TIMING AND EXTENT OF GLACIATION IN THE PANGNIRTUNG FJORD REGION, BAFFIN ISLAND: DETERMINED USING *IN SITU* PRODUCED COSMOGENIC ^{10}Be AND ^{26}Al

A Thesis Presented

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ABSTRACT

Over 140 pairs of *in situ*-produced cosmogenic ^{10}Be and ^{26}Al show that the Pangnirtung Fjord region, southern Cumberland Peninsula, Baffin Island was ice covered during the late Wisconsinan. The Duval moraines, which appear to mark the most extensive ice advance in the region, formed between 25,000 and 10,000 years ago. The cosmogenic data imply that climatic and glaciologic conditions were sufficient to support an expanded ice margin in the eastern Canadian Arctic during the late Wisconsinan and that full fjord deglaciation occurred rapidly after about 10,000 years ago. The timing of deglaciation is constrained by exposure ages on recessional moraines, on an uplifted glaciomarine delta, and on valley bottom bedrock and erratics. Weathered erratics and bedrock tors above approximately 1000 m asl indicate that the highest regions were either ice free or covered by cold-based, non-erosive ice during the late Wisconsinan.

This is the first study to use cosmogenic nuclides to constrain the timing and extent of glaciation in the Canadian Arctic. The revised chronology I have determined based on cosmogenic nuclide dating for the Pangnirtung Fjord region is at odds with existing glacial chronologies based on relative dating. The previous chronologies suggest that this region was ice free during the late Wisconsinan and have been used to infer the maximum extent of the northeastern margin of the Laurentide icesheet and global climate records.

CITATION

Material for this thesis has been submitted for publication to *Geology* on March 18, 1997 in the following form:

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CHAPTER 1: Introduction

This thesis follows the guidelines of a 'journal article thesis' as outlined by the graduate college of the University of Vermont. The journal articles contain the methods and major results of this study. Chapter 1 provides an overview of the project and field area, and states the significance of my research. Chapter 2 contains a comprehensive literature review, focusing on the glacial history of the field area, as well as the method and theory behind cosmogenic isotope dating and a review of relevant cosmogenic isotope literature. In particular, I focus on other studies that have used cosmogenic isotopes to date glaciated terrain. Chapter 3 is a journal article submitted to *Geology*. This article focuses on the major implications of this study in a concise manner. The figures presented in this chapter utilize all of the data generated in this study. Chapter 4 is a journal article for submission to *Geological Society of America Bulletin*. This article focuses on a sub-set of the data in order to present a more detailed analysis of the results. A comprehensive bibliography follows the main text as does a set of appendices including: (A) elevation corrections used for samples, and (B) a table of all of the data generated, including raw isotopic measurements.

Purpose of study

This study produced exposure ages of glacial features in the Canadian Arctic, using *in situ*-produced cosmogenic isotopes ^{10}Be and ^{26}Al, in order to address the long-standing controversy concerning the vertical and lateral extent of the Laurentide icesheet during the last 60,000 years. The Pangnirtung Fjord area was chosen as the study site for three reasons: (1) the Duval moraines, a major moraine system, are well exposed and appear to separate two weathering zones (WZ-2 and WZ-3), (2) the area lacks adequate numerical dating control of the glacial landforms, and (3) the predominant lithologies are quartz-rich

Precambrian gneisses and quartz monzonites, which provide sufficient material (i.e. quartz) for cosmogenic isotope dating using ^{10}Be and ^{26}Al. Using over 140 pairs of ^{10}Be and ^{26}Al model exposure ages I have re-interpreted the glacial history of the Pangnirtung Fjord area, which has more wide-ranging implications for the extent of the Laurentide icesheet. In addition, my data challenge the weathering zone concept, a long-standing paradigm in arctic glacial chronologies, which is described in Chapters 2 and 3.

Field area

Baffin Island, the world's fifth largest island, is located in the eastern Canadian Arctic, west of the Greenland sub-continent (Fig. 1.1). Baffin Island belongs to the Arctic physiographic region referred to as the Shield, which also includes the southeastern coast of Ellesmere Island and eastern Devon Island. This region is underlain almost entirely by the ancient crystalline Precambrian Shield rocks. The mountainous zone along the eastern coast of Baffin Island contains most of the major ice fields and glaciers, with the exception of the Barnes Ice Cap, which is located near the center of the island. The upland areas along the eastern coast, as well as all coasts of Cumberland Peninsula, are dissected by steep-walled fjords, similar to the western coast of Greenland. These fjords have been the focus of many Quaternary studies, including this study of the Pangnirtung Fjord region, located on southern Cumberland Peninsula, adjacent to Cumberland Sound (Fig. 1.2).

A fjord is a narrow inlet of the sea between steep walls, with a mouth that opens into the sea and a head where the sea meets the land, typically at a steep-sided U-shaped valley. The origin of fjords along the eastern coast of Baffin Island has been the subject of some debate. Whereas most text books commonly describe fjords as the product of glacial scouring and subsequent submergence below sea-level (Flint, 1971), some researchers believe that the fjords on Baffin Island are primarily structurally controlled (Dowdeswell and Andrews, 1985). A similar argument has been suggested for fjords on other islands in the Canadian Arctic (England, 1987). Rifting between West Greenland and Baffin Island occurred in late Cretaceous/early Tertiary time, presumably resulting from an offshoot of the Mid-Atlantic Ridge. Models suggest that isostatic uplift associated with rifting (Hay, 1981) can account for the topographic asymmetry of Baffin Island, with the highest peaks occurring along a gently rising plateau toward the eastern coast. It is postulated that rifting and uplift caused incision along deep fault-controlled canyons, and that the fjords and troughs along the eastern coast of Baffin Island are the result of tectonic activity (Dowdeswell and Andrews, 1985). These regions of structural weakness were then exploited by glacial activity, which over-deepened and modified the shape of the valleys and fjords, leaving peaks, including many flat-topped summits, in between.

