the section.

4.3.7.3 Interpretation

The evidence presented from Windswept Valley depicts several degrees of glaciation. The most extensive of these is marked by the erratic of volcanic tuff deposited by a former ice sheet which inundated mountain summits. This, however, is the sole evidence for such an advance. Furthermore, the configuration of the former ice sheet produced flow patterns that are dissimilar to those of the contemporary ice caps, even if they were a few hundred metres thicker. The flowline indicated by the provenance of the erratic is parallel to the Inlet, and it seems surprising that a major re-entrant such as Clements Markham Inlet did not divert this flow. Nevertheless, recent computer modelling has indicated that a major ice sheet, grounded along the Continental Shelf of northern Ellesmere Island to 100 m below sea level, would produce a weak ice stream crossing the high terrain just NW of Clements Markham Inlet and would explain the occurrence of this erratic (Fisher, pers. comm., 1983). Further evidence of such an extensive ice cover may be provided by the chert pebble conglomerate erratics which are found within some of the moraines in Windswept Valley and are also reported throughout northern Ellesmere Island (Christie, 1967). However, these rocks outcrop on the SE side of Clements Markham Inlet and S of Markham Fiord, and may not have required a large ice sheet to deposit them (Fig. 2.2; Mayr et al., 1982).

The second most extensive glaciation is recorded by moraine remnants, striae, and a series of high-level ice marginal channels. These features were caused by a local valley glacier system. Downwasting and upvalley retreat of these valley glaciers resulted in the nesting of successive landforms, upvalley to the local cirques. Although this succession exists it should be cautioned that these features may not necessarily relate to deglaciation from the same glacial cycle. For example, the style of glaciation may have been similar during several glacial cycles, only their relative extent could have differed. Certainly, time-breaks are indicated by the relative extent of weathering between: the moraines near the cirques, the moraines further downvalley or upslope, and the isolated erratics within weathered bedrock. However, these differences are only relative and greater resolution cannot be obtained. Moreover, changing lithologies and microenvironments in the area produce inconsistencies which hamper the temporal quantification of weathering zones.

Evidence indicating the maximum extent of these valley glaciers into Clements Markham Inlet was not found. Nevertheless, when entering the proglacial sea that inundated the valley mouths, the glaciers may have calved or flowed into ice shelves. Calving certainly occurred during the deglacial stage as indicated by the section 1 km SW of Windswept Valley described above. Here, the lowermost silty colluvium is interpreted as the product of slope instability brought on by the marine transgression early in the glacial cycle. This material is of upslope origin and clasts are angular and blocky, with no glacial modification indicated. After this initial period, more quiescent, deep water conditions prevailed during which time silt was deposited in horizontal beds. A gradual increase of rock debris within the stratified silts may indicate the approach of a glacier producing large volumes of ice-rafted debris. Unlike the lowermost unit, this rock debris contains occasional erratics and faceted clasts. Finally, uppermost sandy gravels mark the postglacial regression of the sea. The high volumes of ice-rafted debris described above may indicate the former presence of an ice shelf or floating glacier tongue. Elsewhere, especially near the mouth of Windswept Valley, surface silts of marine origin mixed with rock debris are widespread on the upper slopes. However, this silt cover is thin and the relative amount of subsequent, subaerial mixing with the underlying weathered rock is not known.

Additional evidence for floating, or at least calving glaciers in the valley 1 km SW of Windswept Valley is provided by the 95 m beach and the bordering lateral moraine at 130 m. Silt deposits in Windswept Valley indicate that the marine limit in this area is above 103 m. Because the highest water level indicated in the SW valley is only 95 m, a glacier occupying the valley must have prevented the sea from making contact with the land when the sea stood higher. It should be noted that it is not a coincidence that the lateral moraine lies at 130 m. Given the elevation of bedrock in the valley floor, the glacier that formed the lateral moraine must have been at least 111 m thick. Because the average density of glacier ice is 0.88 gm cm^{-2} (Paterson, 1969), about 98 m of water was required to float the snout of this former glacier. Hence, the ratio of ice thickness to water depth required to float the glacier is clearly demonstrated at this site. Moreover, because calving is more effective than melting for disposing of glacier mass, a stillstand will occur when a glacier attempts to advance, or retreat, beyond its grounding line while the glacier adjusts to the new ablation regime. Lastly, the ca. 8.6 ka BP age obtained for the beach also provides an estimate for the age of the lateral moraines within this valley.

The older shell dates reported from this area provide minimum estimates for local deglaciation. Nonetheless, the dates indicate that the mouth of Windswept Valley was ice-free ca. 600 years before the smaller SW valley (although the large standard errors may place the events closer together). In addition, the difference of 260 years on shell dates from the seaward (9195 BP) and the landward (8935 BP) slopes of the bedrock sill (crossing the mouth of Windswept Valley) may indicate a gradual retreat over several hundred meters. After this, rapid retreat is indicated by marine fauna found 2.5 km upvalley at ca. 72 m (estimated at >ca. 7.8 ka BP, Ch. V). It is surprising that marine bivalves (P. arctica) could successfully invade such a narrow valley, far inland where freshwater input was likely substantial. Of course, if the marine limit (ca. 114 m) initially extended further upvalley, as shown in Figure 4.10, these dates would not reflect glacial retreat.

The marine limit within Windswept Valley is difficult to determine because colluvium from rapidly weathering bedrock has buried or destroyed any strandlines above 103 m. Silty rock debris is found above 103 m, however, this may have originated from subaerially weathered mudstone in the area. Alternatively, the gravel terraces up to 120 m within Windswept Valley may mark the marine limit. However, when all the terraces are plotted within the longitudinal profile of Windswept Valley (Fig. 4.10), it is apparent that they may represent only a single former river profile graded to a sea level of 25.5 m. Alternatively, their ages may be dissimilar so that they represent remnants of successively lower river gradients which developed as the sea regressed downvalley. Lastly, if the glacier within Windswept Valley had retreated upvalley of the fossiliferous site, ca. 2.5 km inland, during the marine limit stage, the abrupt termination of descending lateral moraines, together with the high base level of alluvial fan gradation ca. 114 m, may indicate the marine limit at ca. 114 m (Fig. 4.10).

4.3.8 The Outermost NW Shore of Clements Markham Inlet to Cape Colan

4.3.8.1 General Description

This area is characterized by the steep, SE facing flanks of Mount Foster (ca. 920 m) and a smaller unnamed mountain 7 km to the NE (650 m, Fig. 4.2). The steep seaward slopes are interrupted by several short valleys leading from small, ice-free cirques a few kilometers inland. A broad valley and bay separate these uplands from Cape Colan, which forms the NW entrance into Clements Markham Inlet (Fig. 4.11). Cape Colan has a low summit

Figure 4.11 Cape Colan.

Note the minor moraine near the marine limit at 93 m a.s.l.. Also note the undulating surface of the sea [illegible] shelf along the northern coast. Radiocarbon dates 44 and 45 refer to Table [illegible] scale 1:30,000. (National Airphoto Library, A-16604-85)

(ca. 160 m) which descends abruptly on the E side to a lowland. The steep E slope contains a number of incipient cirques occupied by permanent snowbanks.

4.3.8.2 Geomorphology and Stratigraphy

The cirque floors on Mount Foster are generally covered by coarse till and colluvium and minor moraine remnants are found near the valley mouths. Small, raised deltas are also found at the valley mouths, however, distinct higher terraces are scarce. The largest delta, SE of Mount Foster, has the highest terrace in this area at 99 m. It is comprised of very coarse angular debris, including large subrounded boulders. An adjacent lower delta occurs at 73 m. The valley mouth E of Mount Foster contains isolated deposits of silty gravel with large clasts and rudimentary foreset bedding. Relative sea level could not be determined at the latter site because subsequent regrading to lower levels at 31 and 35 m had taken place.

The small unnamed mountain NE of Mount Foster has lower slopes that are slightly less steep, consequently more sediment accumulated here. The upper seaward slopes are covered by coarse, angular rock rubble containing infrequent subrounded erratics. The colluvium forms a bench (rock glacier?) at ca. 106 m. Marine silts lie on the mountain side up to 87 m (near site 44, Fig. 4.11). As well, a small horizontally stratified silt outlier abruptly overlies smooth, striated bedrock in a level area at 84 m. The striae show a SW to NE orientation, approximately parallel to the long axis of the Inlet (Fig. 4.11). The underlying limestone must weather rapidly because striae are absent and fretting is prevalent where the surface is exposed, a few centimeters from the covering silts. Abundant shells of H. arctica, in growth position, were obtained from the silt outlier (site 44) and these dated 8905 ± 150 BP (S-2104). No relative sea level for this sample could be determined (87-95 m). A major silt, sand and gravel bench, immediately to the S, lies at 84 m.

An outlying ridge of extremely frost shattered shale, ca. 1 km to the NE of the above site, also has a silt mantle. At its upper limit the silt is thoroughly mixed with angular bedrock fragments (95 m). Conversely, the seaward slope bears a well developed beach at 78 m, and pockets of undisturbed silt lie near present sea level. The contemporary littoral zone is extensively bulldozed by sea ice movement from the ENE. This process of sea ice abrasion was also found on older raised shorelines several meters higher.

Cape Colan represents the northernmost part of the study area. Most of the gentle slopes are covered by soliflucting rock debris, whereas more stable surfaces contain patterned ground and desert varnish. Numerous permanent snowbanks occupy low-elevation hollows. Limestone outcrops near the summit exhibit ice-molded forms which have subsequently undergone intensive frost shattering, surface scalloping, and joint-block heaving. Although no striae were observed here during this study, Christie (1967) reported striae having a NW-SE orientation in this area. Nevertheless, this worker felt the gross morphology of the ice-molded forms indicate a former flow to the NNE.

The steep slope, E of the summit at Cape Colan, contains several nivation hollows or incipient cirques. Near one of these features, an arcuate moraine emanates from the cliff face for a distance of 100 m at 93 m. The moraine is comprised of silty rock debris, similar to the surrounding material. Rare shell fragments were found just downslope, at 67 m, where silts with a high content of angular rock rubble contain whole valves of H. arctica and isolated bands of redeposited plant material (site 45). The shells dated 8550 ± 135 BP (S-2116). The unusually high content of rock debris does not decrease significantly in these silt bodies until ca. 1 km downslope from the moraine at 16 m.

Steep slopes and rapidly weathering bedrock characterize the outer NW coast of Clements Markham Inlet, consequently no till sections were found. Furthermore, where till was observed it occurs only as thin discontinuous veneers that indicate minor glacial deposition and insufficient sediment supply for subsequent delta building. Significant amounts of fine grained sediments were found only in the Cape Colan area. Here, the general stratigraphic sequence depicts glacial action, of some unknown age, preceding marine deposition which is followed by regressive deposits. The exception to this occurs where moraines are partially derived from the marine sediments.

4.3.8.3 Interpretation

The preceding descriptions indicate that the cirques on Mount Foster supported small glaciers, but their seaward extent is unknown. The lack of prominent raised deltas in this area could have been caused by two conditions. First, there may have been insufficent sediment input for delta building due to the small catchment areas, and second, very steep submarine slopes disperse incoming sediments so that delta building is inhibited, particularly during rapid postglacial emergence. Nonetheless, minor terraces were built during later times of decelerated emergence.

The date on fossiliferous silts NE of Mount Foster suggests that local deglaciation occurred just prior to ca. 8.9 ka BP. The abruptness of the contact between striated bedrock and the silts indicates that deglaciation was rapid. However, the apparent age of this glacial abrasion, coupled with the orientation of the striae, is enigmatic. Indicated ice flow is parallel to the Inlet and suggests a major trunk glacier flowing down Inlet, just before the marine incursion ca. 8.9 ka BP. If this is the case, the presence of such a glacier is contradicted by older dates (ca. 9.8 ka BP) on in situ shells found at numerous sites further up the Inlet. A chronological hiatus may explain the incongruity, but it is unlikely that the underlying striae would be so well preserved if they are much older than the silts. Although, if the site remained below sea level, the striae, could be indefinitely preserved. The most likely explanations are that: the flow was from a local cirque glacier whose seaward flow was diverted N by topography; or that these striae were due to the geomorphic action of an ice shelf of either glacial or sea ice origin. In either of these cases, an earlier marine inundation up the Inlet is not prevented and therefore these explanations do not contradict the older dates previously discussed.

The sculptured bedrock forms on the summit of Cape Colan also suggest overriding by ice, however, as with other high-elevation features in Clements Markham Inlet the absolute age of this event is presently indeterminable. What is known is that the lowlands were inundated by the sea at least by ca. 8.6 ka BP and probably much sooner (see Ch. V). Very high amounts of rock debris within the marine silts indicate frequent ice-rafting either from glaciers or an ice shelf environment. Moraines deposited by ice from incipient cirques bordering the marine limit at Cape Colan could either have been built into the marine limit sea (ca. 95 m) by niche glaciers, or they postdate the marine transgression entirely, being neoglacial in age. In the latter case, it would mean that no evidence of pre-emergence ice margins was found in the Cape Colan area other than an isolated bench of colluvium at 98 m. In any event, plant debris within the local silts suggests ice-free areas existed by at least 8.6 ka BP.

4.3.9 Piper Pass

4.3.9.1 General Description

The preceding sections described the NW coast of Clements Markham Inlet, whereas Piper Pass is the first area described on the SE coast, beginning at the mouth of the Clements Markham River. The N end of Piper Pass forms part of the lowland occupying the head of the Inlet. To the S, it becomes a narrow trough <2 km wide with 1000 m of relief where it cuts through the United States Range to join the Lake Hazen Basin (Fig. 1.1). At present Piper Pass is free of major glaciers, however, several smaller glaciers, damming three lakes on the valley floor, extend from mountain ice caps on either side of the trough (Figs. 2.1, 4.2).

4.3.9.2 Geomorphology and Stratigraphy

Most of Piper Pass is occupied by a contemporary sandur and several low-level terraces. Three larger side glaciers, flowing from the ice caps, completely obstruct the N flow of the drainage, consequently forming the ice-dammed lakes. Several former water levels are indicated around these lakes and they are probably related to stages of glacial expansion or base level changes caused by changes in sea level. However, these distinct lake levels were observed only from aircraft and no ground observations were obtained because the lakes occur in central and southern Piper Pass, outside the field area. The side glaciers have formed trimlines and moraines a few hundred meters from present margins. Numerous rock glaciers, particularly along the W facing valley side, have complex forms and may be, in part, modified lateral moraines. Furthermore, thermokarst features on many valley-side deposits attest to high ice contents. Coarse grained alluvial fans are also common in the area and a few open system pingos were observed on them.

The north end of Piper Pass is mantled by thick deposits of marine silts up to 85 m. These glaciomarine silts are extensively gullied and in places have a sparse lag of surface debris. A cover of coarse rock debris, including chert conglomerate erratics, mantles bedrock up to 160 m. Striations in the north-central part indicate a N flow aligned with Piper Pass (Fig. 4.12).

Active sand wedges, ventifacts, and desert varnish attest to a harsh contemporary environment in this area. The marine silts also exhibit distinct N-S flutings and yardangs which indicate strong winds from Piper Pass. Athough such winds were experienced during the 1979-1980 and 1981 field seasons, numerous fluted areas have been subsequently truncated by major gully erosion suggesting that the large flutes were formed long ago, possibly during initial emergence.

Figure 4.12 North End of Piper Pass.

Note the kettled, glaciofluvial terraces built into a 103 m a.s.l. marine limit (upper right). Striae indicate a northward flow of ice from Piper Pass ( ). The arrow points to the till exposure described in the text. Radiocarbon dates 4 to 6 refer to Table 5.1; scale 1:60,000. (National Airphoto Library, A-16603-112)

A small valley, trending almost perpendicular to Piper Pass and parallel to the Inlet, connects the NE end of the Pass to the Arrowhead Delta complex (Fig. 4.12, upper right). The flat floor of this valley is composed of coarse sandy gravel with a number of kettle lakes. Large terraces of the gravel lie within Piper Pass at the SW end of this valley at 103 m and 96 m. Marine silts are exposed on the flanks of these terraces, however their exact stratigraphic relationship to the terraces is difficult to determine (discussed later). A number of silt outliers form distinct benches at lower elevations on the steep slopes of these terraces. The flat upper surface of one of these silt outliers contained abundant <u>in situ H. arctica</u> at 67 m (site 6) which dated 7860 ± 330 BP (UQ-262).

Two main sections were discovered in the N end of Piper Pass. The first of these occurs downslope from the 103 m gravel terrace. Here, the base shows ca. 3 m of sand exhibiting cross-stratified and dipping plane beds which have subsequently been faulted. The sands are abruptly overlain by generally barren silts whose beds dip 5^{o} to the SSW. The silts are up to 10 m thick and some upper beds are fossiliferous, containing several species of bivalves. As reported above, a sample of these shells dated ca. 7860 BP. These beds also contain intercalated mats of hair-like organic fibre a few centimeters thick. The stratigraphic relationship between the silts and the abutting gravel terraces is difficult to establish because loose material from the terraces obscures the contact. Nonetheless, it is likely that these terraces represent a coarsening towards the valley sides due to increased proximity of the ice.

The second section was observed where the river flowing from Piper Pass is diverted NE by a bedrock ridge as it enters the Inlet (Fig. 4.12). The cut-bank exhibits a 1 to 2 m layer of till overlying planed-off limestone and gypsum bedrock which, in turn, is overlain by stratified marine silts. This sequence is similar to the one described at Till Island (sec. 4.3.1), and the NE flow indicated by its fabric could be related to ice from either the Clements Markham River valley or Piper Pass. However, no formal fabric analysis was undertaken at this site.

4.3.9.3 Interpretation

Erratics, striae, and till in northern Piper Pass indicate the presence of a former valley glacier flowing northward. Whether this ice merged with the glaciers in the Clements Markham River valley is not known, as medial moraines are absent. The retreat of the glacier from northern Piper Pass is marked by glaciofluvial terraces which graded into the marine limit sea at 103 m. During this time (ca. 7860 BP or earlier), a smaller glacier occupied the aforementioned cross valley at the NE end of Piper Pass. No evidence of the terminal position of the main glacier in Piper Pass was discovered, but fine grained marine deposits, 5 km up Piper Pass, suggest that the terminus was beyond this point during the marine limit stage.

The overall stratigraphic sequence generally depicts rapid recession from the area of confluence of Piper Pass and the Clements

Markham River valley. However, in the area of the marine limit glaciofluvial terrace, the marine silts are underlain by cross-stratified sands rather than till. Given this fining-upward sequence, several interpretations can be proposed using the model discussed in Chapter III. The first interpretation suggests that the fining-upward sequence reflects a gradual rise in sea level up to the marine limit. However, this condition is inconsistent with glaciers blocking a high sea from this area, because a rapid transgression would follow ice retreat in such a case. The second interpretation is that the stratified sands could have been deposited during the initial transgression, early in the glacial cycle, when the ice front was still a considerable distance upvalley, and therefore a hiatus could exist between the sand unit and the overlying silt unit. In fact, the sharp boundary between the two units may be erosive, and furthermore, faulting within the sands may have been caused by subsequent overriding by ice. A third interpretation is that the sands were deposited in a proglacial, subaqueous environment by a center of meltwater outflow which later shifted and allowed silt deposition during the subsequent quiescent interval. In summary, the proglacial environment under which the sands were deposited may have been marine, glaciomarine, or even glaciolacustrine. As noted, contemporary proglacial lakes are common within Piper Pass. Moreover, Christie (1967) suggested the existence of a former proglacial lake at the S end of Piper Pass, 30 km NE of Lake Hazen. The existence of a former proglacial lake at the NE end of Piper Pass, therefore, remains a possibility if it was dammed by the main Inlet ice.