The Laurentide icesheet: theories and controversy

For the past few decades, Baffin Island has been the focal point of a debate concerning the vertical and lateral extent of icesheets in the eastern Canadian Arctic during late Wisconsinan time. Two different models for the configuration of the northeastern sector of the Laurentide icesheet during the late Wisconsinan (about 24 to 6 ka; ka = 1000 years before present, and will be used throughout this thesis) have persisted in the literature since the early 1940's (Ives, 1978). One reconstruction hypothesizes that a large, extensive icesheet with both land- and marine-based domes existed during the late Wisconsinan (Hughes *et al.*, 1977; Denton and Hughes, 1981; Grosswald, 1984), constituting what is commonly referred to as the "big ice" or "maximum ice" model. In contrast, many researchers (Miller, 1973; Miller and Dyke, 1974; Andrews, 1975; Dyke and Prest, 1987) believe that ice during the late Wisconsinan glaciation was limited to small, land-based icecaps, leaving the coastal zone of Baffin Island ice-free. This second hypothesis is often referred to as the "small ice" or "minimum ice" model. Figure 1.3 illustrates the contrasting icesheet configurations, showing the maximum and minimum reconstructions of the Laurentide icesheet in the late Wisconsinan, as compiled by Vincent and Prest (1987). Baffin Island marks the northeastern margin of the former Laurentide icesheet and therefore has been a major focal point for Quaternary research. However, the lack of reliable ages for glacial chronologies has contributed to the persistent controversy concerning the late Wisconsinan configuration of the Laurentide icesheet in this region. Numerous researchers have questioned the accuracy of mapped limits of late Wisconsinan ice (e.g., Ives 1978, Mayewski *et al.*, 1981) based on the reliability of the methods used for determining these limits.

In situ-produced cosmogenic isotopes

Cosmogenically produced nuclides are a useful tool for constraining landscape erosion rates, as well as exposure ages of landforms. Since Davis and Schaeffer (1955) first demonstrated that cosmic rays produce otherwise-rare nuclides in terrestrial material, advances in the fields of analytical chemistry and nuclear physics have allowed quantitative measurement of isotopes including those produced by the interaction of cosmic rays with rock and soil (Elmore and Phillips, 1987).

Primary cosmic rays are protons that continuously enter Earth's atmosphere. The flux of such protons is regulated by Earth's magnetic field, therefore deflection of primary cosmic rays is greater at the equator than at the poles. As these protons enter Earth's atmosphere, they collide with nuclei to produce atmospheric radionuclides (Lal, 1987), as well as a secondary shower of cosmic rays, primarily neutrons. The interaction of these secondary cosmic rays with target atoms in terrestrial material produces cosmogenic isotopes *in situ*. At sea level, the production rates of cosmogenic isotopes are generally very low, on the order of 10 to 100 atoms g^{-1} yr^{-1}. Production increases at high latitudes and elevations (Lal, 1991).

The production pathways of the commonly studied cosmogenic nuclides (^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al, and ^{36}Cl) include spallation, muon capture, neutron activation, and alpha-particle interaction (Fabryka-Martin, 1988). This study utilized the cosmogenic nuclides, ^{10}Be and ^{26}Al, which are produced primarily through spallation reactions. These spallation reactions involve the bombardment of high-energy neutrons that split a target atom into various particles. For instance, ^{10}Be is primarily produced through spallation of O, Mg, Si, and Fe, whereas ^{26}Al is produced primarily from the spallation of Si, Al, and Fe. The use of ^{10}Be and ^{26}Al together is ideal as they are both produced from spallation of the atoms that make up the mineral quartz, SiO_2. Quartz is resistant to weathering, common in many rocks, and often contains little native Al, which is important when making isotopic measurements. Early models suggesting that cosmogenic nuclides build up in surface materials over time (Davis and Schaeffer, 1955) have been verified by numerous studies (Nishiizumi *et al.*, 1986, 1989; Phillips *et al.*, 1986; Marti and Craig, 1987; Cerling, 1990) and demonstrations that spallation reactions primarily occur in the upper few meters of Earth's surface (Kurz, 1986a,b; Brown *et al.*, 1992; Fig 1.4).

When using cosmogenic isotopes to date glacial landscapes, an assumption is made that glacial erosion removes the upper few meters of rock and soil from the surface, producing a landscape that has no inheritance (cosmogenic isotopes from prior exposure). A further assumption is that during ice retreat, the surface is immediately exposed to cosmic ray dosing and therefore the calculated model exposure ages yield the time since deglaciation.

Significance of research

This study is significant for two reasons: (1) The data provide the first direct numerical ages on glacial landforms in the Canadian Arctic, and (2) this is the largest cosmogenic nuclide study to date, including over 140 pairs of ^{10}Be and ^{26}Al analyses.

Specifically, the exposure ages determined for glacial landforms in the Canadian Arctic are important for the following reasons:

- The data provide the first direct ages on moraines and weathering zones in the Canadian Arctic, bringing numerical chronology to a region that has thus far remained largely undated.
- The data challenge the weathering zone concept, which has been used extensively in the Canadian Arctic, Scandinavia, and Antarctica, to construct models of global icesheet extent.
- The data suggest that an expanded Laurentide icesheet persisted throughout the late Wisconsinan; this finding has important implications for climate modeling and source areas for Heinrich events.

In addition, the exposure ages described in this thesis help resolve discrepancies between the terrestrial and marine records around Cumberland Sound. A major landform within the field area is an extensive moraine system, the Duval moraines, that formed during a significant glacial advance on southern Baffin Island, which appears to represent the maximum ice advance on southern Cumberland Peninsula . However, the previously accepted early Wisconsinan age (ca. 60 ka) for the Duval moraines in the Pangnirtung Fjord area (Dyke, 1977, 1979) was at odds with recent marine evidence for late Wisconsinan (~10 ka) till on the floor of Cumberland Sound (Jennings, 1993, 1996). My data show that the Duval moraines date 10 to 22 ka, suggesting that the ice margin was expanded in both Pangnirtung Fjord and adjacent Cumberland Sound during late Wisconsinan time.