4.3.10 Arrowhead Delta

7

4.3.10.1 General Description

The Arrowhead Delta is a complex delta system found immediately to the E of Piper Pass (Fig. 4.13). The Arrowhead valley lies parallel to Piper Pass and is headed by a major outlet glacier from the Grant Ice Cap 11 km from the coastline (Fig. 4.2). The mouth of this valley is connected to Piper Pass via the cross valley described in the previous section, whereas a NW trending trough connects the head of the valley to the Breakthrough Delta (next section). Raised marine sediments at the mouth of the Arrowhead River were first described by Christie (1967) and a gravity station established there during 1957/58 is still shown on the 1:250,000 topographic map of Clements Markham Inlet.

4.3.10.2 Geomorphology and Stratigraphy

The overall distribution of sediments in this delta (Plate 4.7) is similar to other deltaic sequences described. An outer belt, of thick, horizontally bedded and gullied, marine silt ca. 1.5 km wide represents bottomset deposition (Plate 4.8). The silts are, in turn, overlain by proglacial, sandy gravel terraces at higher elevations. Near the delta apex numerous kettles and hummocks suggest an ice-contact glaciofluvial origin for the gravels. These terraces generally grade down to a lip elevation of ca. 103 m. Subsequent river degradation has incised both the upper terraces and thick silt

Figure 4.13 Arrowhead Delta.

The upper delta lies at the confluence of the Arrowhead Valley and the cross valley to Piper Pass (upper right corner of the photo). Note the kettled, ice contact topography with numerous former meltwater channels. Kame terraces and outwash surfaces grade down to the marine limit at ca. 103 m a.s.l.. Several rock glaciers (RG) have developed below steep cliff-faces and may in part be lateral moraines. Also note the descending delta terraces and the gullied bottomset silts. Radiocarbon dates 9 to 16 are described in Table 5.1; scale 1:30,000. (National Airphoto Library, A-16607-103)

Plate 4.7 The Arrowhead Delta. The Arrowhead glacier occupied the right of the photo and emitted sediment into a high sea of 103 m a.s.l. at ca. 8,000 BP. Coarse terraces formed near the ice front, meanwhile bottomset silts and fine sands settled out in deeper, more distal areas. Several coarser terraces prograded over the fines as sea level dropped. Also note the rock glaciers which formed along the base of the steep cliffs in the upper center. Some of these may be rock glacierized lateral moraines (Fig. 4.13).

Plate 4.8 Thick deposits of bottomset silts in the Arrowhead Delta are dissected by asymmetric gullies up to 40 m deep. North is to the left of the photograph.

deposits to form several lower delta terraces at 70, 50 and 33 m (Fig. 4.13). The Arrowhead Delta has been further altered by a secondary stream to the W. Moreover, buried bedrock ridges and protuberances have redirected drainage and formed gorges during sea level lowering. The end result is that a complex series of smaller deltas lie inset and superimposed on the initial deposits.

The area upslope of the delta apex contains till veneers, coarse glaciofluvial terraces, and lateral moraines ca. 120 m. Many of the moraines have been rock glacierized.

Stream cuts on the W side of the delta complex have exposed a basal unit of cross stratified sands and gravels, 10 m thick, overlain by 3 m of diamicton containing many gravels similar to the underlying unit, plus occasional striated clasts. The diamicton is interpreted as a till, but it was not exposed for more than 50 m laterally in section. Ten meters of rhythmically bedded silts comprise the uppermost unit in the section. Seaward, along the same cut, only cross stratified sands underlie the upper silt unit, whereas further seaward, the silts are underlain by horizontally laminated fine sand. Like the other areas described, the stratigraphically lowest silts are barren in this area. However, some stratigraphically low, fossiliferous silts were found on the E side of the Arrowhead River at 60 m (site 16); as well as on the W side at 75 m (site 9). These _in situ_ shells of _H. arctica_ and _M. truncata_ dated 7965 ± 65 BP (SI-4316) and 7900 ± 335 BP (UQ-263) respectively.

Younger silt bodies are inset into the seaward periphery of the Arrowhead Delta. One such deposit originally reported by Christie (1967) contains detrital plant beds (site 13) which dated 6400 ± 60 BP (SI-4314; Stewart, 1981). Stewart (1981) and Stewart and England (1983) discussed the environmental implications of the various fossil species and they suggested that ameliorated conditions occurred at this time. Hiatella arctica in growth position were collected from bottomset sands, overlain by foreset sands grading to 39 m (site 12), and dated 6050 ± 300 (UQ-277). This sample is considered to provide the maximum age estimate for the 39 m relative sea level. The W side of the delta contained the oldest driftwood yet found within Clements Markham Inlet. These samples, found on the surface of dissected marine silts at 93 and 83 m (sites 10, 11), dated 8545 ± 110 BP (S-2210) and 8915 ± 115 BP (S-2211), respectively (Stewart and England, 1983).

4.3.10.3 Interpretation

Ice-contact glaciofluvial sediments and moraines indicate that the Arrowhead Glacier reached Clements Markham Inlet. During this stage a tongue of ice also flowed NE from Piper Pass, through the cross valley, and merged with the Arrowhead Glacier. The limits of these glaciers are unknown, but the old driftwood suggest that this part of Clements Markham Inlet was ice-free by at least 8.9 ka BP. However, deglaciation of the delta zone did not occur until 8 ka BP, as at the mouth of Piper Pass. While the narrow mouth of the Arrowhead Valley was occupied by the glacier, the marine limit formed downvalley at 103-104 m. Stagnant ice within the cross valley also produced large volumes of glaciofluvial sediments which prograded into the marine limit sea.

Following the establishment of the marine limit, postglacial emergence continued and was accompanied by discrete pulses of deltaic progradation. Paleoenvironmental fluctuations during this interval are discussed in detail by Stewart (1981) and Stewart and England (1983).

4.3.11 Breakthrough Delta

4.3.11.1 General Description

Breakthrough Delta lies 7 km to the E of the Arrowhead Delta complex (Fig. 4.1). It is the last major raised delta system along the SE coastline of Clements Markham Inlet. The delta lies at the mouth of a valley network which presently drains a glacier emanating from the Grant Ice Cap and a number of smaller cirque glaciers, ca. 9 km inland. One of the main tributaries is connected to the Arrowhead Glacier Valley via an upper col. Consequently, given the past expansion of the Arrowhead Glacier, additional ice would have entered the Breakthrough drainage basin (Fig. 4.2). The lower part of the drainage basin becomes progressively more flat-lying and open, allowing for the accumulation of sediment during the last transgressive/regressive cycle.

4.3.11.2 Geomorphology and Stratigraphy

Till veneers and minor moraine ridges characterize the flat lying areas of the Breakthrough Delta region above 120 m a.s.l.. Several subdued moraines and ridges of ice-contact gravels lie close to the apex of the delta proper. Also, small arcuate till ridges, and till veneers are found adjacent to the delta at lower elevations, some extending below the marine limit (Fig. 4.14).

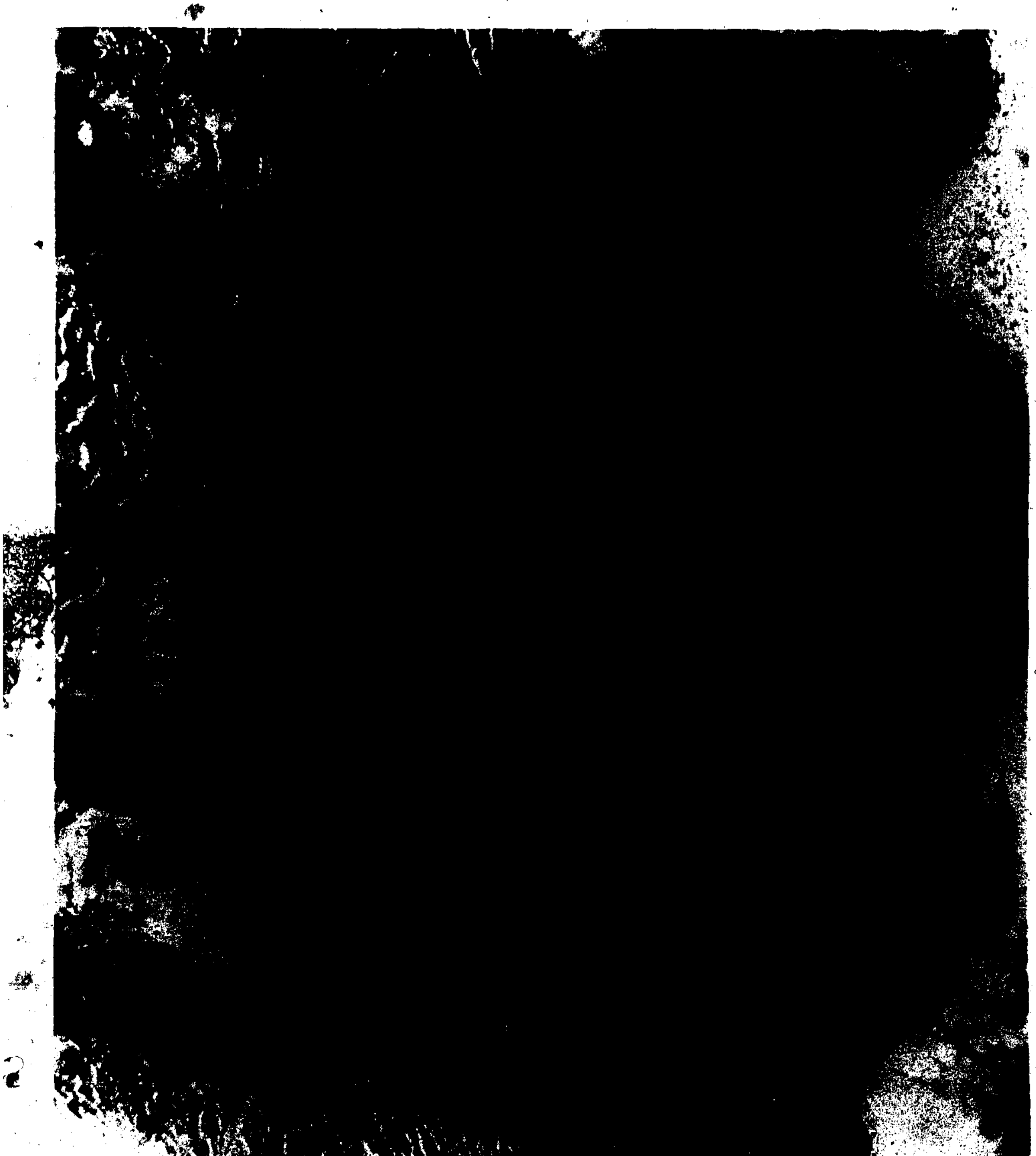
Breakthrough Delta is composed of a series of descending gravel terraces which have prograded over deep-water silts. Silt outliers are also found in the periphery of the delta where modern fluvial erosion has isolated them. Major gravel terraces occur at 117.5 m and 70 m. Abundant, disarticulate *H. arctica* valves were found in a silt pocket at 100 m, about 200 m downslope from the uppermost terrace (site 17). These shells dated 9745 ± 255 BP (S-1981). Subsequent, lower terraces are associated with a seaward shift in the apex of the delta, as the river progressively downcut through the bedrock.

4.3.11.3 Interpretation

The distribution of the moraines in the flat-lying areas below the junction of the main tributary valleys, indicates that the area was occupied by a number of glaciers that coalesced. At some point in time, certainly before 9745 ± 255 BP (S-1981), the glaciers extended ca. 6 km beyond their present margins. During this stage, an

Figure 4.14 Breakthrough Delta.

Widespread galciofluvial and till veneers are found in the upper-central part of the photo. The marine limit delta formed at 117 m a.s.l., during which time the main channel was to the right of present (arrow). The dashed line represents the approximate glacier margin during the time of marine limit formation. Radiocarbon dates 17 and 18 refer to Table 5.1; scale 1:30,000. (National Airphoto Library, A-16604-77)



ice contact marine limit delta was built into a relative sea level of ca. 117 m. Subsequent stages of glacier retreat are documented by minor moraine remnants at various positions upvalley (Fig. 4.15). The moraines shown are thought to have been produced by glaciers which flowed into the Inlet from the side valleys, rather than by main ice in the Inlet, because the delta was prograding into the Inlet (hence the Inlet was ice-free) at the same time side glaciers were still occupying the mouths of the side valleys.

4.3.12 The Outer SE Coast of Clements Markham Inlet

4.3.12.1 General Description

The SE coast of Clements Markham Inlet, NE from Breakthrough Delta, to Hamilton Bluff, is characterized by ca. 35 km of steep mountain flanks truncated by two major valleys and a number of smaller ones headed by cirques (Fig. 4.2). The two major valleys; one 2 km E of the Breakthrough Delta and the other 7 km SW of Mount Beverley, contain sandar which drain a number of glaciers occupying the NW flanks of the United States Range (Fig. 2.2). The first valley, adjacent to Breakthrough Delta, is formed by the junction of three main tributary valleys ca. 3 km from the sea. In the second valley, the sandur terminates in a lake which, in turn drains into a large bay on the central, SE shore of the Inlet (Fig. 4.2). The coastline, from this bay to Hamilton Bluff, is bordered by several valleys leading to ice-free cirques. These valleys decrease in size down Inlet.

4.3.12.2 Geomorphology and Stratigraphy

Although the two main valleys which presently contain glaciers have most of the glacigenic sediments; even these do not have significant stratigraphic sections. Consequently it was not possible to find sufficient marine fauna for dating control.

In the major valley directly E of Breakthrough Delta, the upper reaches of the three tributary valleys contain large areas of local rock rubble, colluvium and thin, discontinuous till veneers. Marine sediments are found only downvalley where the three tributary valleys meet. These marine sediments are comprised of isolated gravel terraces at various elevations, and horizontally bedded silts abutting steep hillslopes. The mouth of the E tributary valley contains a flat-topped sand and gravel terrace at ca. 111 m and a number of subdued moraines lie just upslope. Below the junction of the tributaries, the valley bottom is dominated by a large outwash fan building into the sea. A steep cliff, forming the E boundary of the fan, is covered with colluvium, associated rock glaciers, and lateral moraine remnants. The lateral moraines have also become rock glacierized in several areas.

In the second major valley molded and striated bedrock was observed up to at least 350 m along the W side of the large bay. The striae are aligned parallel with the bay (NW-SE) and are formed on oxidized, orange-brown dolomitic sandstone and mudrock of the Imina Formation (Mayr, et al., 1982). Only the striated surfaces appear oxidized because freshly exposed, frost-rived surfaces are grey in colour. At approximately the same elevation on the opposite side of the bay a number of benches are cut into the side of Mount Beverley across the bedrock strike. Flat uplands and broad hanging valleys are covered by till veneers.

Concentrations of marine washed till and outwash separate the head of the bay from the large lake. Gravelly moraines, containing large boulders, lie on the valley sides bordering this lake (Fig. 4.2). The moraines grade downvalley from ca. 100 m to 90 m, terminating in the vicinity of till concentrations at the head of the bay. The lake also has a number of strandlines formed above it, and a prominent gravel terrace lies at ca. 90 m, 1 km upvalley from the lake. Further upstream, the valley is dominated by the contemporary sandur, with scree slopes, lateral moraine remnants, and outwash terraces along the valley sides.

The remainder of the SE coast of Clements Markham Inlet out to Hamilton Bluff, has only small deltas at some valley mouths, whereas, thin, discontinuous tills, mixed with high proportions of local rock rubble, occupy the highlands. Stream-cuts through the deltaic sediments show coarse, cobbly gravels whereas finer sediments are rare. The largest of these deltas lies 5 km W of Mount Beverley and at this site two _in situ_ shell samples were obtained from sandy gravel, foreset beds. A sample of _H. arctica_ obtained from an regraded 42 m terrace (site 38) dated 6595 ± 125 BP (SI-4317), whereas the other, obtained from foreset beds related to a relative sea level of at least 55 m (site 37), dated 6480 ± 130 BP (UQ-261). The two outermost deltas, near Hamilton Bluff had minor terraces and gravel accumulations at 55 and 42 m respectively. Neither of these terraces was found to be fossiliferous.

4.3.12.3. Interpretation

Moraines, ice marginal channels cut in bedrock, and high level striae indicate that major coalescent glaciers, emanating from the Grant Ice Cap, occupied parts of the SE coastline of Clements Markham Inlet, especially in the vicinity of the two major valleys described.

Although dating control is lacking, the major valley immediately E of Breakthrough Delta does contain a local marine limit terrace at ca. 111 m. This elevation suggests that deglaciation, to at least the junction of the three main tributary valleys, occurred sometime not long after the marine limit was established at the Breakthrough Delta where shells at 100 m, related to the 117 m sea level, dated ca. 9.7 ka BP.

Less is known about the major valley and bay SW of Mount Beverley. However, the striae indicate that a glacier, at least 350 m thick, occupied the mouth of the Bay at some unknown time in the past. During this stage smaller glaciers probably flowed from the upland cirques on Mount Beverley, coalescing with the main glacier. This stage of ice advance may correlate with the extensive advance indicated by the outermost moraines in the previously described valley E of Breakthrough Delta (Fig. 4.2). If this advance is older than ca. 9.7 ka BP, the striae have remained intact despite subaerial exposure for at least the duration of the Holocene. Although resistance to surface weathering may be a characteristic of the Imina Formation, it is noteworthy that striae on other lithologies within the study area are only found on recently exhumed surfaces. Because the striated surfaces have an orange-brown oxidation, recent exhumation of an old striated surface cannot be invoked here in order to explain their preservation. Furthermore, this ice molded rock is undergoing destruction under the present environment, and presuming that these conditions have prevailed for some time, these forms should have been obliterated. Moreover, if this destructive weathering has prevailed since deglaciation of the site the orange-brown surface oxidation would not have formed in the first place. Consequently, because the oxidized surface is present, the immediate environment after deglaciation may have been dissimilar to now, and/or the oxidation may have case-hardened the surface so that the striae are preserved, despite their age. A significant ice margin within the limits of the outermost advance is indicated by the moraines at 90 to 100 m, inland from the large bay. The glacier probably terminated on the seaward side of the lake at this time. Although several lake levels are indicated at lower elevations, the upper levels indicated by terraces at ca. 90 m are probably not lacustrine but related to marine incursion which followed ice retreat. The marine limit on the opposite shore of the Inlet suggests that the limit in this bay should be at least 110 m.

Evidence of glacial action along the outermost SE coast of Clements Markham Inlet is very limited and stratigraphic sections were not found. Till veneers are mainly comprised of local bedrock rubble, and are difficult to distinguish from colluvium. Sediment production, in general, was and still is very small so that deltas are limited. Raised deltas appear as non-graded cones that may represent submarine fans because, based on the elevation of the marine limit on the opposite shore of the Inlet, the marine limit sea did not register in this area. Furthermore, inhibited delta building during the time of the marine limit sea implies that the former glaciers, which would have supplied most of the sediment, must have also been small at this time. This contrasts markedly with the large deltas, and corresponding evidence of extensive glaciation, at the head of the Inlet. Like the NW shore of Clements Markham Inlet, evidence on the SE shore indicates progressively less extensive glaciation as one moves from the head of the Inlet to the outermost coast.

4.4 Ice Retreat Map

An integrated summary of the surficial geology (Fig. 4.2) and geomorphic descriptions (Sec. 4.3) within Clements Markham Inlet is provided by the Ice Retreat Map (Fig. 4.15). In the previous section, numerous ice-contact indicators such as kame terraces, moraines, meltwater channels (Plate 4.9), and perched deltas were identified within the study area. These features were mapped and correlated in order to approximate former ice configurations within the area. The resulting sequences of nested ice margins within the valley systems

Plate 4.9 Ice marginal channels near the head of the Inlet, along the northwest side. These nested channels dip down the Inlet to the right and mark the terminus of a trunk glacier which once occupied the head of Clements Markham Inlet. The Gypsum River delta lies to the right of the photograph.

suggest a pattern of glacier retreat from the Inlet shore to the contemporary ice margins. Although most of the inner margins remain undated, ages of the ice margins along the coastline can be estimated from the dated fossiliferous marine deposits.

4.4.1 Construction

Constructing the Ice Retreat Map involved plotting the known ice marginal features derived from airphotos and field work on a base map. Secondly, all adjacent features which were thought to depict the same glacier profile were connected up and downvalley. Lastly, a certain amount of a cross-valley symmetry was assumed so that corresponding glacier margins, where missing, were plotted at the same elevation on opposite sides of the valley. Obviously the map contains a certain amount of interpolation because evidence of glacier termini occupying valley bottoms is commonly absent due to postglacial sandur activity. Furthermore, no provision was made for floating ice tongues or ice shelves within the Inlet which, due to calving, may have had more abrupt termini than shown in Figure 4.15. Nevertheless, rapidly descending ice margins near valley floors could be projected to approximate terminal positions. Additional problems also arise when the ice thickness was greater than the height of the valley sides because in such a case no estimate of ice extent (thickness) can be made. Moreover, present glaciers in this area have a wide range of profiles, from very steep to nearly horizontal, depending on underlying topography, so that correlation along the valley may be inaccurate.

4.4.2 Interpretation

Despite the above constraints the Ice Retreat Map is thought to give a good impression of successive glacier margins as they occurred within Clements Markham Inlet. Retreat to contemporary margins appears to have taken place from expanded ice caps having a configuration similar to present. Although the spatial distribution of radiocarbon dates on the marine deposits provides chronological control for some of the margins, it is felt that a number of the northernmost ice margins, particularly those constructed entirely from ice marginal channels, may represent an older glacial cycle. These undated features may belong to one, or several, more extensive glacial cycles which preceded the last ice limit. Discussion of this aspect will follow in later chapters.

Gross chronological control provided by the location of dated marine shells, indicates that the majority of Clements Markham Inlet was free of a fiord glacier before 9845 $\pm$ 485 BP (25, Table 5.1), whereas, the head of the Inlet was occupied by glaciers until ca. 8 ka BP. The early interval, depicted in Figure 4.15, is characterized by a grounded glacier at the head of the Inlet while smaller side glaciers along the sides of the Inlet lie in contact with the high proglacial sea (full glacial sea). It will be proposed in the subsequent chapters that the ice margins which existed before 10 ka BP are the limits of the last glaciation (Plate 4.9). Subsequently, a number of sequential dates in the Moraine Creek area and the mouth of Windswept Valley (Figs. 4.7, 4.9) suggest slow retreat during the period ca. 10-9 ka BP. In fact, the glacier at the head of Breakthrough Delta was only 6 km from the contemporary terminus at 9.7 ka BP. This period of slow retreat probably continued until 8 ka BP, during which time large valley glaciers stood at the mouths of Piper Pass and Arrowhead Valley. Following 8 ka BP, rapid deglaciation resulted in the entire lower Clements Markham River Valley becoming ice-free within a few hundred years (by 7.6 ka BP).

The Ice Retreat Map also depicts numerous younger still-stands during recession upvalley. These were probably caused by minor glacioclimatic fluctuations that occurred while the valley glacier network disconnected. The fluctuations probably have related intervals of delta building which could be dated, but the existing stratigraphic record does not permit this correlation with specific ice margins. However, because significant deltaic progradation ceased along the NW coast of the Inlet in the period 6.5 to 7 ka BP (e.g. Omega Bay), the valleys in this area may have been entirely deglaciated by this time.

The scenario presented in Figure 4.15 is also confirmed by the stratigraphy of the Inlet. The NW shore is characterized by intercalated sequences of marine and glacigenic sediments suggesting that the side glaciers were calving or floating into a high relative sea level (10-8 ka BP). On the other hand, exposures in inner Clements Markham Inlet show a clayey till overlying bedrock, suggesting the presence of a grounded ice mass during this time. Furthermore, subsequent overstep conditions into marine sediments depict rapid marine inundation with an absence of intercalated sequences which would be produced by sedimentation from a nearby ice margin.

Further interpretation of the events described in this chapter is provided by reconstructing the sea level history of Clements Markham Inlet in Chapter V.

Chapter V

Sea Level History

5.1 Introduction

Chapter IV presented the geomorphic and stratigraphic data and, together with 15 radiocarbon dates, a general model of ice retreat was proposed. In this chapter further interpretations of these radiocarbon dates are made and the associated relative sea levels are discussed in terms of glacioisostatic adjustments. It will be shown that sea level changes and differential emergence through time provide additional information on the glacial history introduced in Chapter IV.

Table 5.1 lists the available radiocarbon dates from the study area and indicates their related sea levels. The specific location of each sample is shown in the figures within Chapter IV. Most dates are from marine shells whose stratigraphic positions were discussed in Chapter IV. However, additional dates were obtained on driftwood samples collected from fine marine sediments or from beaches where the related sea level could be determined.

5.2 Emergence Curves

The theory of postglacial emergence in the High Arctic has

Table 5.1 Radiocarbon Dates and Related Sea Levels

	Site Location	Lab. No.	Material	Age BP	Stratigraphic and Enclosing Material	Sample Elevation (m)	Related Sea Level (m)	Lat. N	Long. W
1.	Corner Delta	SI-4761	Shells	7635 ± 80	Colluviated Silts	82	102	82°35'	68°56'
2.	Piper Pass	GSC-2973	Driftwood	4660 ± 60	Beach	22.5	22.5	82°34'	68°43'
3.	Piper Pass	S-2083	Driftwood	5780 ± 100	Beach	22.5	>22.5	82°34'	68°43'
4.	Piper Pass	SI-4315	Driftwood	6445 ± 85	Surface	43	45	82°35'	68°37'
5.	Piper Pass	UQ-282	Driftwood	6730 ± 80	Water-cut bench	59	59	82°36'	68°25'
6.	Piper Pass	UQ-262	Shells	7860 ± 330	In situ, silts	67	103	82°36'	68°25'
7.	Piper Pass	GSC-2975	Driftwood	7830 ± 80	Surface	72	>82	82°37'	68°25'
8.	Owl Perch Delta	UQ-260	Shells	7540 ± 95	In situ, silts	60	93	82°40'	68°45'
9.	Arrowhead Delta	UQ-263	Shells	7900 ± 335	In situ, silts	75	103	82°38'	68°15'
10.	Arrowhead Delta	S-2210	Driftwood	8545 ± 110	Surface	93	>93	82°38'	68°00'
11.	Arrowhead Delta	S-2211	Driftwood	8915 ± 115	Surface	84	>84	82°38'	68°00'
12.	Arrowhead Delta	UQ-277	Shells	6050 ± 300	In situ, proximal bottomset f. sands	20	39	82°38'	68°12'
13.	Arrowhead Delta	SI-4314	Organics	6400 ± 60	Detrital, silts and f. sands	—	—	82°39'	68°05'
14.	Arrowhead Delta	GSC-3031	Driftwood	2180 ± 60	Beneath ice-pushed ridge	5.5	7	82°39'	67°46'
15.	Arrowhead Delta	L-251B	Driftwood	2190 ± 180	Surface	7	7	82°39'	67°46'
16.	Arrowhead Delta	SI-4316	Shells	7965 ± 65	In situ, silts	60	104	82°37'	67°50'
17.	Breakthrough Delta	S-1981	Shells	9745 ± 255	In situ, silts	100	117	82°38'	67°34'
18.	Breakthrough Delta	SI-4766	Driftwood	4340 ± 65	Surface	1.2	>1.2	82°35'	67°35'
19.	Gypsum River Delta	UQ-258	Shells	5720 ± 150	In situ, f. sand and silt	13	ca. 36	82°42'	68°08'
20.	Gypsum River Delta	S-2103	Shells	7965 ± 135	Surface fragments	104	104	82°42'	68°08'
21.	Gypsum River Delta	S-2124	Shells	8660 ± 155	In situ, sandy foresets	65	≥96	82°42'	68°07'
22.	Gypsum River Delta	S-2140	Wood	23 850 ± 850	Detrital, stratified sands	85	—	82°42'	68°07'
*23.	Gypsum River Delta	S-2137	Shells	10 690 ± 520	In situ, silts	65	124	82°42'	68°04'
24.	NE Gypsum River Delta	S-2135	Shells	6865 ± 130	In situ, silts	46	>46	82°42'	68°02'

25.	Moraine Creek	S-2123	Shells	9845 ± 485	<u>In situ</u>, silts	87	≥111-116	$82^{\circ}42'$	$67^{\circ}57'$
26.	Moraine Creek	S-2105	Shells	9270 ± 265	<u>In situ</u>, silts	87	111	$82^{\circ}43'$	$67^{\circ}57'$
27.	Moraine Creek	S-2115	Shells	8605 ± 140	Surface, silts	98	>98	$82^{\circ}42'$	$67^{\circ}57'$
28.	Eider Delta	S-2136	Shells	7640 ± 140	Surface, silts	78	>78-118	$82^{\circ}44'$	$68^{\circ}50'$
*29.	Eider Delta	S-2106	Shells	8535 ± 140	Surface fragments	81	>81-111	$82^{\circ}48'$	$67^{\circ}46'$
30.	Omega Bay	SI-4765	Shells	6860 ± 80	<u>In situ</u>, silty sand foresets	46	56	$82^{\circ}44'$	$67^{\circ}32'$
31.	Omega Bay	S-2127	Shells	7275 ± 135	Surface fragment	69	>69		
32.	Omega Bay	S-2125	Shells	5255 ± 110	<u>In situ</u>, coarse foresets	ca. 10	22-25	$82^{\circ}44'$	$67^{\circ}30'$
33.	Omega Bay	S-2126	Shells	6195 ± 120	<u>In situ</u>, coarse foresets	ca. 20	40	$82^{\circ}45'$	$67^{\circ}30'$
34.	W of Omega Bay	S-2117	Shells	7850 ± 130	<u>In situ</u>, silts	73	>73-110	$82^{\circ}45'$	$87^{\circ}27'$
35.	N Outermost Island	SI-4763	Driftwood	4215 ± 75	Surface, silts	21	21	$82^{\circ}46'$	$67^{\circ}10'$
36.	S Outermost Island	SI-4764	Driftwood	5515 ± 65	Behind beach ridge	8	8	$82^{\circ}45'$	$67^{\circ}56'$
37.	Mt. Beverley Delta	UQ-261	Shells	6480 ± 130	<u>In situ</u>, coarse foresets	40	55	$82^{\circ}42'$	$67^{\circ}05'$
38.	Mt. Beverley Delta	SI-4317	Shells	6595 ± 125	<u>In situ</u>, coarse topsets	31	>42	$82^{\circ}42'$	$67^{\circ}06'$
39.	Windswept Valley	S-2107	Shells	8605 ± 260	Whole halves in beach gravels	94	95	$82^{\circ}47'$	$67^{\circ}06'$
40.	Windswept Valley	S-2114	Shells	8935 ± 145	Surface fragments	82	>82-104	$82^{\circ}47'$	$67^{\circ}06'$
41.	Windswept Valley	S-2133	Shells	9195 ± 275	Surface, silts	85	>85-104	$82^{\circ}47'$	$67^{\circ}00'$
42.	Windswept Valley	S-2134	Shells	7250 ± 365	Surface	22	>22	$82^{\circ}47'$	$67^{\circ}00'$
43.	NE of Windswept Valley	S-2113	Shells	8345 ± 140	Surface fragments	85	>85-110	$82^{\circ}48'$	$66^{\circ}55'$
44.	S of Cape Colan	S-2104	Shells	8995 ± 150	<u>In situ</u>, silts	87	87-95	$82^{\circ}54'$	$66^{\circ}25'$
45.	Cape Colan	S-2116	Shells	855[illegible] ± 135	Surface, silts	67	81-90	$82^{\circ}56'$	$66^{\circ}20'$

*This sample redated 7940±130 (S-2483)

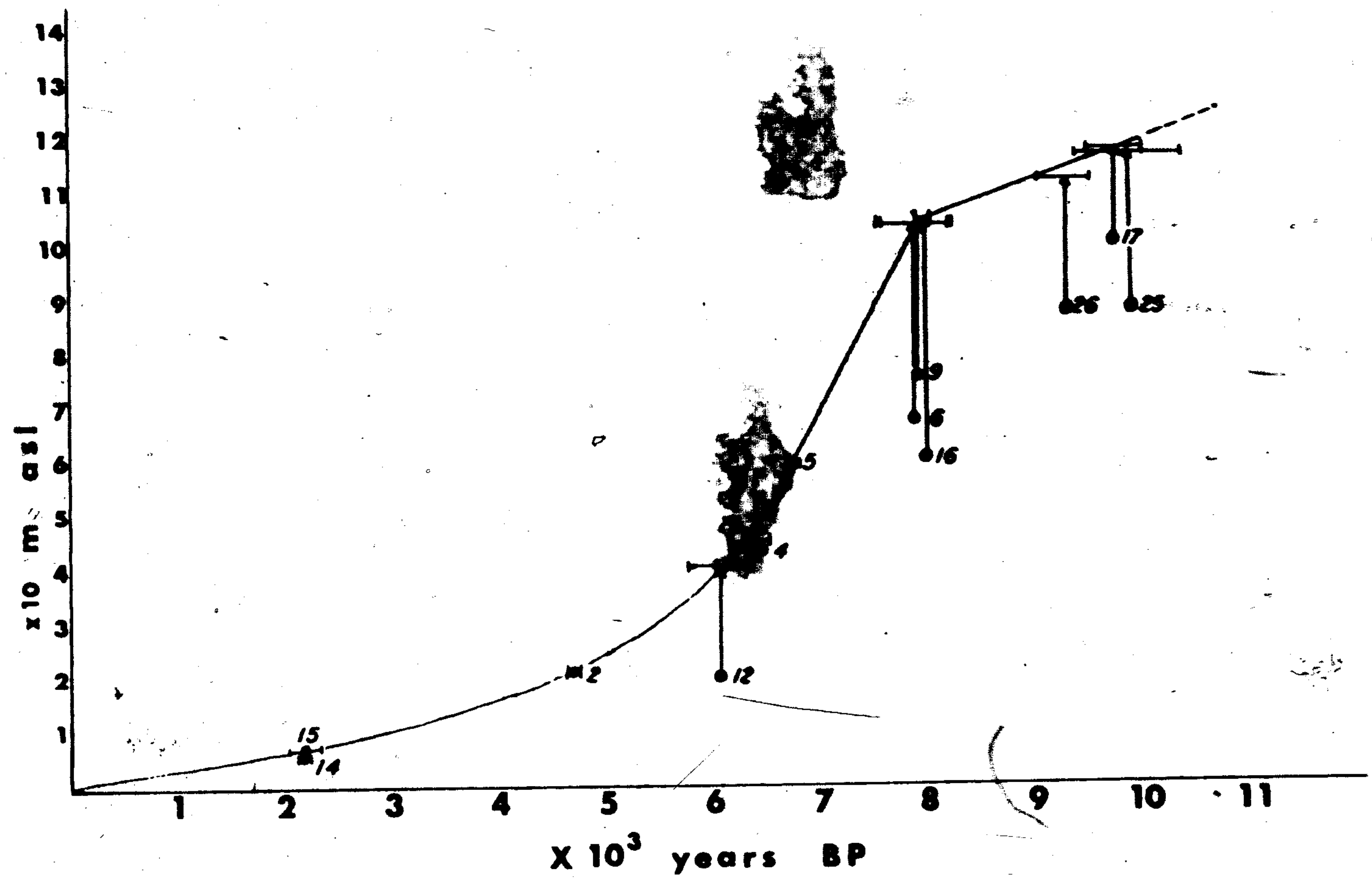
been discussed in Chapter III. Emergence curves portray the position of relative sea level for a specific site through time; but, because of differential emergence (Ch. II), they do not portray sea level changes over a wider area. Information provided in the subsequent section of this Chapter shows that, due to such differential emergence, there is a whole family of nested curves within Clements Markham Inlet.

Due to the abundance of low-elevation driftwood at the head of the Inlet, the most complete emergence curve was obtained from the Arrowhead Delta - Piper Pass area. Furthermore, concurrence of radiocarbon dates obtained on both marine shells and driftwood made it possible to cross-check the sea level points, resulting in a more reliable curve.

Figure 5.1 shows the pattern of emergence determined for the head of the Inlet. Note that the style follows a 'normal', smoothly decelerating curve from ca. 7.9 ka BP to the present (cf. Andrews, 1970; Blake, 1975). However, the earliest segment is dissimilar, showing only slow emergence from at least 11-10.5 ka to 7.9 ka BP (0.72 m 100 y^{-1}; cf. England, 1983).

Because the head of the Inlet was not deglaciated until ca. 7.9 ka BP, older dates on marine sediments can only come from outside the Arrowhead - Piper Pass area. Breakthrough Delta lies 7 km NE from the head of the Inlet and here the local marine limit is recorded by an ice-contact delta terrace at 117 m. Immediately downslope from

FIGURE 5.1 Emergence Curve for Inner Clements Markham Inlet.



this gravel terrace, in situ shells of H. arctica were collected from silt at 100 m (site 17, Fig. 4.5) and these provide a minimum age of 9745 ± 255 BP (S-1981) for the marine limit. Plotted on Figure 5.1, it is clear that the point 117 m at 9745 BP does not fall on the same uniformly decelerating curve described by the next lower points at 7.9 ka BP. Moreover, Breakthrough Delta is so close to these 7.9 ka BP sites that it is unlikely that this point could fall on a 'normal' curve simply shifted in time due to its earlier deglaciation. In the following section it will be shown that differential emergence in the Inlet indicates that the Breakthrough Delta date (9745 BP) is necessarily linked to very slow initial emergence prior to 7.9 ka BP.

The highest sea level observed in Clements Markham Inlet occurs in the Gypsum River Delta. This site is located on the opposite side of the Inlet from Breakthrough Delta, however it lies on approximately the same isobase (section 5.3, 5.4), and therefore, both sites can be plotted on the same emergence curve. In the lower Gypsum River a former sea level at 124 m is marked by an ice contact, marine limit terrace. Because the elevation of this marine limit is 7 m above the Breakthrough Delta marine limit, it must also be older than 9745 BP. A specific date on this marine limit was not found, but by projecting the slope of the emergence curve up to the 124 m marine limit an approximate age for it is obtained (10.5 to 11 ka BP). Therefore, the 124 m marine limit at the Gypsum River Delta suggests that the period of slow emergence defined at the Arrowhead Delta and the Breakthrough Delta began at least 10.5-11 ka ago. This period of slow emergence is also verified by two sea level estimates at Moraine

Creek (25, 26; Table 5.1). However, these points fall slightly below the emergence curve because this site is a short distance down-isobase from the Gypsum River Delta.

On the other hand, if one argues that the older dates are not related to these assigned sea levels, but are actually related to much higher, unrecognized sea levels which occurred in a preceding interval of 'normal' postglacial emergence, then these sea levels must approach 200 m (given ca. 3.8 m 100 yr^{-1}, the slope of the curve defined by the post 8 ka BP emergence). It should be noted here that Holocene marine limits at this elevation have not been reported in the High Arctic. In fact, the only comparable Holocene marine limits are found on the Canadian mainland associated with the demise of the Laurentide Ice Sheet (Dyke, 1979).