The previous discrepancies between the terrestrial and marine records emphasize the uncertainty in relative dating of terrestrial deposits and illustrate the need for assigning accurate numerical ages to these landforms. Cosmogenic isotope measurements provide the most direct age estimates to date for the deglaciation of southern Baffin Island and this study demonstrates the usefulness of utilizing cosmogenic age dating for refining glacial chronologies in the Canadian Arctic.

Although this project focused primarily on a single fjord, the implications are far reaching in the field of Quaternary studies. The late Wisconsinan (< 22 ka) age of the Duval moraines, determined from this study, changes the glacial chronology of the Pangnirtung Fjord area significantly. Glacial chronologies are used to estimate the maximum extent and volume of the Laurentide icesheet, constrain sea level histories, and infer paleoclimate. In addition, the weathering zone concept, the basis for many glacial chronologies, has been used extensively in the Canadian Arctic, maritime Canada, Scandinavia, and Antarctica without a clear understanding of the numerical age of weathering zones or the processes involved in their formation. This study is one of the first to date directly landforms that had been previously defined as occurring within of marking the break in distinct weathering zones on the basis of relative dating.

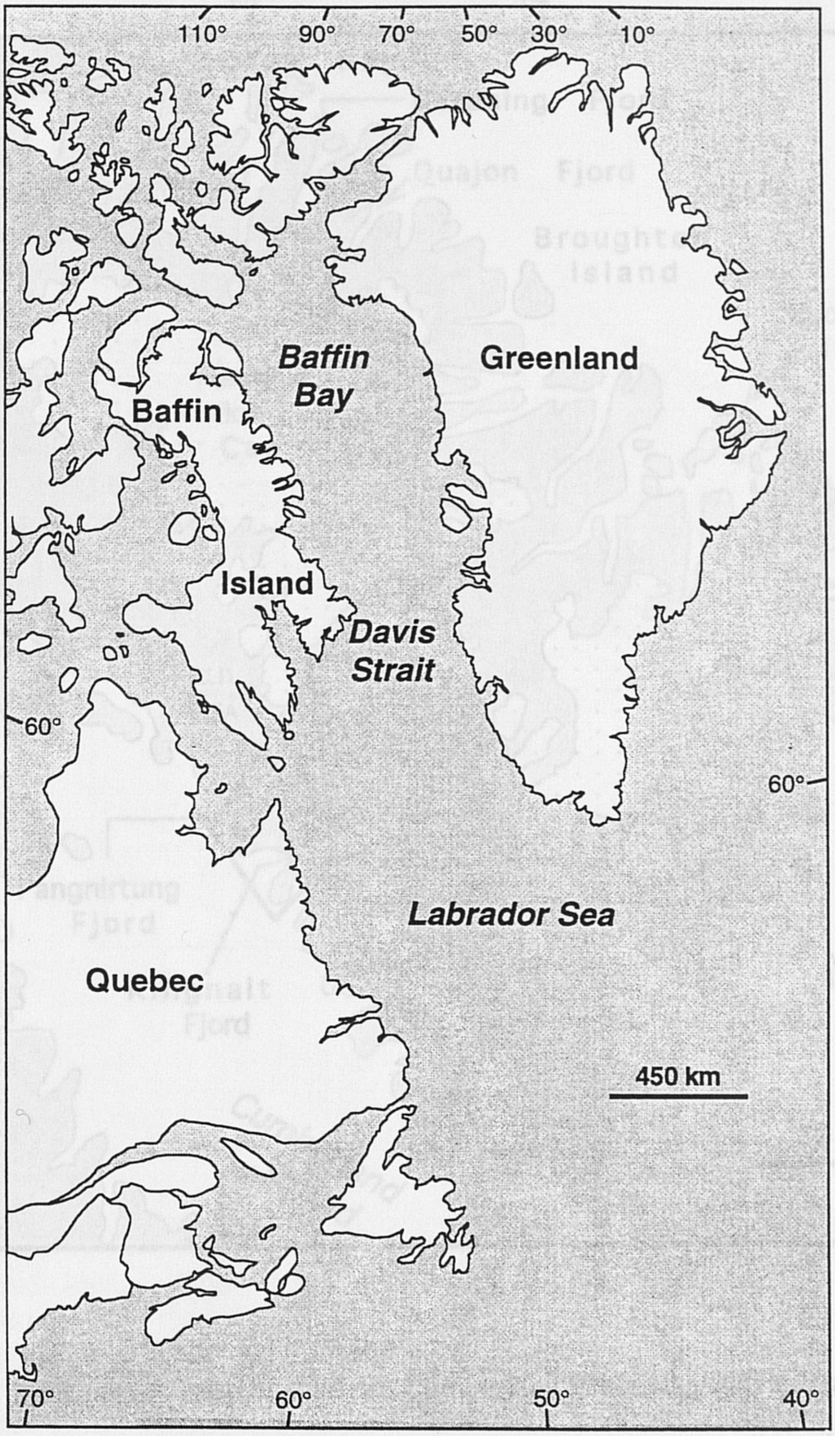

Figure 1.1 Location map of Baffin Island in the eastern Canadian Arctic.

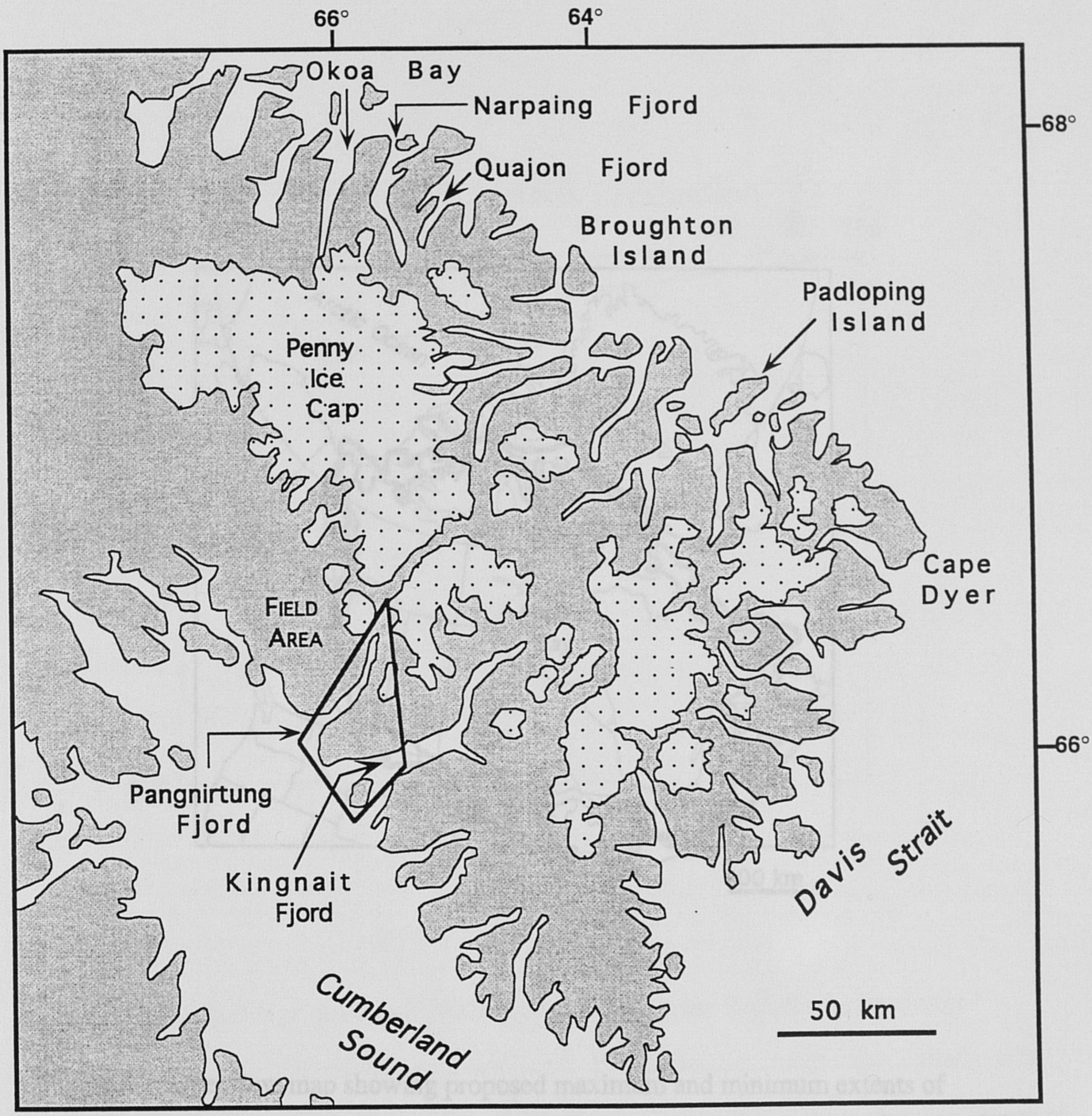

Figure 1.2 Location map of southern Cumberland Peninsula, Baffin Island. Field area is indicated by trapezoid with thick lines. Approximate ice cover indicated by stippled pattern.

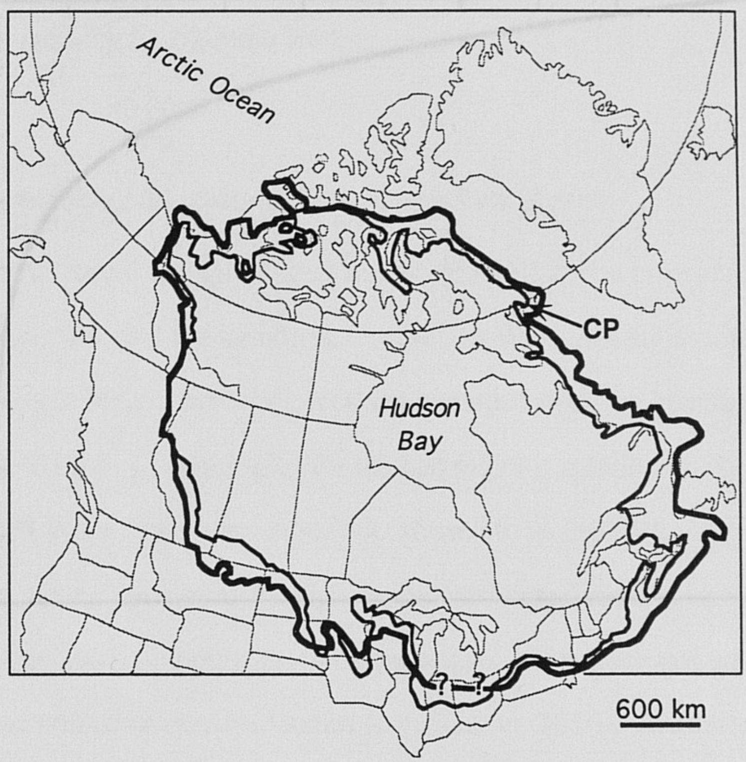

Figure 1.3 Overview map showing proposed maximum and minimum extents of Laurentide icesheet. The maximum extent (thick solid line) has been proposed as early Wisconsinan and the minimum extent (thin solid line) has been proposed as late Wisconsinan by many workers. Cumberland Peninsula on Baffin Island labeled CP (adapted from Vincent and Prest, 1987).