The initiation of rapid emergence occurred at ca. 7.9 ka BP when relative sea level stood at 103-104 m. This is based on three separate, minimum estimates on the age of the local marine limit in the Arrowhead Delta - Piper Pass area (7860 ± 300 BP, UQ-262; 7900 ± 335 BP, UQ-263; and 7965 ± 65 BP, SI-4316; sites 6, 9, 16). In each case, the shell samples were in situ *M. truncata* and *H. arctica*, found in rhythmically bedded bottomset silts representing the earliest migration of abundant fauna into the area following deglaciation (Fig. 4.15). Moreover, because these samples are from sites 2 to 3 km apart and elevations of 67, 75 and 60 m respectively, their similar ages suggest the initial entry of a fauna. Because bottomset sediments are commonly deposited contemporaneously with the marine limit topset sediments, it is concluded that these dates provide the best age estimate for the 103-104 m ice-contact terraces, and therefore the minimum age of deglaciation for this area.

The youngest segment of the curve is derived from driftwood logs found associated with prominent beaches. At the Arrowhead Delta dates of 2180 ± 60 BP (GSC-3031) and 2190 ± 50 BP (L251-B) were obtained on driftwood related to a 7 m beach (sites 14, 15; Stewart, 1981; Crary, 1960; respectively). A driftwood date of 4660 ± 60 BP (GSC-2973), associated with a 22.5 m beach (site 2); another driftwood date of 6445 ± 65 BP (SI-4315) associated with a 45 m strandline (site 4); and a date of 6730 ± 80 BP (UQ-282) associated with a 59 m strandline (site 5), were obtained a few kilometers closer to the Inlet head and these provide the upper driftwood points on the curve. Moreover, a shell date on a sample of <u>in situ Hiatella arctica</u>, (6050 ± 300 BP, UQ-277) collected from sandy foreset beds related to a former sea level at ca. 39 m, is in good agreement with the slope of the emergence curve indicated by the driftwood samples discussed above.

Several other driftwood dates obtained from this area gave anomalously old ages for their elevations (Table 5.1; sites 3, 7, 10, 11 and 18). The occurrences of younger driftwood or shells at elevations near, or above, these older samples clearly indicate that the anomalous dates were obtained on driftwood that had been redeposited downslope before becoming embedded in younger littoral deposits.

The pattern of postglacial emergence directly reflects the ice load history which is responsible for the crustal adjustments. Therefore, the emergence curve presented here indicates a slowly diminishing ice mass from before 10 to 8 ka BP, followed by rapid retreat causing unrestrained crustal rebound similar to other recently deglaciated, arctic areas (cf. Andrews, 1968, 1970). This relationship appears valid as geomorphic evidence presented in Chapter IV substantiates the minor retreat of small glaciers within the Inlet 10–9 ka BP followed by rapid recession of the main glaciers at the head of the Inlet ca. 8 ka BP.

The shape of the emergence curve has important implications. The early period of slow emergence reflects the initial thinning of ice following the culmination of the last glacial cycle. This segment of the curve could only have been recorded at or beyond the maximum extent of the last glaciation. Moreover, major stillstands are marked at the mouths of several valleys (Chapter IV), and although in a number of cases the maximum extent of glaciers could not be determined, other areas indicate glaciers calving into a high sea which likely represents their maximum positions with no evidence of more extensive positions.

A slow isostatic response to initial thinning of ice before the onset of rapid deglaciation has been suggested by Clark et al., (1978). Isostatic recovery occurring prior to deglaciation is termed restrained rebound. Although by strict definition restrained rebound occurs only at a site covered by thinning ice and therefore cannot be observed (Andrews, 1970b), restrained rebound will nonetheless be recorded in the peripheral depression of the ice sheet (Walcott, 1970). Furthermore, a period of high, near stable sea levels representing the apex of the glacial cycle should predate the time of slow emergence. This has recently been discussed by England (1983) who presents evidence for a stable sea level, at the marine limit, for several thousand years prior to initial emergence on NE Ellesmere Island (the full glacial sea).

It is generally thought that there was a worldwide sea level rise during early postglacial time due to water transfer as the Laurentide and Fennoscandian Ice sheets melted (<18 ka BP; Andrews, 1970). However, arguments presented in Chapter III imply that for relative sea level to be stable, isostatic uplift and eustatic sea level rise must be of equal magnitude. Although in situ thinning of ice sheets may cause restrained rebound, which could offset the eustatic rise in sea level during this time, recent hydro-isostatic modelling of global sea level changes suggests that the post-16 ka BP eustatic sea level rise in high latitudes was ≤ 0 (Clark et al., 1978). Moreover, Newman et al., (1979) could not find a coherent world wide pattern of postglacial eustatic sea level rise, based on current field data. Lastly, England (1983), did not find any evidence of ice recession in NE Ellesmere Island between 11-8 ka BP, although relative sea level remained essentially stable, thereby implying no local eustatic changes. It is therefore concluded that the eustatic component in Clements Markham Inlet was probably negligible and that emergence history closely approximates true crustal uplift, hence the ice load history during the early Holocene.

In summary, the emergence curve presented in Figure 5.1 represents a site at the margin of a major ice limit and therefore records the early proglacial, as well as the postglacial, sea level history. Slow emergence from a high sea (124 m a.s.l.) in the zone of the peripheral depression records the onset of deglaciation at least 11-10.5 ka BP, whereas rapid deglaciation did not occur until 8 ka BP. Geomorphic evidence (Ch. III) corroborates the ice load history recorded by the shape of the emergence curve. Flights of recently emerged beaches at the head of the Inlet indicate that emergence continues at about 30 cm per century (Plate 5.1); therefore isostatic equilibrium has not yet been achieved.

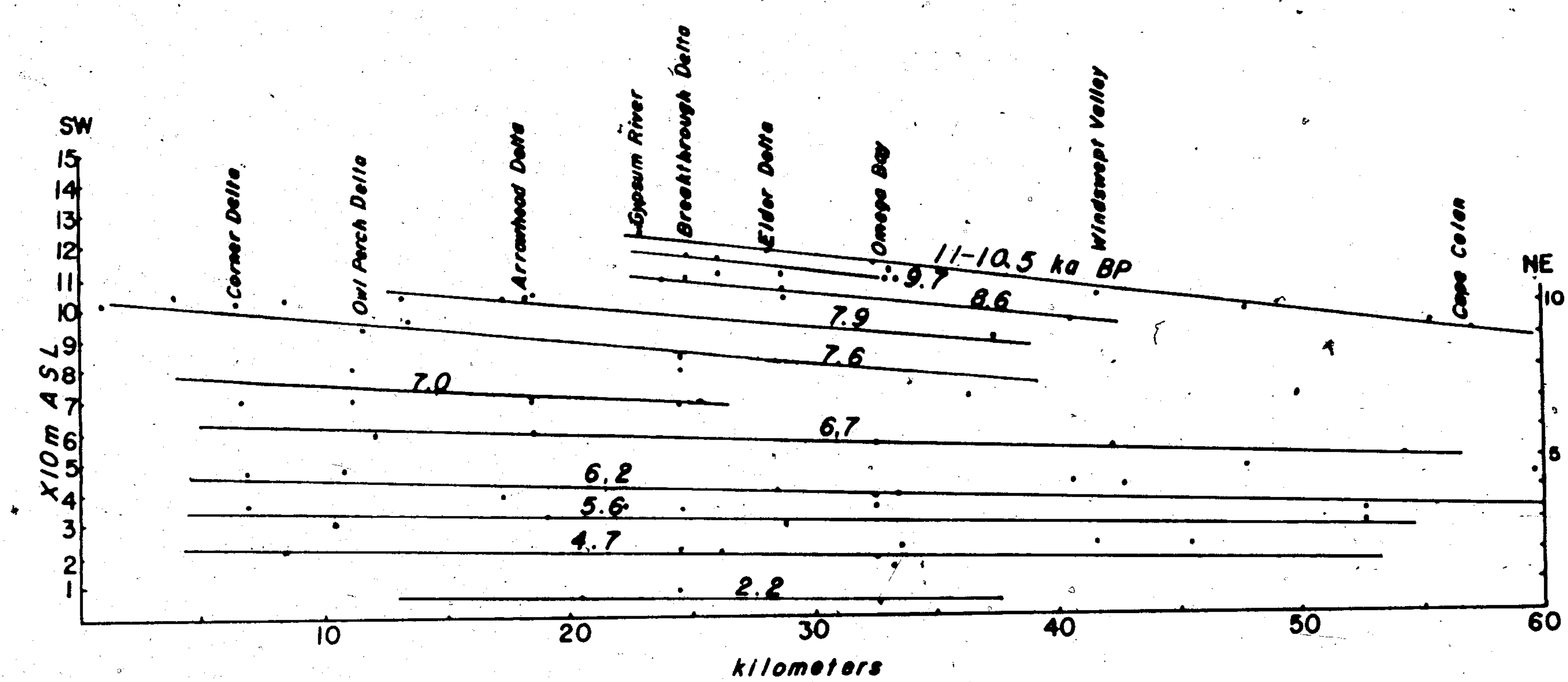
5.3 Equidistant Diagram

The elevations of sea level features along the sides of Clements Markham Inlet can be plotted against distance (Figure 5.2). The relative merits of such equidistant diagrams were discussed in Chapter III. Ideally, the distance axis on the diagram should project toward the center of maximum uplift so that it is orthogonal to the trend of isobases. Because the isobase pattern was unknown, the distance axis was rotated until the best agreement was reached between the shoreline profiles and dated sea level features. The shoreline remnants were correlated by radiocarbon dates, or, where dates were not available, the ages of intervening features were derived from the

Plate 5.1. Recently emerged beaches (<1 m a.s.l.) near the Arrowhead Delta suggesting that emergence is still ongoing at the head of the Inlet. The width of the flight of beaches shown is ca. 100 m.

FIGURE 5.2

Equidistant Diagram along Clements Markham Inlet



of many sites having local marine limits below the highest sea indicate the persistence of glaciers until later times (see Ch. IV). Conversely, initial emergence from the highest strandline must have occurred synchronously throughout the Inlet because it was initiated outside the last ice limit as soon as unloading began.

The equidistant diagram (Fig. 5.2) indicates a number of discrete strandlines. As discussed in an earlier review (Ch. III), reasons for shoreline and delta terrace formation are variable; for example, some features are erosional whereas others are depositional. Andrews (1970a) suggested that a correlation between delta formation and pulses of glacial activity was plausible. Although some of the upper strandlines presented here are clearly related to former ice margins, the relationship of lower water planes to the ice margins is not clear. Moreover, any number of water planes could be drawn to explain the scattered intermediate sea levels shown in Figure 5.2. Hence, the correlation of various Holocene strandlines found in Clements Markham Inlet to a Neoglacial record was not attempted here.

The tilting of the observed strandlines was caused by increased rates of emergence towards the former center of glacial loading. This differential emergence resulted in a family of nested emergence curves within Clements Markham Inlet. For example, emergence curves derived from sites farther out the Inlet (i.e. distal to the curve shown in Fig. 5.1) mimic its shape, but lie below it. However, because the rate of emergence decelerates thorugh time the curves approach each other later in the Holocene when the tilt of emergence curve at the head of the Inlet. Shoreline features of the same age were joined by straight line segments which depict the tilting of specific water planes. The best-fit projection of the distance axis has a bearing of 49°, nearly parallel to the Inlet and approximately orthogonal to the northern coastline.

The equidistant diagram shows a series of former water planes that dip down the Inlet. This dip decreases with successively lower and younger water planes. The uppermost strandline, which must date between 11-10.5 ka BP, has a maximum elevation of 124 m, at the Gypsum River, and it declines to 92 m a.s.l. at Cape Colan, over a distance of ca. 35 km. This uppermost strandline is considered to represent the proglacial sea which existed during the culmination of the last glaciation (Sec. 5.2). Subsequent, lower strandlines are subparallel until ca. 8 ka BP, after which there is a rapid decay in the rate of differential emergence.

The elevation of the marine limit increases from Cape Colan to the Gypsum River, whereas inland (SW) of the Gypsum River it drops in elevation. This is because inland of the Gypsum River the marine limit is younger due to the exclusion of the older, high sea by glaciers. Consequently, in this area (within the ice limit) a certain amount of initial emergence occurred before ice recession allowed the sea to contact the land. Similarly, further out the Inlet, a number of valley mouths also do not register the full glacial sea which suggests that glaciers occupied them during this early interval (ca. 11-10.5 ka BP). As discussed in previous sections, the stratigraphy successive water planes decreases. In fact, differential emergence of the lower water planes could not be resolved in the field.

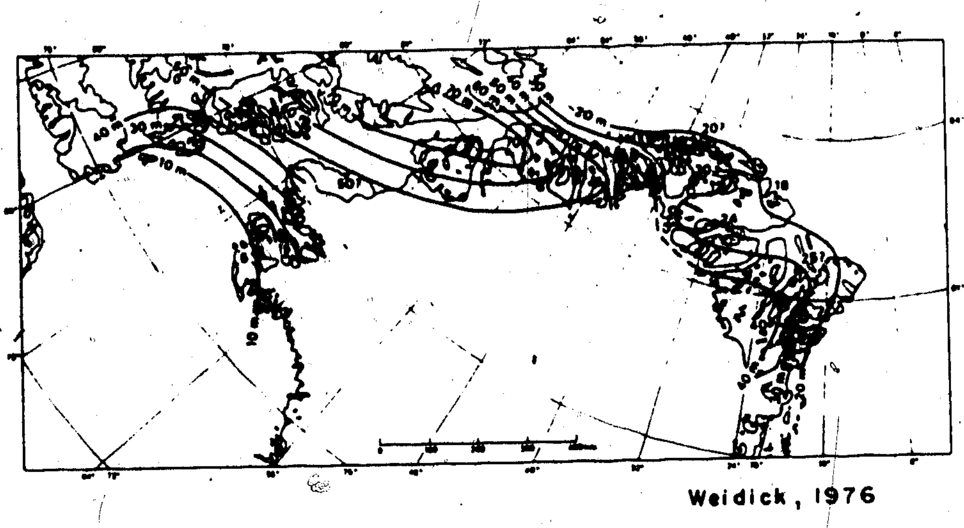
The strandlines are inclined toward the maximum former ice load which is toward present day ice caps in the heart of the Grant Land Mountains. Moreover, the upper strandline has a gradient of 0.9 m km^{-1} which is similar to values reported for early, postglacial shorelines in other Arctic areas (cf. Andrews, 1970b; England, 1983). The projection of the uppermost shoreline intersects present sea level ca. 137 km NE of Gypsum River. Given that the Gypsum River site lies near the last glacial limit, the width of the peripheral depression here is within the limits of the flexural parameter (crustal strength) calculated by Walcott (=110-140; 1970). Furthermore, given a ratio of mantle/ice density of ca. 3.7, it can be calculated that almost 500 m of ice occupied the head of the Inlet during the maximum of the last glaciation. Because of the complexity introduced by mountainous topography on ice sheet configurations and thickness (Reeh, 1982) further predictions from geophysical models based on simple monolithic ice sheet profiles are not attempted here.

5.4 Isobases

The equidistant diagram indicates the magnitude and direction of shoreline diplacement within Clements Markham Inlet. This two-dimensional information is now used to reevaluate the pattern of isobases on northern Ellesmere Island. The isobases describe the trend surfaces of shoreline displacements in the region for various times.

As noted in the review chapter, a number of published isobase maps exist for NE Ellesmere and NW Greenland (England, 1976, 1982; Wiedick, 1976; Fig. 5.3). In each case the isobases increase in value from the Lincoln Sea, toward northern Ellesmere Island and NW Greenland. Moreover, although England's and Wiedick's patterns are fundamentally different further inland on NW Greenland, both isobase patterns continue smoothly along the northernmost coast of Greenland onto Ellesmere Island, where the isobases trend obliquely across the northern tip of the island. As discussed previously, the variance between these two reconstructions is largely due to the scarcity of data. In fact, England's (1976) depiction has only two control points on the N coast of Ellesmere Island. With additional data, England (1982) showed essentially the same pattern, but with higher values on the N coast of the island.

The shoreline displacements presented in Figure 5.2 indicate an isobase pattern approximately orthogonal to Clements Markham Inlet. This suggests that the previously published isobases should swing more sharply northward as they cross Robeson Channel from Greenland. Furthermore, shorelines within the Inlet are higher than previously indicated for a given age, consequently the revised isobases must be displaced towards the Lincoln Sea. Figures 5.4, 5.5, and 5.6 show the isobases for 10, 8 and 6 ka BP respectively, based on the Clements Markham Inlet data plus additional information from NE Ellesmere Island (England, 1983) and preliminary results from Hall Land, NW



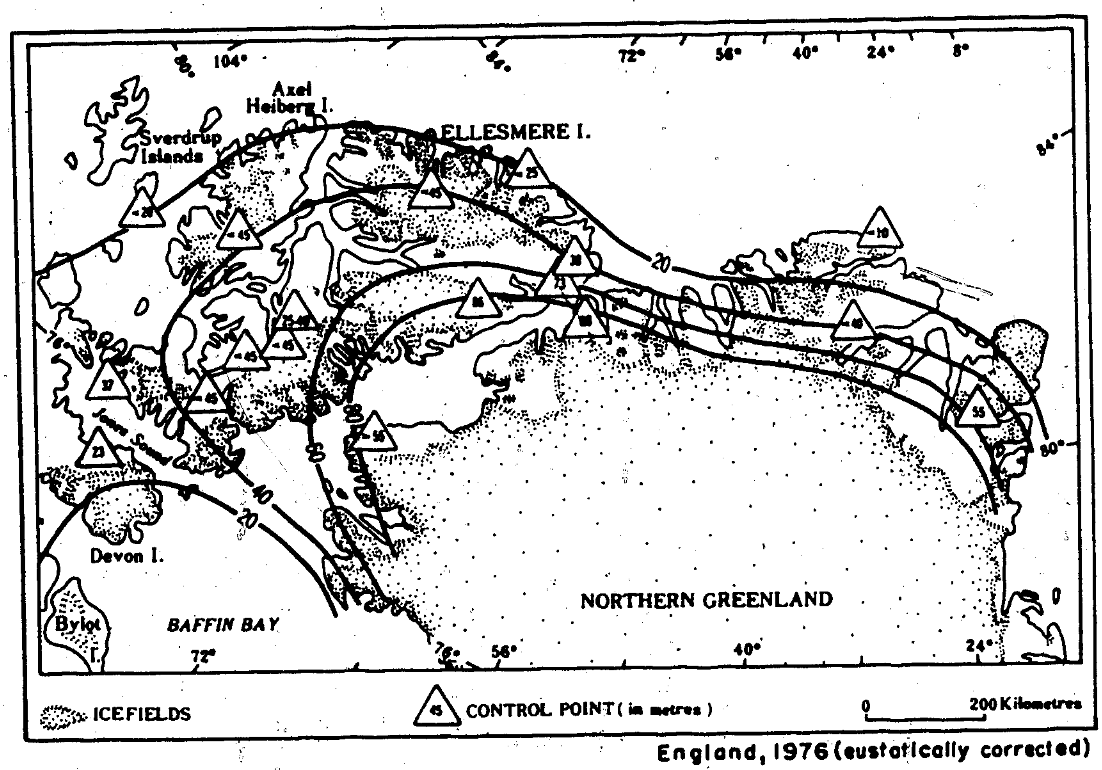


Figure 5.3 Regional isobases over the Queen Elizabeth Islands and adjacent northern Greenland coast for 6000 BP.