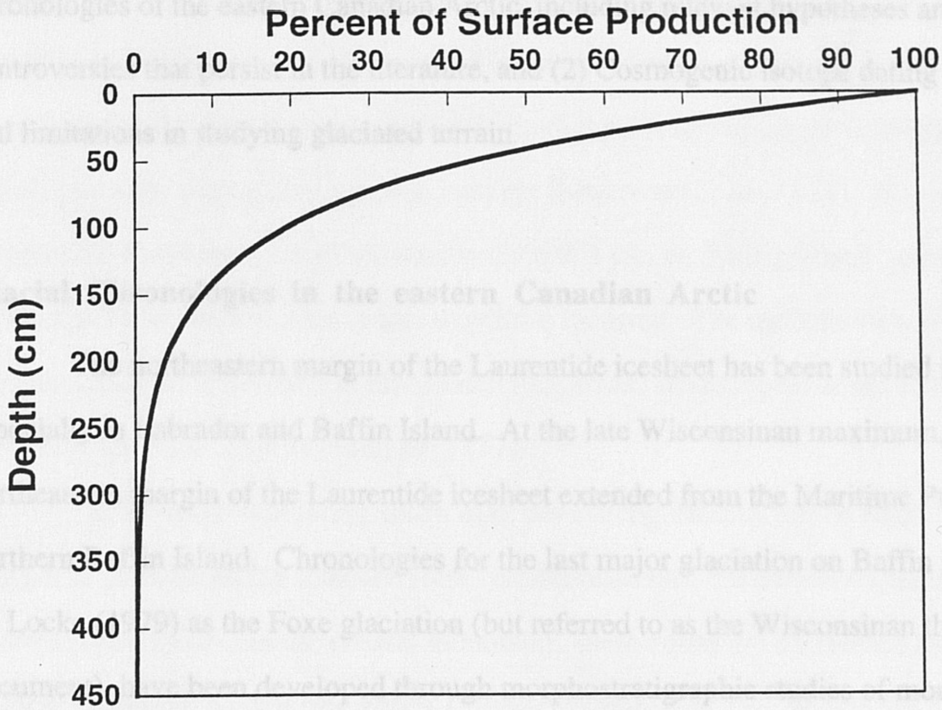

Figure 1.4 Percentage of surface production of cosmogenic nuclides for spallation reactions vs. depth below Earth's surface (cm). Production decreases exponentially with depth. Modeled using, $P_x = P_0 e^{-\rho x/\Lambda}$, $\Lambda = 165$ g cm^{-2}, $\rho = 2.7$ g cm^{-3}, for rock (after Bierman and Gillespie, 1995).

CHAPTER 2: Literature review

Introduction

This chapter considers literature relevant to this study in two sections: (1) Glacial chronologies of the eastern Canadian Arctic, including relevant hypotheses and controversies that persist in the literature, and (2) Cosmogenic isotope dating and its use and limitations in studying glaciated terrain.

Glacial Chronologies in the eastern Canadian Arctic

The northeastern margin of the Laurentide icesheet has been studied intensely, especially in Labrador and Baffin Island. At the late Wisconsinan maximum, the northeastern margin of the Laurentide icesheet extended from the Maritime Provinces to northern Baffin Island. Chronologies for the last major glaciation on Baffin Island, defined by Locke (1979) as the Foxe glaciation (but referred to as the Wisconsinan throughout this document), have been developed through morphostratigraphic studies of moraines and shoreline features, relative dating based on the degree of bedrock and surface boulder weathering, and soil development studies (e.g. Nelson, 1980). Radiocarbon ages and amino acid racemization ages of fossil mollusks have also been used extensively in assigning dates to Quaternary deposits and landforms (Miller, 1980; Miller *et al.*, 1977; Miller *et al.*, 1988; Nelson, 1981, 1982; Brigham, 1983; Mode, 1980; Locke, 1987). In addition, many studies have focused on the dating of raised marine deposits to reconstruct relative sea level and use isostatic uplift models to determine the timing and extent of glaciation on eastern Baffin Island (e.g., Pheasant and Andrews, 1973; Andrews, 1975; Dyke, 1977, 1979; Miller et al., 1977; Andrews, 1980; Brigham, 1983)

Weathering zone concept

Weathering studies in the eastern Canadian Arctic date back to the turn of the century (Daly, 1902), when investigators first began trying to differentiate glaciated from unglaciated terrains. The early work of Daly (1902), Coleman (1920), Odell (1933), and Tanner (1944) resulted in different interpretations of the observed "zones of weathering" and began a debate that persists today. Fieldwork conducted over the past two decades has contributed to the modern view of Wisconsinan ice extent that centers around the development of the weathering zone concept (Mayewski *et al.*, 1981). In general, weathering zones are units of the land surface that can be distinguished based on the amount of subaerial weathering that they have incurred. The modern view of weathering zones focuses on these units as altitudinally separated stratigraphic zones that record different lengths of exposure (Dyke, 1977).

Three broad weathering zones have been recognized along the entire northeastern margin of the Laurentide icesheet, from Newfoundland to the northern Queen Elizabeth Islands (Ives, 1978). In the coastal mountains, these zones are delineated by altitude, from highest to lowest, weathering zones 1, 2, and 3 (WZ-1, WZ-2, WZ-3; Fig. 2.1). The boundaries between these zones are thought, by most, to reflect the former margins of fjord glaciers (Boyer and Pheasant, 1974) and the upper limit of WZ-3 is commonly believed to represent the maximum extent of ice during the Wisconsinan glaciation (Ives, 1978). Some moraine systems within WZ-3 on eastern Baffin Island have been divided further, based on relative dating, into early Wisconsinan moraines (EWM), mid Wisconsinan moraines (MWM), and late Wisconsinan moraines (LWM; Figure 2.1). The debate concerning the vertical and lateral extent of the late Wisconsinan Laurentide icesheet has persisted largely because of different interpretations of these weathering zones.

Most weathering zone studies have focused on relatively small areas; however, researchers have presumed that weathering zones can be correlated over long distances. In correlating these units over distances of thousands of kilometers, researchers must make two assumptions: (1) in a particular area, each weathering zone represents subaerial weathering over a specific period of time, and (2) similar degrees of weathering seen in different regions represents weathering over a similar period of time. Such assumptions presume that the effects of variations in lithology, microclimate, or distance from ice source, are negligible.

Researchers over the past two decades have developed the weathering zone concept, in part, to determine glacial chronologies for the eastern Canadian Arctic (Pheasant and Andrews, 1973; Andrews *et al.*, 1975; Dyke, 1977, 1979; Miller *et al.*, 1977; Nelson, 1980). Figure 2.2 is a compilation of the previous glacial chronologies inferred along the eastern coast of Baffin Island (locations of all except Clyde Forelands are indicated on Fig. 1.2). The time-distance diagrams present a concept of progressively less extensive ice advances during the Wisconsinan glaciation, with the most extensive glaciation occurring in the early Wisconsinan (ca. 120 to 60 ka). The outermost limit of late Wisconsinan (ca. 8 ka) ice is mapped roughly parallel to the fjord heads on the eastern coast of Baffin Island.

An alternative hypothesis to the inference that altitudinal zones of increased weathering reflect increasing exposure time is based on a glaciological, rather than stratigraphic, interpretation of observed weathering zones (Sugden, 1977; Sugden and Watts, 1977). Sugden and Watts suggest that differing glacial thermal regimes could be responsible for the observed weathering zones. If cold-based ice persisted on mountain tops while warm-based ice flowed through the fjords and valleys, the lowermost weathering zone (WZ-3) would have been produced by active ice, which would be responsible for the relatively fresh appearance of rocks in this zone. Cold-based ice in the altitudinally highest zone (WZ-1) would have altered the landscape minimally, preserving the pre-glaciation landforms. The intermediate zone (WZ-2) represents a zone of transitional basal ice conditions, described by Sugden (1977) as a "warm-freezing" zone, where ice is at the pressure melting point but there is also some regelation. Sugden and Watts (1977) suggest that as ice passes from a warm-based regime (as in WZ-3) to a cold-based regime (as in WZ-1) erratics may be incorporated into the cold-based ice.

Interpretations of weathering zones as stratigraphic units led to the model of a restricted late Wisconsinan configuration of the Laurentide icesheet. In contrast, interpretations of the weathering zones as glaciologic units, which have been subaerially exposed for relatively the same period of time, support the model of an expanded late Wisconsinan configuration of the Laurentide icesheet. In both cases, individual observer biases and opinions seem to influence the interpretation of field data (Ives, 1978). Observer bias is difficult to avoid in relative weathering studies, as many of the criteria are qualitative. In addition, quantifying rates of weathering is difficult. For instance, questions such as how long does it take for weathering pits or tors to form, and what factors control the development of these features remain unanswered.

Uplift curves and glacial unloading

The deglacial history of icesheets has been inferred through studies of glacio-isostasy, the response of Earth's surface to the loading and subsequent unloading of ice and water (Andrews, 1975). If, theoretically, when an icesheet covers the land surface for a sufficient period of time an equilibrium state is achieved such that the surface is vertically displaced proportional to the density of the load (ρ_i) and the density of the mantle (ρ_m) in the ratio ρ_i/ρ_m. The displacement of Earth's surface that results is directly related to the thickness (h) of the ice, $h\ (\rho_i/\rho_m)$. With the onset of deglaciation in coastal areas, the sea transgresses across the recently depressed surface until such time as the surface becomes elevated by the isostatic response from the removal of the ice load (Andrews, 1980).

Evidence for glacio-isostatic uplift comes from coastal areas and large interior lakes where raised features such as shingle beaches and glaciomarine deltas mark the former marine limit, which marks the maximum level that the sea transgressed across an area (Andrews, 1980). It has been assumed that during the late glacial and post-glacial period the coastal zones covered by the retreating Laurentide icesheet were in contact with the sea and therefore the date of deglaciation should be synchronous with the time of formation of marine limit. The marine limit can be demarcated by different features: (1) a glaciomarine delta formed by glacial meltwater and sediments draining into a higher sea level, (2) the upper limit of perched erratic boulders that have been dislodged by wave action, (3) the uppermost beach ridge, (4) the highest occurrence of marine fossils, and (5) the lowermost limit of grounded moraines (Andrews, 1980). However, not all of these features are preserved in all coastal regions. In the eastern Canadian Arctic, differing marine limits from fjord to fjord indicate a complex ice retreat pattern. In addition, if ice retreat occurred early and ice load were limited, the original marine limit may be obliterated as sea level transgressed forming a higher, younger marine limit (Andrews, 1975).

The preservation of marine shells and wood in raised glaciomarine deposits has allowed for radiocarbon and amino acid racemization (AAR) dating of these deposits. Assuming that the date of marine limit formation is synchronous with the date of deglaciation over short intervals, researchers have constructed numerous elevation-time graphs and relative sea-level curves for the eastern coast of Baffin Island. The sea-level history of the Cumberland Sound region is summarized by Dyke (1979), who acknowledges that there is a lack of adequate samples that would allow for the construction of a chronologically well-controlled relative sea level curve. Figure 2.3 shows what Dyke describes as the "speculative" sea-level curves for the Cumberland Sound region, including Pangnirtung and Kingnait Fjords. Dyke hypothesizes that the sea-level history for the head

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A major observation along the eastern Canadian Arctic coast has been an unconformity between sediments dated ≥ approximately 45-50 ka and the overlying early Holocene marine transgression sediments, which primarily date between 8 and 10 ka (Andrews, 1975; Miller *et al.*, 1977). Researchers along the eastern coast of Baffin Island have noted a consistent lack of sediments dating in the interval between about 10 and 45 ka (Andrews, 1975). One explanation for this gap could be that the coast was inundated with glacial ice during this time, suggesting that ice was more extensive than the present coast during the late Wisconsinan (e.g., Blake, 1966). However, Andrews (1975) believes that the ice did not extend beyond the fjord heads, pointing out that there is a lack of till associated with the hiatus in sediments dated > 45 ka and the early Holocene sediments. Andrews (1975) suggests that sea level was below present level between 10 and 30 ka, arguing that many radiocarbon ages are > 26 ka from elevations between 10 and 80 m asl along the outer coast.