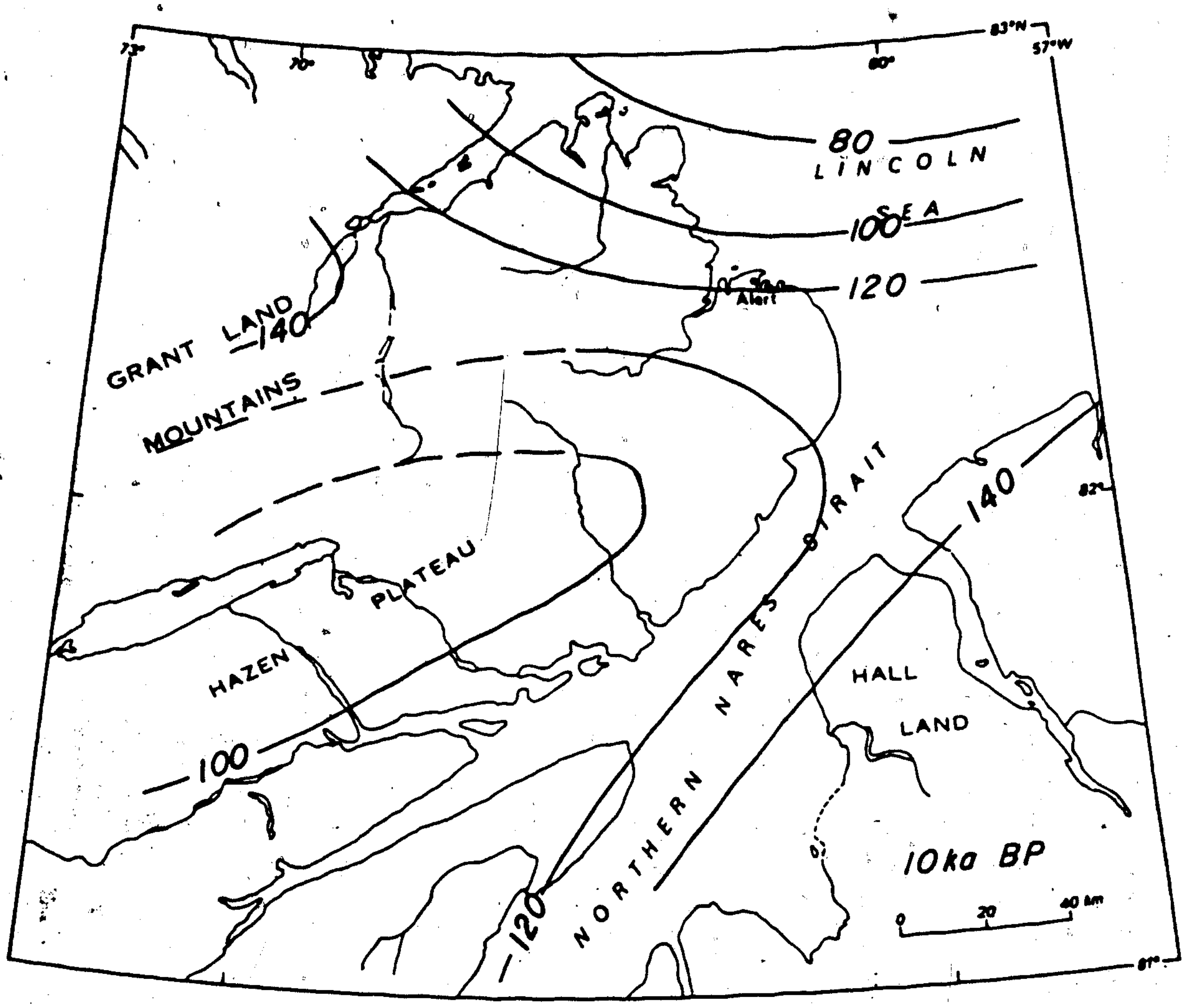


Figure 5.4 Isobases on the 10,000 BP shoreline over northern Ellesmere Island and northwestern Greenland (meters a.s.l.).

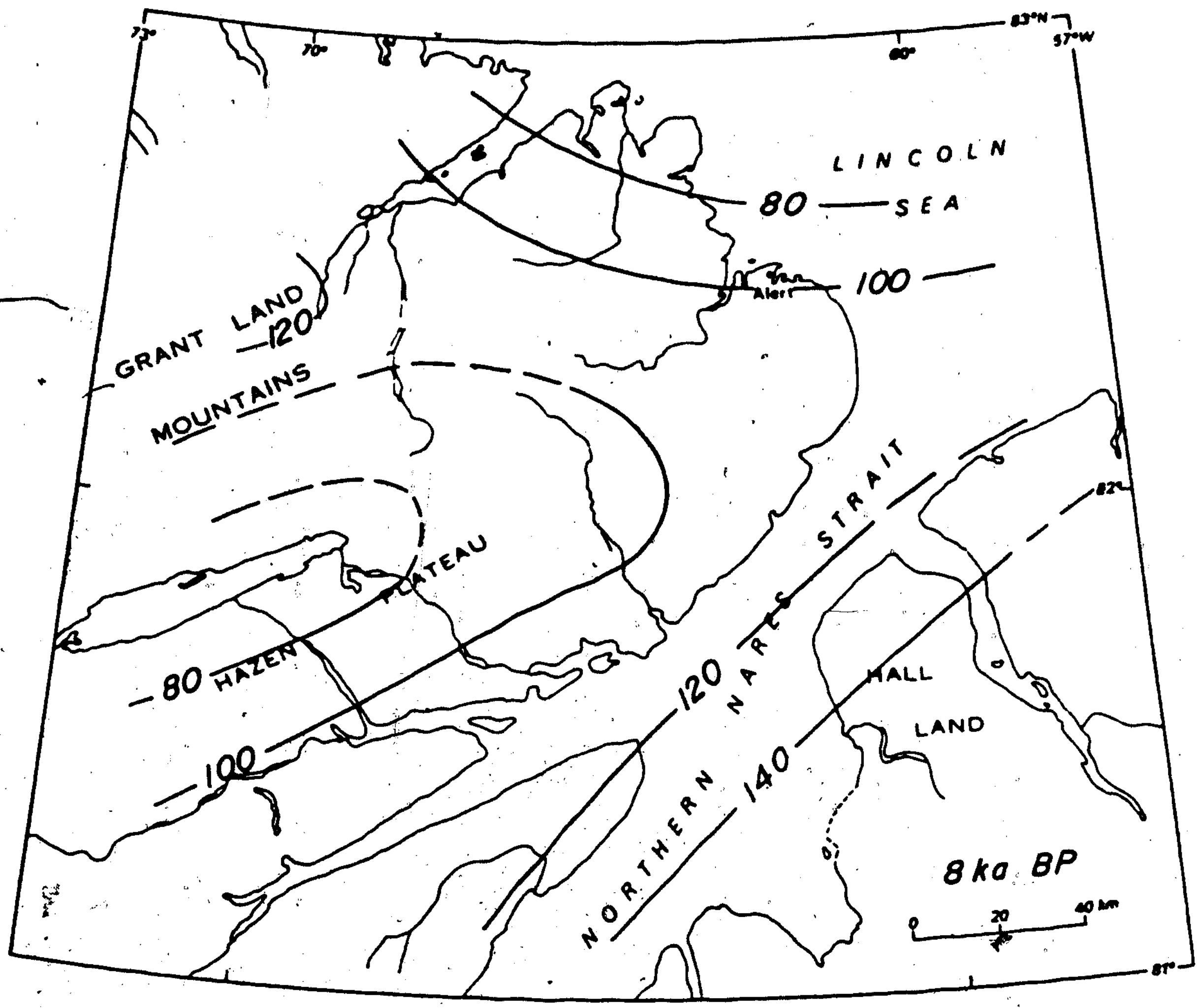


Figure 5.5 Isobases on the 8000 BP shoreline over northern Ellesmere Island and northwestern Greenland (meters a.s.l.).

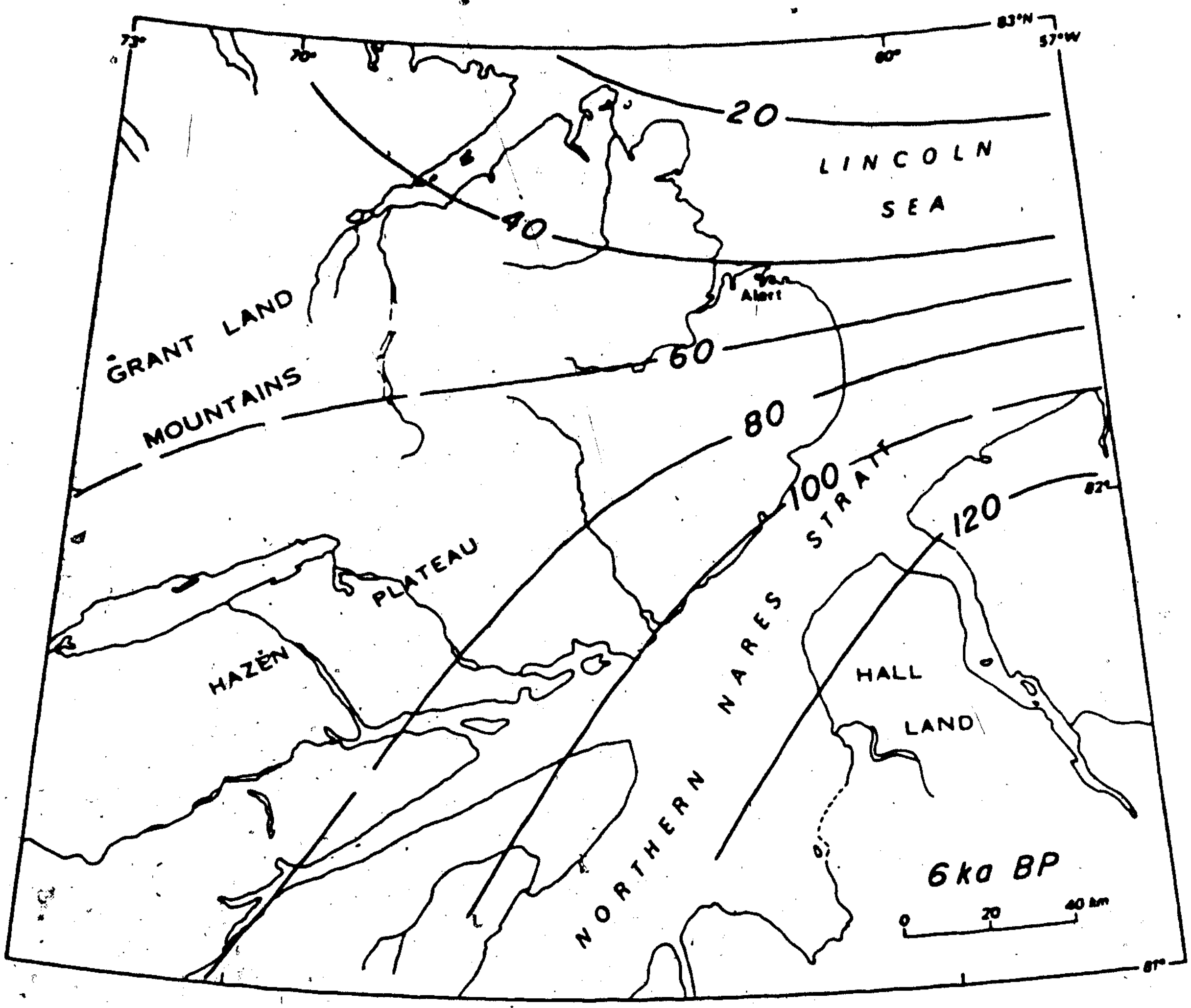


Figure 5.6 Isobases on the 6000 BP shoreline over northern Ellesmere Island and northwestern Greenland (meters á.s.l.).

Greenland (England, pers. comm., 1983). Although the isobase pattern at 6 ka BP is similar to those previously published (Fig. 5.3), alterations of England's 1976 and 1982 maps for earlier times are warranted (Figs. 5.4 and 5.5). The revised isobase maps (Figures 5.4 and 5.5) indicate that the recession of the ice caps in the Grant Land Mountains had a more significant effect in deflecting the isobases from NW Greenland than was previously envisaged. A separate center of uplift within the Grant Land Mountains produces a cell of low emergence over the Hazen Plateau. The cell diminishes with continued emergence of the Grant Land Mountains between 10 and 8 ka BP. Moreover, with post-8 ka BP accelerated emergence, the cell is gone by 6 ka BP (Fig. 5.6).

Much of the above scenario is governed by the different history of glacial unloading, hence emergence, between northernmost and NE Ellesmere Island (England, 1983). In NE Ellesmere Island, England (1983) describes a "full glacial sea" characterized by a high stable sea level that marks the marine limit from at least 11 to 8 ka BP. The full glacial sea predates any glacial unloading by at least 1 to 3 ka. This period was followed by an interval of slow initial emergence (0.7 m 100 yr^{-1}) from 8 to 6.2 ka BP, and a subsequent interval of 'normal', rapid emergence that extends to the present. As can be seen, a time-lag of ca. 2 ka occurs between the onset of initial slow emergence in this area compared to the Clements Markham Inlet area where it began at least by ca. 11-10.5 ka BP (Fig. 5.1). In fact, this earlier unloading history extends at least as far E as Alert, where England's (1983, Fig. 7) revised emergence curve is very similar to that for Clements Markham Inlet. However, despite the different unloading histories between N and NE Ellesmere Island, the isobases shown for 10 ka BP (Fig. 5.4) closely reflect the maximum depression, and hence, the total crustal loading caused by the last glaciation. Consequently the low cell in the isobases described above is not simply an artifact of different glacioclimatic responses between these two areas in late glacial and early postglacial time, but rather it reflects the relative distribution of the ice load.

Further clarification of the nonsynchronous nature of regional emergence is provided in Figures 5.7 and 5.8. These figures are derived from the isobase maps (Figs. 5.4, 5.5 and 5.6) and record isopleths of the amount of emergence that occurred between 10-8 ka, and 8-6 ka BP respectively. Figure 5.7 depicts the interval 10-8 ka BP during which uniform, slow emergence occurred over a large area in the NW (20 m), while the SE remained stable. Figure 5.8, on the other hand, indicates rapid emergence post 8 ka BP in the NW sector, while the formerly stable area in the SE begins to emerge at approximately the same rate as did the NW sector in the earlier interval (10-8 ka BP). Subsequently, the post 6 ka BP recovery along Nares Strait (Fig. 5.6) acquired similar high values as occurred in the NW in the post 8 ka BP period (ca. 3.5 m 100 yr^{-1}).

The isobase patterns describe a general SE migration of postglacial crustal response through the area during the early Holocene, with a 2 ka lag between the N coast and the northern part of Nares Strait. Emergence along the Ellesmere Island side of Nares

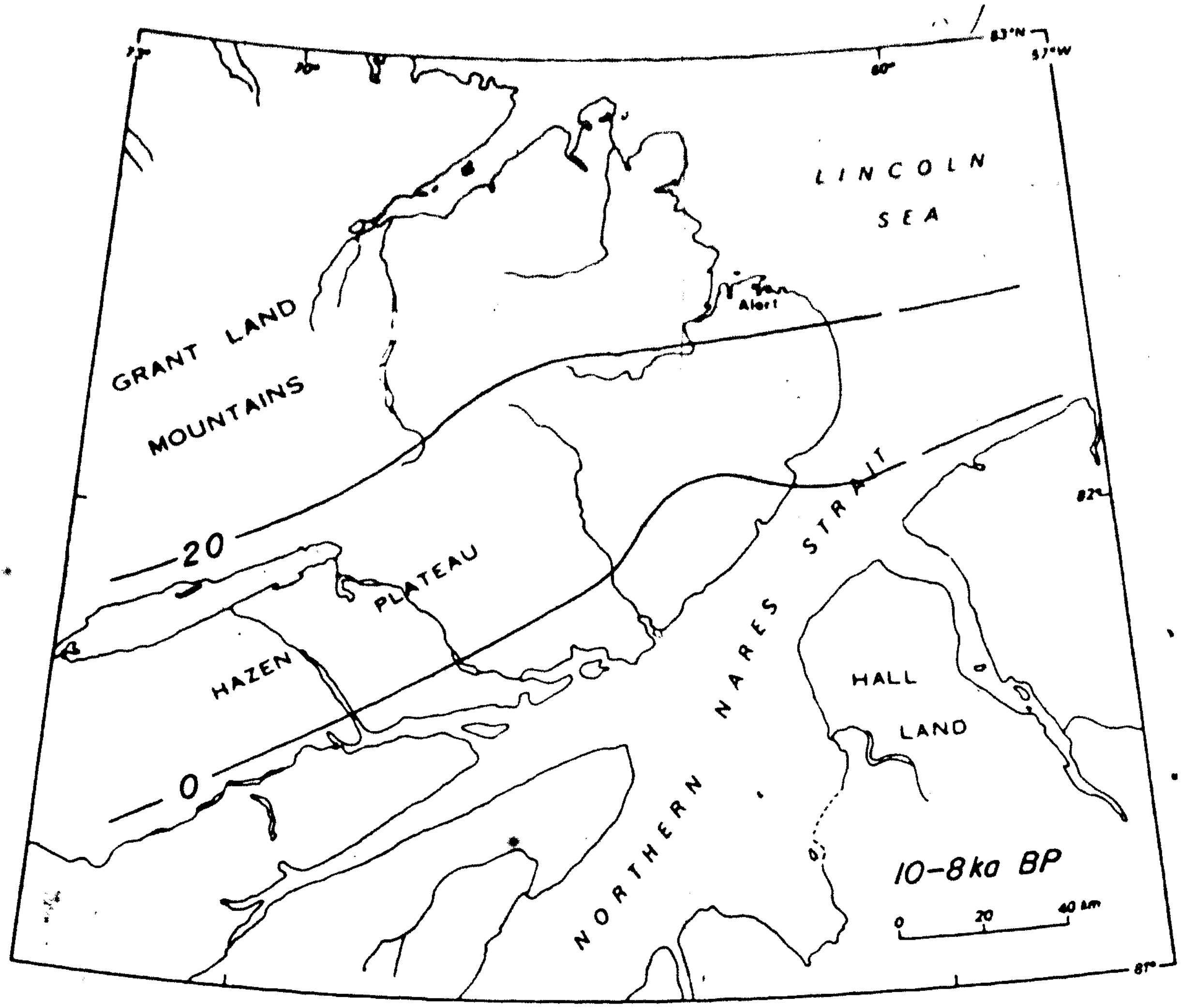


Figure 5.7 Emergence (meters) over northern Ellesmere Island and northwestern Greenland between 10,000 and 8000 BP.