Southern Cumberland Peninsula

Dyke (1977, 1979) studied a large area, primarily southwest of the Penny Ice Cap (Fig. 1.2), and initially mapped what he termed the Duval moraines and related features along Pangnirtung Fjord and near the mouth of Kingnait Fjord (Fig. 2.4). Dyke's study is one of the most comprehensive relative weathering studies to date, as he attempted to use specific, quantifiable characteristics to assign boulders and bedrock to different classes of weathering based on a 9-point classification scheme. Dyke (1977,1979) concluded that the Duval moraines altitudinally differentiate an older, more extensive glaciated terrain from a younger glaciated terrain produced by a restricted ice margin. Dyke asserted that the Duval moraines date between 100 and 60 ka based on amino acid data from a single disarticulated mollusk fragment on a nearby, stratigraphically related glaciomarine delta. However, the shell fragment from the delta was not *in situ*, therefore the age of the delta and the Duval moraines can not be reliably inferred, as the shell fragment may have been reworked (Jennings, 1993).

The Ranger moraines (Fig. 2.2), located at some fjord heads, are believed to delimit the maximum extent of late Wisconsinan (ca. 8 to 5 ka) ice in the southern Cumberland Peninsula area; however, moraines have not been identified at the head of Pangnirtung Fjord. The glacial chronology for southern Cumberland Peninsula, as proposed by Dyke (1977, 1979) and characterized by progressively less extensive glaciations throughout the Wisconsinan, is consistent with the chronologies proposed for the northern side of Cumberland Peninsula (Miller, 1973; Boyer and Pheasant, 1974; Nelson, 1980; Brigham, 1983), as well as other chronologies shown in Figure 2.2 from the eastern coast of Baffin Island.

Mayewski *et al.* (1981, p. 115) question the Cumberland Sound chronologies, based on the reliability of the radiocarbon ages, and assert that restricting the extent of late Wisconsinan glaciers to fjord heads on the basis of the Cumberland Sound chronologies is problematic. Furthermore, the minimum ice model reconstruction for the late Wisconsinan configuration of the Laurentide icesheet, which has its roots in the eastern Canadian Arctic, is controversial based on the reliability of the dating constraints (Mayewski *et al.*, 1981).

Prior to the work of Jennings (1989, 1993) and Jennings *et al.* (1996), the glacial history of the Cumberland Sound region was interpreted entirely from terrestrial records (Miller, 1973; Dyke, 1977, 1979; Dyke *et al.*, 1982). While the terrestrial record shows that Cumberland Sound served as a drainage way for regional and local ice, previous interpretations suggest that ice did not extend into Cumberland sound in at least the last 70,000 years. However, Jennings' (1989, 1993) work on sediment cores from

Cumberland Sound demonstrates that the sequence of till overlain by ice-proximal to ice-distal glaciomarine sediments was deposited during the late Wisconsinan (~10 ka). The marine record suggests that Cumberland Sound ice advanced to its maximum position by about 11 ka, then retreated up the Sound by 10.2 ka. In addition, evidence from the marine sediment cores shows that an earlier advance of Cumberland Sound ice took place sometime after 34 ka, but was less extensive than the younger advance (Jennings *et al.*, 1996). The Cumberland Sound ice reconstructions presented in Jennings (1993) and Jennings *et al.* (1996) provide evidence that ice extended well beyond the previously mapped ice margin in Cumberland Sound during the late Wisconsinan; however, it was difficult to reconcile this new evidence with the alleged early Wisconsinan age of glacial features in Pangnirtung and Kingnait Fjords (Fig. 2.5).

Data from marine sediment cores from some areas adjacent to Baffin Island suggest that late Wisconsinan ice deposited till on parts of the continental shelves (MacLean, 1985; MacLean *et al.*, 1986; Praeg *et al.*, 1986). This evidence may support a late glacial maximum extending beyond the fjord heads on other parts of Baffin Island as well.

Exposure dating using cosmogenic isotopes

Previous geomorphic studies have primarily used six *in-situ* produced cosmogenic isotopes: ^{3}He, ^{10}Be, ^{14}C, ^{21}Ne, ^{26}Al, and ^{36}Cl. Recent studies have shown that, particularly in glaciated regions, *in situ* -produced cosmogenic isotopes provide useful information about exposure ages and histories of the geologic surfaces (e.g. Cerling, 1990; Phillips *et al.*, 1990; Nishiizumi *et al.*, 1991; Brook *et al.*, 1993, 1996).

A sampled surface that has been exposed to cosmic dosing relatively rapidly and has had minimal erosion since exposure will yield an isotope abundance that is a function of the exposure time and the isotopic production rates (Bierman, 1994). The model exposure age is calculated using equation 1, in which N is the measured isotopic abundance after background correction (atm g^{-1}), P is the isotopic production rate (atm g^{-1} y^{-1}), λ is the decay constant for unstable isotopes (y^{-1}), and t represents the time of exposure (y):

$$t = \frac{-\ln\left(1-\frac{N\lambda}{P}\right)}{\lambda} \qquad (1)$$

Assumptions

The three most important assumptions in the interpretation of measured isotope abundances as model exposure ages are: (1) there is a known and constant rate of isotopic production, (2) at the time of initial exposure, there is no isotopic inheritance (i.e. from prior exposure history), and (3) there has been no significant erosion of the sampled surface since exposure. I discuss each of these assumptions below.