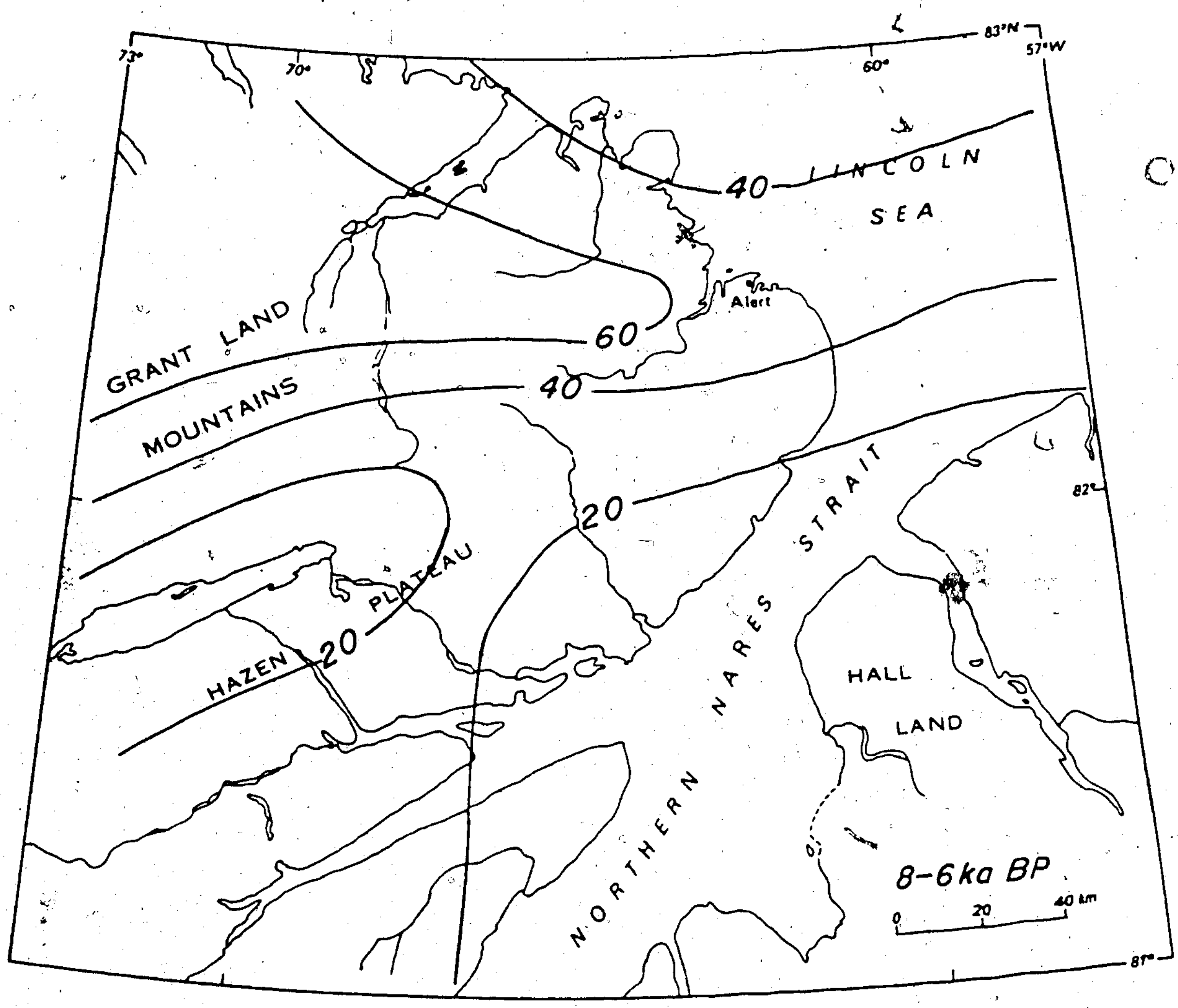


Figure 5.8 Emergence (meters) over northern Ellesmere Island and northwestern Greenland between 8000 and 6000 BP.

Strait is considered to be primarily governed by reduction in the NW Greenland ice load, (England, 1976, 1982, 1983). This may explain the lag in initial emergence in this area as opposed to the earlier and largely independent rebound further to the NW. Consequently, earlier retreat of the Ellesmere Island ice may explain the time-step in emergence between the N and NE (England, 1983; p. 911). Moreover, it has been suggested that the large isotopic shift at ca. 10.5 ka BP in the Greenland and Devon Island ice cores indicates an abrupt climatic amelioration (Dansgaard <u>et al.</u>, 1973; Koerner and Fisher, 1981). If this interpretation is correct, and the approximate response times for the Ellesmere Island ice caps and the Greenland Ice Sheet are ca. 2.5 ka and 4 ka respectively (Fisher, pers. comm. 1983), it is possible that the Ellesmere Island ice reacted first ca. 8 ka BP, followed by the Greenland ice ca. 6.2 ka BP. Because it is likely that some reduction in glacier mass would occur after the onset of the proposed warming at 10.5 ka BP, and before rapid retreat (ca. 8-6 ka BP), this may explain the slow glacial retreat that was already under way in Clements Markham Inlet by ca. 9.8 ka BP. On the other hand, England (1982, 1983) found that glaciers on NE Ellesmere Island began to slowly retreat ca. 8 ka BP, a time when deglaciation in Clements Markham was well under way. To reconcile earlier unloading in the N, England (1983) suggested that a significant difference in glacioclimatic regimes may occur between the N and S sides of the Grant Land Mountains. It appears that the dated ice margins bear this out.

5.5 <u>Summary</u>

Shoreline displacements in Clements Markham Inlet indicate the sea level history for the last 11,000 years. The emergence curve derived for the head of the Inlet is characterized by two distinct intervals: 1) slow emergence from 11-10.5 ka to 7.9 ka BP at a rate of 0.72 m 100 yr^{-1}; and 2) rapid emergence after 7.9 ka BP which decelerated towards present sea level. It is noteworthy that the slow emergence in the first period is similar, in rate, to the initial rate reported on NE Ellesmere Island, which occurred ca. 2 ka later (England, 1983; 0.7 m 100 yr^{-1}). The two intervals in Clements Markham Inlet are considered to represent a period of slow unloading by ice, followed by rapid deglaciation beginning ca. 8 ka BP, and this is also confirmed by the stratigraphy and geomorphology described in Chapter IV. Because the complete emergence curve is recorded near the head of the Inlet, and because the marine limit extends from the outermost coast to this site, this synchronous marine limit (11-10.5 ka BP) records a full glacial sea which occupied areas beyond the maximum extent of the last glaciation. This is also partly substantiated by the stratigraphy near the Gypsum River, and at more distal sites, where local glaciers debouched into the full glacial sea. Evidence for an interval of earlier emergence, indicating recession from a more extensive ice cover has not been found. The equidistant diagram shows a series of water planes tilting toward the former center of maximum ice load within the Grant Land Mountains, and defines the trend of the isobases which run approximately parallel to the northernmost coast. The uppermost water plane, defining the regional marine limit, marks the tilt of the full glacial sea. Conversely, local marine limits at lower elevations within certain valley mouths define the retreat of more extensive valley glaciers which contacted the full glacial sea.

On a regional scale, shoreline displacements along the northern coast of Ellesmere Island are apparently beyond the isostatic influence of the NW Greenland Ice Sheet. New isobase patterns based on this study define a center of uplift in the northern Grant Land Mountains which in turn lead to a low cell in the Lake Hazen area, to the south. Southeast from the cell, shorelines rise toward the NW Greenland ice load (England, 1976, 1983). The local Grant Land Mountain rebound decays rapidly after ca. 8 ka BP, while the depression on its SE side is maintained until ca. 6.2 ka BP. This difference in emergence between the N and SE sides of the Grant Land Mountains is probably due to dissimilar glacioclimatic regimes, as well as to the differential response times affecting the Greenland and Ellesmere Island ice masses (England, 1983; this study).

Chapter VI

Discussion

6.1 Introduction

The objective of this study was to expand the glacioisostatic data base for northern Ellesmere Island and to relate it to the late glacial stratigraphy. It was assumed that this area would have a discrete isostatic response independent of the Greenland ice load, and that the Arctic Ocean would have a greater glacioclimatic influence here than on NE Ellesmere Island. Furthermore, the information concerning the Quaternary history of Clements Markham Inlet is relevant to the question of ice extent in high latitudes during the last glaciation (late Wisconsin time) and this provides constraints for interpreting long-term ice core records in the region. This chapter summarizes the major findings from Clements Markham Inlet in order to further discuss the conflicting hypotheses of minimum versus maximum ice coverage during the last glaciation.

6.2 Paleogeography and Chronology

In Chapter III a facies model for glaciomarine sedimentation was presented which serves as an interpretive tool for distinct stratigraphic sequences within the study area. This model stressed that the combination of sediment input, ice proximity, and rate of emergence governs the facies transitions within a stratigraphic column. The resulting stratigraphic interpretations, together with various radiocarbon dates in Chapter IV, were complemented by detailed mapping which led to a reconstruction of former ice margins in the area. Moreover, numerous radiocarbon dates provided chronological control for correlating sections and ice margins throughout Clements Markham Inlet. Lastly, reconstructing the sea level history in Chapter V further refined the interpretations in Chapter IV by showing the relationship between ice margins and sea levels. Furthermore, the differential emergence of raised shorelines defined isobase patterns which provide a regional perspective. The general unloading of northern Ellesmere Island was independent in relation to the emergence on NE Ellesmere Island, which reflects the influence of the bordering Greenland Ice Sheet.

In order to summarize the above information, the paleogeography of Clements Markham Inlet is shown as it may have occurred ca. 11 ka, 8 ka and 6 ka BP (Figs. 6.1, 6.2 and 6.3; respectively). These ice margins, in relation to the former coastlines, identify the zone of interaction between marine and glacial environments during phases of deglaciation. Again, it is possible that ice shelves formed within the Inlet during these stages, or that certain glaciers extended further into the sea than indicated.

At 11 ka BP (Fig. 6.1) the head of Clements Markham Inlet was indundated by confluent glaciers emanating from major valley systems along the northern flanks of the Grant Land Mountains. Major outlet

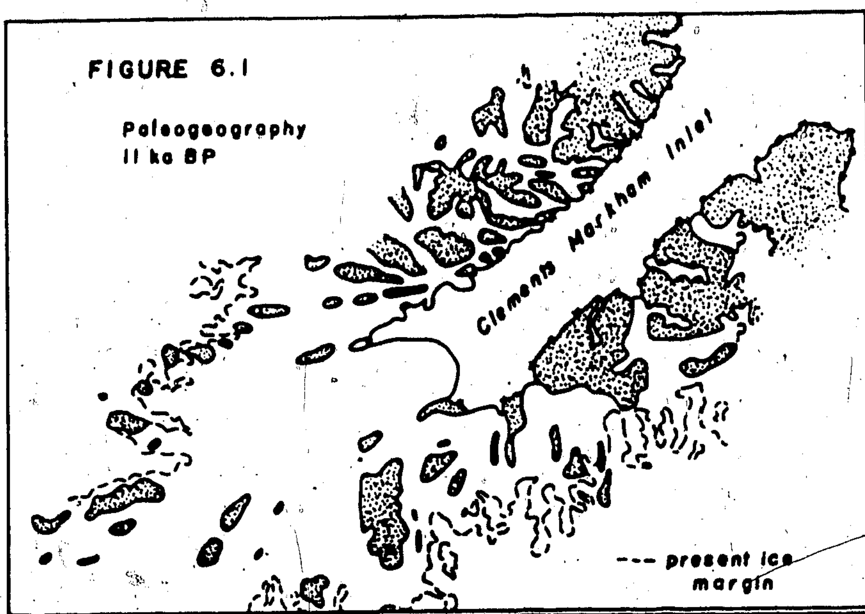


FIGURE 6.1

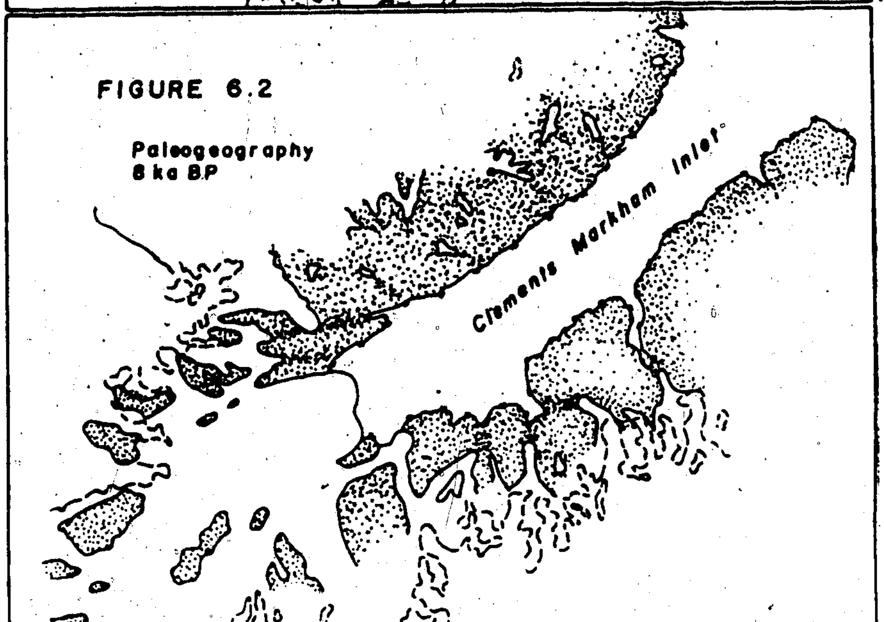


FIGURE 6.2

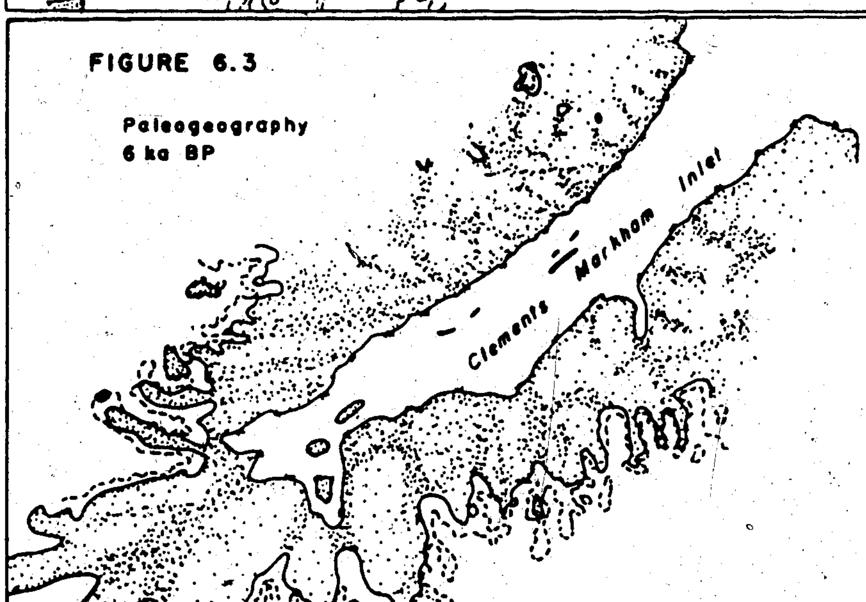


FIGURE 6.3

glaciers advanced out of Piper Pass, Clements Markham River valley, and Crescent Glacier valley (Fig. 2.1), extending 15 to 30 km from present margins. The overall ice cover at 11 ka BP diminishes toward the outer coast because the glaciers were smaller and confined to separate valleys from which they calved into the sea along the steep sides of the Inlet (Fig. 6.1). Widespread stratigraphic sections near the mouth of the Gypsum River document these floating ice margins. The relative sea level at ca. 11-10.5 ka BP was 124 m along the ice limit near the head of the Inlet and it graded down to 92 m at the mouth of the Inlet at Cape Colan (Figs. 5.2, 6.1).

The period between 11 ka and 8 ka BP was marked by slow glacial retreat and slow initial emergence. Nonetheless, by 9.7 ka BP some glacier termini along the SE coast were ca. 6 km from the present ice caps. Although at 8 ka BP (Fig. 6.2) the inner lowlands at the head of Clements Markham Inlet were still inundated by ice, major retreat of the smaller glaciers occurred in the outermost region of the Inlet. During this time the relative sea level dropped to 104 m near the head of the Inlet (Fig. 6.2).

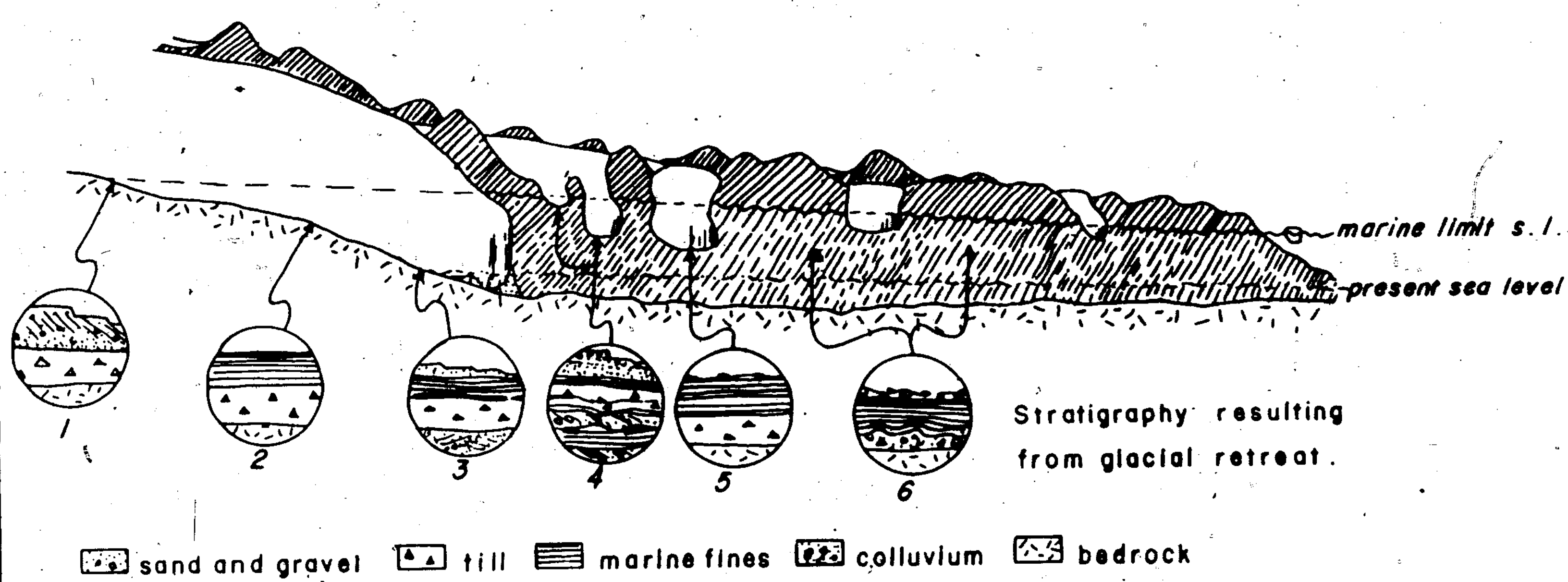
The period after 8 ka BP was marked by rapid deglaciation coupled by falling sea levels that resulted in the head of the Inlet becoming ice-free in less than 400 years. As the valley bottoms became ice-free marine transgressions occurred to maximum distances of 10 km inland from the present coastline. Deglaciation was such that most glaciers probably reached their current positions or even disappeared by ca. 6-5 ka BP when a large decrease in delta building is recorded. Certainly, radiocarbon dates of 6-5 ka BP obtained on organic material adjacent to present glaciers in other parts of Ellesmere Island, indicate that these ice margins are as extensive now as they have been in the last 6000 years (e.g. England, 1978; Lowdon and Blake, 1979; King, 1981; Völk, 1981).

The situation in Clements Markham Inlet 10.5-11 ka BP is also shown in the cross-sectional profile (Fig. 6.4). This diagram shows the general stratigraphic sequences found throughout the Inlet and it includes facies changes during subsequent ice retreat and emergence. For example, the central lowlands at the Inlet head were occupied by thick ice which was able to mold the bedrock and deposit till (Fig. 6.4, Section 2). Rapid deglaciation of this area, ca. 8 ka BP, was coupled with a marine inundation resulting in an overstep condition characterized by fine bottomset silts directly overlying the till or glacially scoured bedrock. Moreover, because these central areas are removed from the immediate sediment sources at the mouths of valleys, rapid postglacial emergence exposed these silt plains to extensive subaerial erosion. However, areas of concentrated sediment influx (i.e. valley mouths) produced sufficient coarse debris to cap the fines, producing a flight of successively lower delta terraces as sea level regressed (Fig. 6.3, Section 1). As the rate of emergence slowed in later times, widespread deposition of littoral deposits could occur (Fig. 6.4, Section 3).

Beyond the terminus of thick ice occupying the head of the Inlet (Fig. 6.1), there was a peripheral depression occupied by a full

FIGURE 6.4

Diagramatic sketch of glaciation in Clements Markham Inlet ca. 11 ka BP.



glacial sea where shorelines formed during the period of slow emergence 11 ka to 8 ka BP. The stratigraphy in the full glacial sea is very diverse and depends on the proximity to, and thickness of, the former glaciers flowing out of the side valleys (Fig. 6.4, Sections 4, 5, 6). Generally, the stratigraphy is characterized by intercalated sequences of marine and glacigenic deposits, indicating fluctuating ice margins that floated or calved into the full glacial sea. The marine limit in the full glacial sea formed a continuous strandline along the sides of the Inlet which was only broken in certain areas where glaciers from side valleys debouched into the sea (Fig. 6.4). The upper strandline abruptly ends at the ice limit near the Inlet head, and sites which have local marine limits below the uppermost sea indicate glacier occupation. This is clearly exemplified by lower and younger marine limits at the head of the Inlet, inside the ice limit.

From the evidence presented in this study it is clear where the major 11-10 ka BP ice limits were located. In Chapter IV it was noted that the ice retreat map (Fig. 4.15) showed undated ice marginal channels beyond the proposed ice limit. However, it was suggested that they could date from an older, more extensive glacial cycle (pre last glaciation). Furthermore, volcanic erratics on mountain summits near the outer coast imply a major ice flow parallel to Clements Markham Inlet, which surmounted mountain barriers. Certainly, this implies complete inundation by ice during some unknown interval in the past. Nonetheless, a redeposited willow branch within delta sands from the watershed SE of Mount Rawlinson, near the source of the erratics, provided a radiocarbon date indicating that ice-free areas existed ca. 23,850 ± 850 BP (S-2140, assuming this is a finite date). Trying to establish the age of the more extensive ice cover by defining zones of relative weathering is problematical because of the fast weathering of the sedimentary rocks and varied microclimates. Generally, the northernmost, ice-free areas of Clements Markham Inlet appear to have more frost shattered, rubble covered surfaces than the highlands adjacent to the ice caps, but these differences could not be systematically quantified. Consequently, the question of the glacial history prior to ca. 11 ka BP can only be deduced from the current data and this will be attempted here.

There are two basic scenarios for Clements Markham Inlet. The first is that the maximum extent of the last glaciation is marked by the ca. 11 ka BP margin which was limited to the Inlet head and the mouths of some valleys (Fig. 6.1). This hypothesis has already been suggested in the discussion of sea level history. The alternative hypothesis is that the ca. 11 ka BP margins are merely a recessional stage from a more extensive glaciation. This major advance may, or may not, be associated with the glacial stage during which an ice sheet completely inundated the area and dispersed the mountain summit erratics. The alternative hypothesis, then, would imply that all the margins shown on the ice retreat map are due to a single glacial cycle during which a grounded fiord glacier, or ice stream, extended out into the Lincoln Sea (cf. Christie, 1967). Based on available data this scenario has several important consequences. First, this glacial advance must have occurred after 23.8 ka BP, as vegetated surfaces presumably existed in more proximal areas during this time. Moreover, these sites must have been overrun as the ice advanced to the more distal sites. Secondly, because the marine limit rises toward the head of the Inlet, the fiord glacier must have retreated rapidly (cf. Andrews, 1975). If so, this rapid deglaciation started before 9.8 ka BP, however, it must have slowed or ceased as the ice terminus reached the Inlet head ca. 8 ka BP, causing subsequent proximal marine limits to be lower. Third, sequential retreat of a fiord glacier would also imply that ^{14}C dates on the marine limits should be progressively younger as the head of the Inlet is reached (cf. Blake, 1972; England, 1983). Moreover, postglacial emergence of sites along the sides of the Inlet should be characterized by a series of nested, "normal" emergence curves.

Clearly, the consequences of this alternative hypothesis do not agree with most of the data. First, a systematic decrease in the age of the marine limit was not found as one goes up the Inlet. In fact, the oldest sample was found near the head of the Inlet. Therefore, it would have to be assumed that the age distribution of the 45 radiocarbon dates is random, and that an age progression simply was not found. Furthermore, if the marine limit at Cape Colan was considerably older than a Gypsum River as would be expected if the ice limit reached the mouth of the Inlet or beyond, the ca. 10 ka BP shorelines would lie well below 92 m at Cape Colan and older shorelines would have much steeper tilts. Secondly, the determined trend in marine limit elevations does not agree with the established emergence curve if the alternative hypothesis of extensive glaciation prior to 9.8 ka BP was true. This is because the alternative hypothesis necessitates that the marine limit be younger as it rises towards the head of the Inlet. In turn, this rise in marine limit would mean that the presumed fiord glacier retreated rapidly (cf. Andrews, 1975), hence, crustal unloading should also be rapid ca. 9.8 ka BP. However, the emergence curve indicates slow unloading during this time (10 to 8 ka BP) which is obviously in contradiction with the rapid retreat. Clearly, the only way the emergence curve can agree with the trend in marine limit elevations within the Inlet is if the marine limit was formed synchronously, beyond the last ice limit. Furthermore, the validity of the emergence curve was discussed in Chapter V, where it was noted that even if only the younger part of the curve is approximately correct (< 8 ka BP), then the sea levels related to the older shell dates (9.8 ka BP) would have to be in excess of any elevations yet observed on Holocene shorelines in the entire Queen Elizabeth Islands (if 'normal' rapid emergence began at 9.8 ka BP instead of 7.9 ka BP).

In conclusion, although it may always be argued that the absence of data does not necessarily preclude one or another hypothesis, the first hypothesis, that the ca. 11 ka BP margin marks the last ice limit, accommodates all the current data. Furthermore, the detailed nature of the field work undertaken makes it unlikely that the dated samples misrepresent the overall data set. Lastly, England (1983) also recognized at least two discrete glacial cycles related to the expansion of Ellesmere Island ice caps. Although, as noted in Chapter V, the timing of deglaciation during the last cycle appears ca. 2 ka out of phase with northernmost Ellesmere Island, the overall model from NE Ellesmere Island is quite similar; also involving a full glacial sea beyond the last ice limit.

The favoured hypothesis, however, does have two shortcomings which are addressed here. The first of these deals with the paucity of dates related to the full glacial sea. Emergence from the full glacial sea should be synchronous, therefore the marine limit, outside the influence of local valley glaciers, should date the same everywhere along the sides of the Inlet (cf. England, 1983). Although there are several dates in the 10-9 ka BP range evenly distributed along the Inlet, these must relate to slightly lower sea levels. Furthermore, the full glacial sea must necessarily have been in existence throughout the last glacial cycle, but shells older than 9.8 ka BP have not been found. Nonetheless, Clements Markham Inlet contains vast areas of barren silts which may predate 10 ka BP and the stratigraphy in some areas shows that marine deposits underlie glacigenic units which record the initial transgression during the onset of the last glaciation.

The second shortcoming of the hypothesis is that, if the major ice reached its limit near the Inlet head, why are there no higher, older sea levels related to the earlier, more extensive glacial cycle? This is an important point, but a possible explanation exists if the NW and SE shores of the Inlet are compared. For example, although high sea levels from the last glaciation are well documented on the NW side, there is no evidence of similar sea levels on the SE side. As noted earlier, this is probably due to a combination of steep slopes, rapid mass movement, and little sediment input which resulted in poor preservation of rudimentary sea level indicators on the SE side. This example, then, could be a pertinent analog to explain the absence of the older sea level indicators. Given enough time, even the raised shorelines on the NW side of the Inlet will likely be removed by prolonged weathering and erosion.

6.3 Conclusion

The previous section summarized the data obtained from Clements Markham Inlet and reviewed arguments pertaining to the location of the ice limit. It was concluded that the ice retreat map shows the most plausible sequence of ice configurations during deglaciation, whereas, the most extensive margins probably relate to a more ancient glacial cycle. Moreover, corroboration between the stratigraphy, sea level history, and age distribution of radiocarbon dates is best explained if the most recent glaciation did not completely inundate the Inlet but rather was restricted to about a 40 km advance from contemporary margins in the major valleys. Just as the extent of today's ice cover diminishes from the ice caps of the Grant Land Mountains to the ice-free areas near the mouth of the Inlet, so too did the relative intensity of the last glaciation. Confluent glaciers flowed from expanded ice caps invading the head of the Inlet, while cirques in the outer margins were re-occupied by smaller glaciers (Fig. 6.1).

Finally, this study has answered questions pertaining to the late Quaternary history of northernmost Ellesmere Island that were previously unaddressed. Nevertheless, as with most studies of this nature many new questions are raised. Some of these questions are summarized here and the direction of future work is indicated.

(i) The primary need is to find marine deposits beyond the last ice limit that span the last glacial cycle. Older shells may have been overlooked in the field, particularly at lower elevations, because no simple relationship exists between the appearance of such shells and their age. Perhaps bottom sampling of the Inlet would produce the older shells. Moreover, bathymetric studies may also identify terminal moraines which would define the limit of the major ice within the center of the Inlet.

(ii) Secondly, because the thesis has proposed glacial limits abutting a full glacial sea, one would expect to find evidence of higher, older sea levels related to previous glacial cycles. As noted these were not observed in Clements Markham Inlet, but they may exist in the numerous bays and fiords along the northernmost coast of Ellesmere Island.

(iii) Furthermore, each additional dated sea level in the area would contribute to extending the isobases. This study has shown that the addition of a single emergence curve or cluster of points from a small area can substantially alter the isobase maps. Therefore: (1) There is an urgent need for studies of this sort, and (2) debates about the extent of regional ice cover and ice mass geometry which rely on isobase maps for sole support should cease until the data base is more sound. For example, the emergence data identified an independent center of glacioisostatic response over the Grant Land Mountains, hence a similar study is recommended on the SE side of these mountains, along Nansen Sound, in order to fix the position of the uplift center. This information would be valuable in assessing the size and extent of the Ellesmere Island ice load relative to the Greenland ice and the larger Laurentide Ice Sheet to the south. Of particular interest is that this study has shown that a limited advance of the glaciers in the Grant Land Mountains (<40 km) produced up to 124 m of emergence near the ice limit. This is similar to the amount of emergence found in west-central Ellesmere Island where Blake (1970) proposed the center of the Innuitian Ice Sheet to be. More work on NW Greenland may also clarify the nature of the emergence saddle in the Alert/Lincoln Sea area. As data of this nature become available on northern Ellesmere Island, controls will be provided for the applicaton of theoretical ice sheet and glacioisostatic models (i.e. Reeh, 1982; Quinlan and Beaumont, 1981, respectively).

(iv) This study has shown that the onset of deglaciation within Clements Markham Inlet began ca. 2 ka sooner than south of the Grant Land Mountains. This implies a dissimilar glacioclimatic regime which deserves further study, particularly by the disciplines of glaciology and synoptic climatology, in order to identify contemporary controls on mass balance. Once these present controls are identified, the need for some presently unknown factors, such as changes in the sea ice cover of the Arctic Ocean, may be assessed.

(v) Lastly, during the course of this field work, problems arose concerning the genesis of various ice-contact, glaciofluvial features which may be associated with floating or calving ice margins. Moreover, a number of these features may also be indistinguishable from those left by ice shelves (sea ice and glacial types). Therefore it is recommended that more work on the contemporary geomorphic processes of present ice shelves be undertaken (cf. Sugden and Clapperton, 1981). Isotopic variations within marine pelecypods, reflecting relative meltwater production, may also shed light on the types of modern and ancient proglacial environments (^{18}O, ^{13}C; cf. Hillaire-Marcel, C., 1977; Stewart, unpublished).

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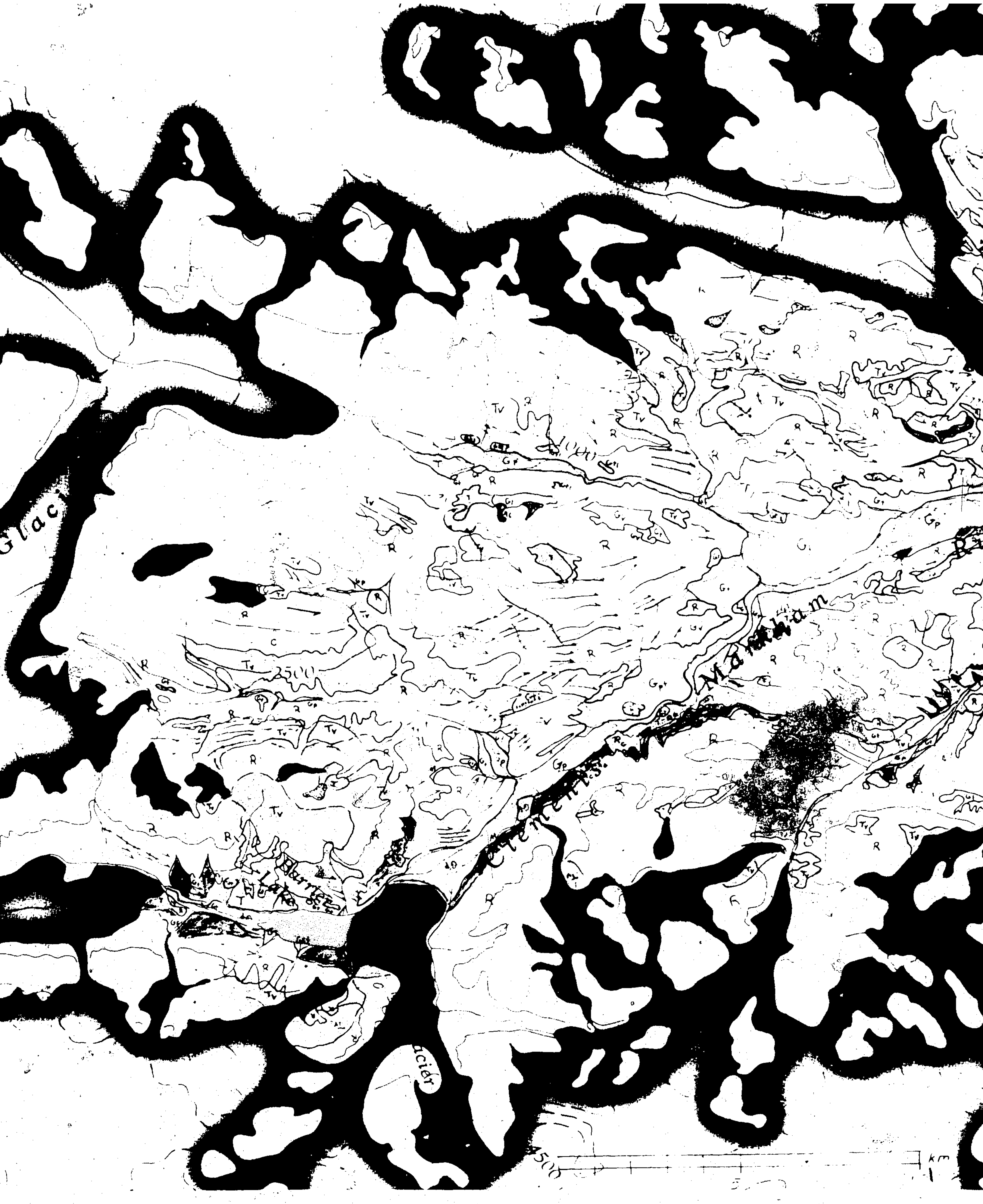

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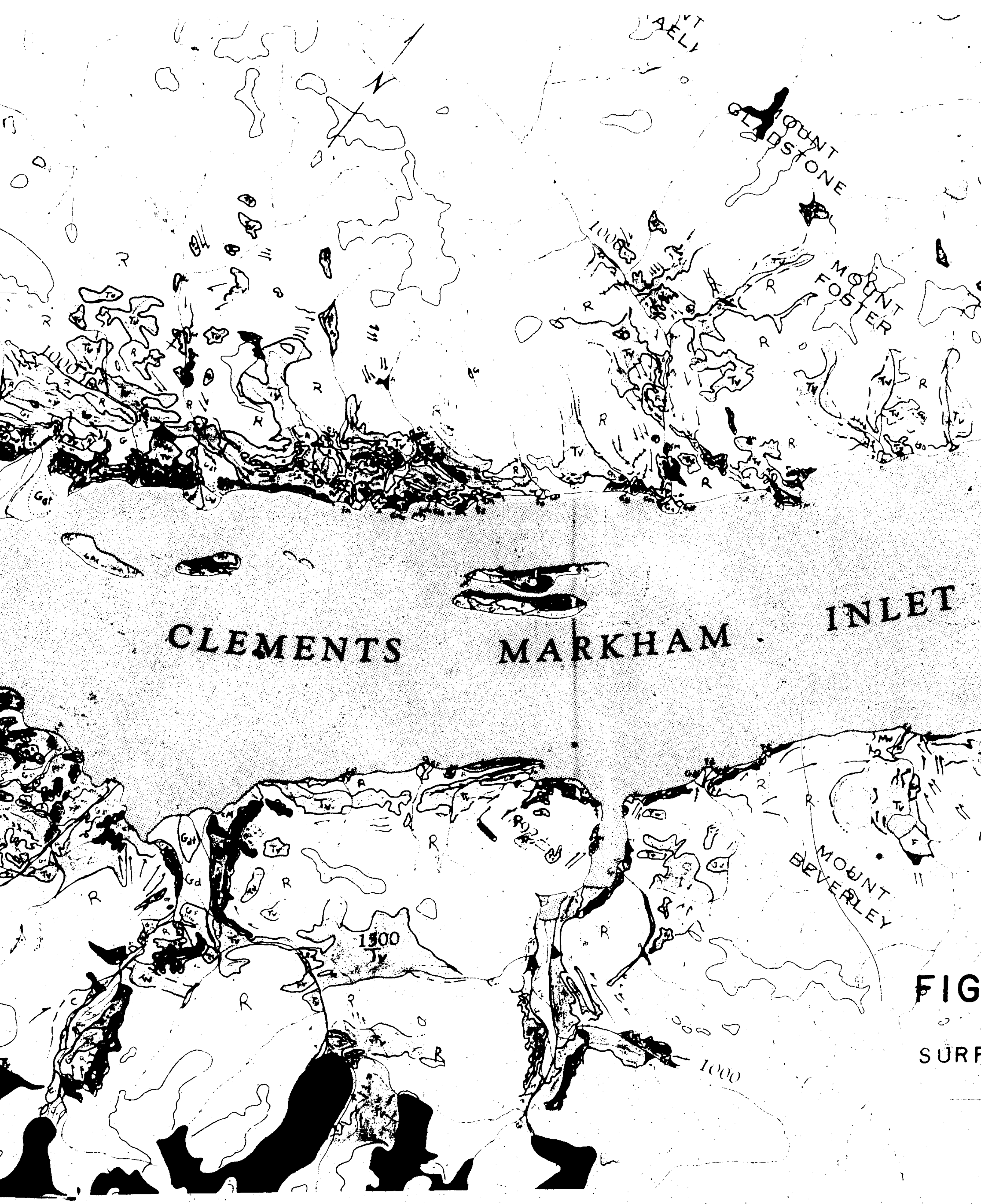
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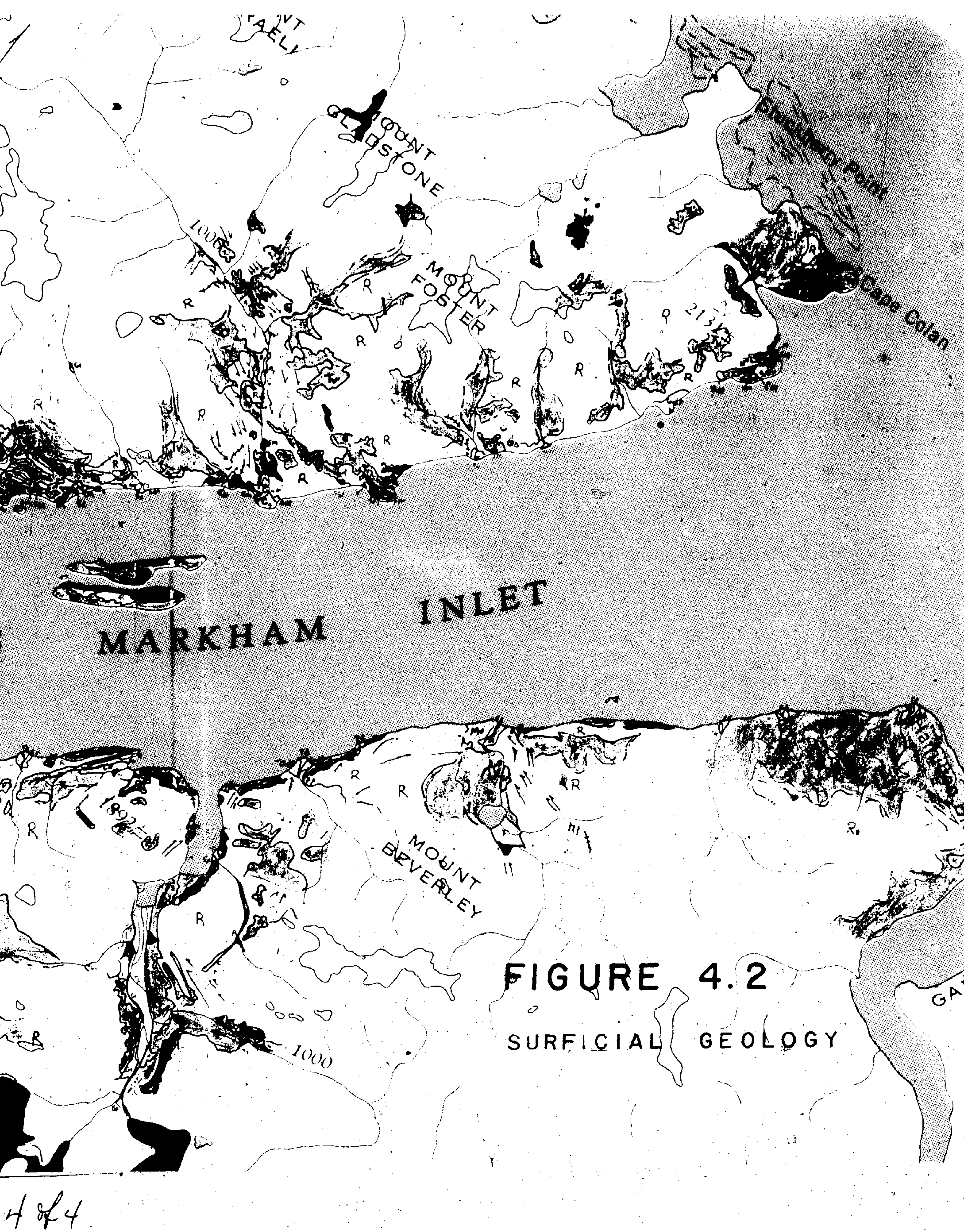


FIGURE 4.2

SURFICIAL GEOLOGY

TABLE 4.1

Surficial Materials of Clements Markham Inlet

Legend for Figure 4.2

Unit

C - Colluvium; unsorted rock debris mantling valley slopes and floors, solifluctted or washed from upslope weathered rock.

RG - Rock Glacier; mobilized unsorted detritus containing an ice core or interstitial ice, may be derived from lateral moraines or colluvium.

A - Alluvium; detrital material which has been transported by water having no apparent glacial source, commonly forms fans along steep valley sides.

F - Fluvial sediments; Gravel and sand deposited on valley floors by rivers having no apparent glacial source, rare in Clements Markham Inlet.

M - Marine sediments (cM - coarse; fM - fine); Glaciomarine sediments deposited in seawater, generally horizontally stratified fines comprising the bottomset component of a delta sequence.

GM - Glaciomarine sediments; coarse unsorted debris, commonly overlying marine silt, probably ice-rafted or ice shelf deposits.

LG - Glaciolacustrine sediments; proglacial lake deposits.

G - Glaciofluvial sediments; gravel and sand deposited beneath and in front of the marginal zone of a glacier.

T - Till; unsorted debris deposited directly by a glacier.

R - Rock; various lithologies and ages, but mainly sedimentary rock which is often very frost shattered.

Descriptor

v - veneer; thin discontinuous, generally < 0.5 m thick

b - blanket

t - terraced; inactive, elevated above present level of activity

f - fan

d - delta; generally the downvalley end of a sandur being deposited in water

i - ice-contact; perched or kettled deposit, e.g. kame terrace (Gti)

p - plain; e.g. outwash plain (Gp)

r - ridged; e.g. beach ridges (Mr)

Symbols

Cirque

Drumlin

Striae

Moraine

Ice-contact face

Kettle hole

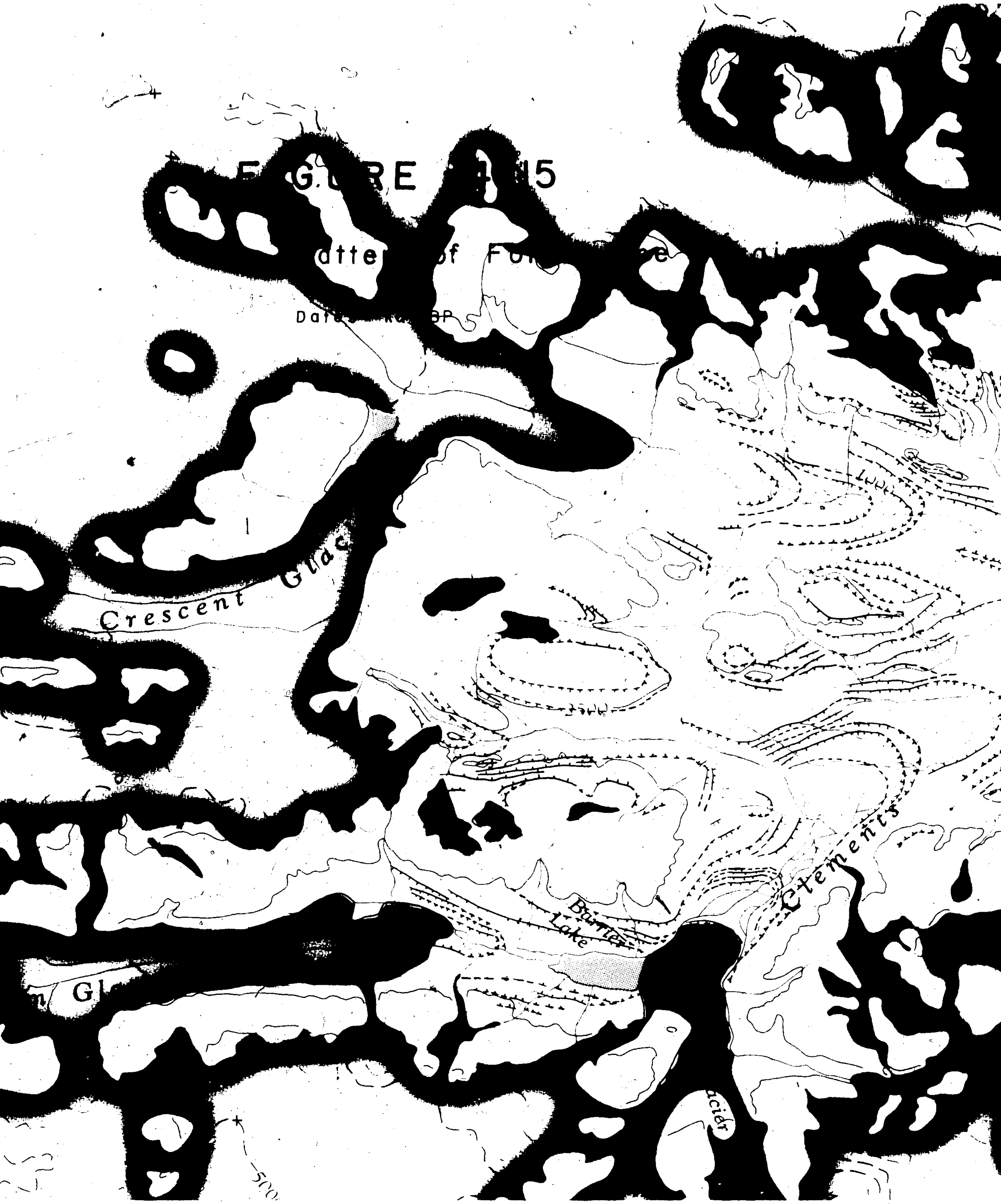
Abandoned channel (large, small, side hill)

Intermediate shoreline features

Escarpment

Canyon

Wind fluted silts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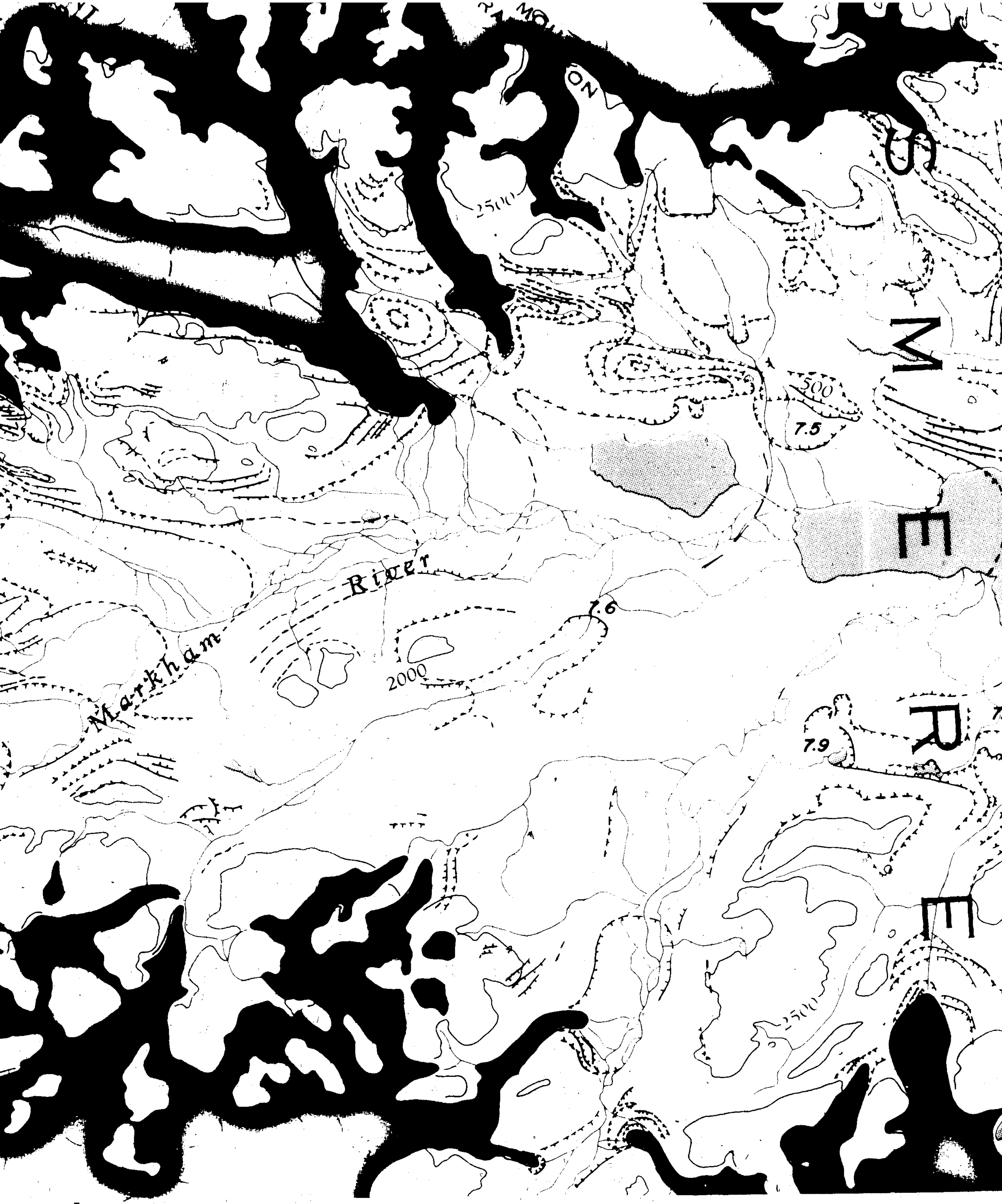
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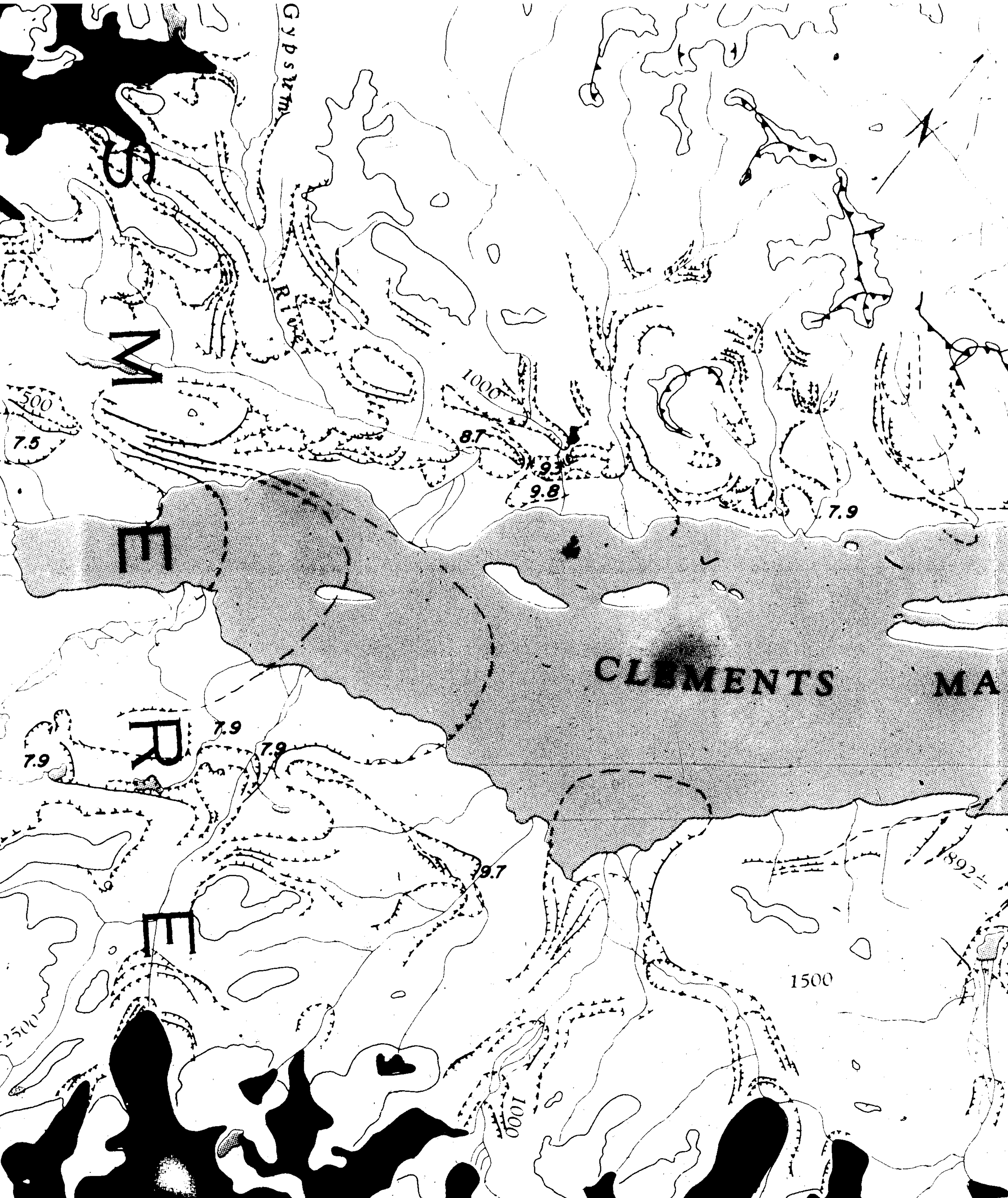

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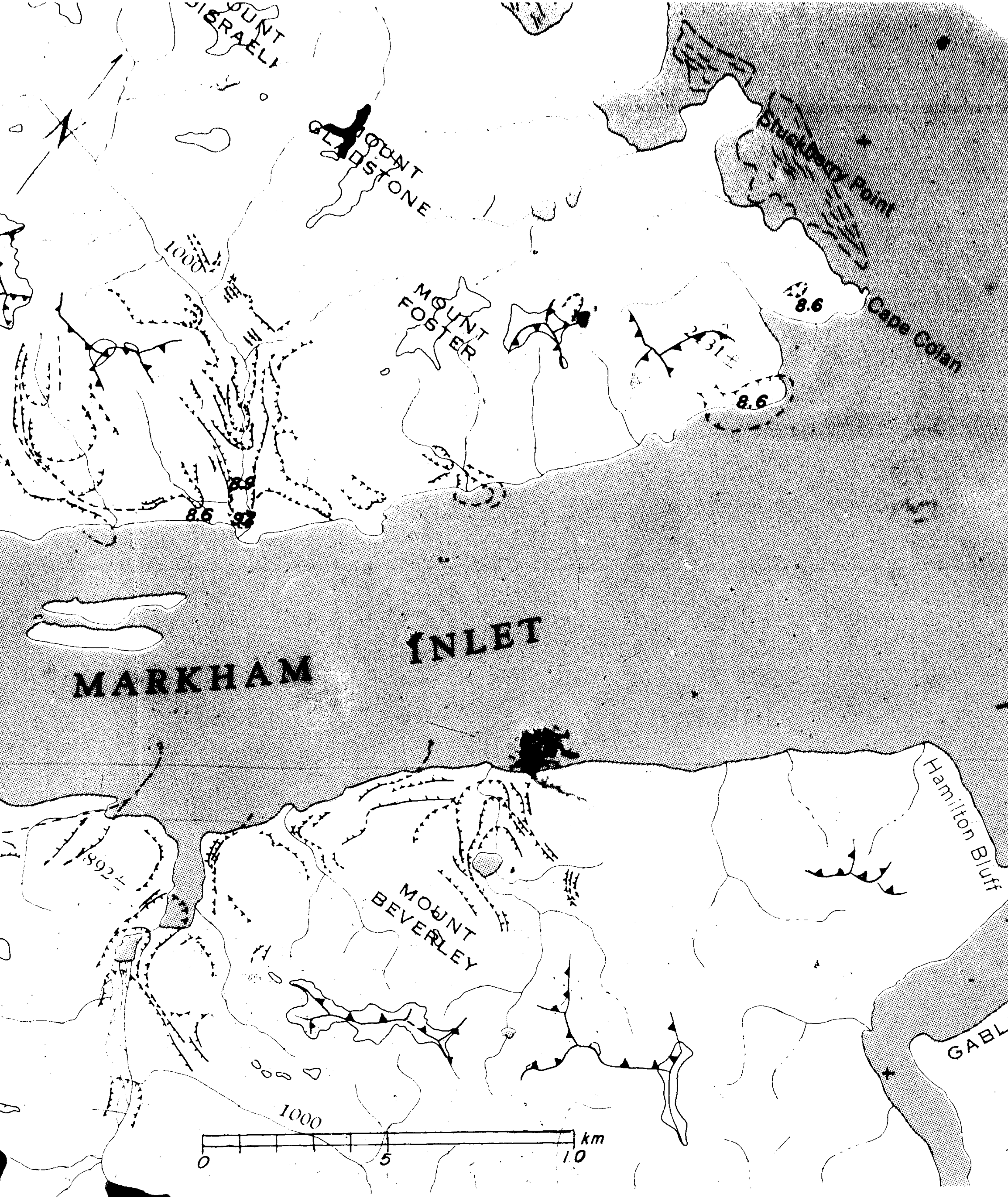
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