<u>1. Production rates</u>

Field based studies have been used to calibrate production rates of commonly used cosmogenic isotopes, including ^{3}He (Cerling, 1990; Kurz *et al.*, 1990), ^{21}Ne (Hudson *et al.*, 1991), ^{10}Be and ^{26}Al (Nishiizumi *et al.*, 1989; Brown *et al.*, 1991a; Larsen, 1996), and ^{36}Cl (Zreda et. al., 1991). These studies empirically determined production rates on surfaces that have ages constrained by other dating techniques. Such calibration integrates isotope production rates over the time period of calibration.

Recent studies (e.g., Meynadier *et al.*, 1992) have shown variations in the strength of Earth's magnetic field; therefore, production rates of cosmogenic nuclides have fluctuated over time. The effect of variations in Earth's magnetic field strength are pronounced at the equator, but decrease poleward. Therefore, the effect of field strength changes in polar regions is minimal (Lal, 1991). Because Baffin Island is at high latitude, changes in the production rate over time have a minimal effect on calculated model exposure ages.

This thesis utilized the ^{10}Be and ^{26}Al production rates of Nishiizumi *et al.* (1989), 6.03±0.3 and 36.8±2.7 atoms g^{-1} y^{-1}, respectively. While the established production rates of Nishiizumi *et al.* (1989) are commonly used in cosmogenic studies, recent studies (Clark *et al.*, 1996; Larsen, 1995; Bierman *et al.*, 1996a) suggest that these production rates may be 10 to 20% too high. The effect of using production rates that are too high will result in calculated model exposure ages that are too young by a similar percentage.

2. Inheritance

The assumption of no isotopic inheritance has been investigated by few workers. Brook *et al.* (1993) suggest that inheritance is not likely to be important for Antarctic glaciers, based on an exposure age of 9 ka from debris on the modern glacier, which when compared to the age of the moraines (> 100 ka) is minimal. Work by Bierman (1993) also indicated that minimal isotopic inheritance occurred in boulders contained in Sierra Nevada moraines. As part of this study, six of seven samples collected from a striated bedrock surface exposed after a glacier retreated following the Little Ice Age have very low isotope abundances (indistinguishable from blanks), suggesting that even short-lived glacial advances are capable of removing isotopic evidence of prior cosmic ray exposure.

3. No erosion

Uncertainties associated with the assumption of no erosion (or minimal erosion) can be constrained using isotopes with differing half-lives. For the ^{10}Be:^{26}Al system, however, this works only in the 0.5 to 3 my range (Brown *et al.*, 1992; Gillespie and Bierman, 1995). For samples younger than 0.5 my, independent estimates on sample erosion rates must be made. In glaciated terrains, investigators have sampled outcrops on which glacially polished and/or striated surfaces are preserved (e.g., Gosse *et al.*, 1995a). On these landforms, the amount of rock surface degradation can be constrained to 1 to 2 cm, which will have a minimal influence on calculated exposure ages (1 to 2%); however it may be difficult to assess if polish or striae were covered by till, soil or snow for an appreciable length of time (Bierman, 1994). Determining a maximum erosion rate for a region, determined using the ^{10}Be:^{26}Al systematics for old samples, is one way to constrain erosion on younger boulders. In the Pangnirtung Fjord area, cosmogenic model erosion rates suggest surfaces are eroding at a maximum of 1.1 meters per million years (Bierman *et al.*, in submission).

Related Studies

Moraines provide the most significant terrestrial records of glacial history; however until recently, little numerical age constraint was available for most important moraine sequences. In this section, I present some of the important cosmogenic studies on moraines that are relevant to this thesis.

Western United States

Phillips *et al.* (1990) used the abundance of *in-situ* produced ^{36}Cl to calculate the exposure ages for 31 boulders on five moraines in the Sierra Nevada of southern California. The relative age of these moraines had already been determined through numerous studies that utilized weathering characteristics along with distinct cross-cutting relationships; however Phillips *et al.* (1990) provided the first direct dating of these moraines. Their ^{36}Cl data tend to cluster tightly in the younger moraines, while the older moraines tend to show more scatter in model ages. Phillips *et al.* (1990) used the maximum boulder ages on moraines to assign ages and determine that the ages of the three youngest moraines, the Younger Tahoe, Tenaya, and Tioga, are 65, 24 and 21 ka, respectively, in agreement with their stratigraphic positions. The ages determined for the two oldest moraines, the Mono Basin moraines (80-119 ka) and the Older Tahoe moraines (130-150 ka) conflict with the cross-cutting relationship between these moraines.

Phillips *et al.* (1996) summarize the ^{36}Cl exposure ages from the Sierra Nevada moraines and attempt to correlate these ages with radiocarbon-dated sediment cores from nearby Owens Lake. The ^{14}C chronology from Owens Lake had previously been correlated with climatic events indicating glacial advances associated with Heinrich events 5, 3, 2, and 1 (Bond *et al.*, 1993), identified by layers of terrestrial sediments in cores from the North Atlantic ocean. Heinrich events result from expulsions of large volumes of icebergs from periodic collapses of the Laurentide icesheet, and represent rapid changes in global climate (Bond *et al.*, 1993). Phillips *et al.* (1996) correlated the mean of the maximum boulder ages on the younger Sierra Nevada moraines with peaks in magnetic susceptibility and organic carbon from the sediment record of Owens Lake, suggesting glacial advances at 49±2, 31±1, 25±1, 19±1, and 16±1 ka. These glacial advances fall within or around Heinrich events 5 (47 ka), 3 (27 ka), 2 (20 ka), and 1 (14.5 ka); however, based on the uncertainty in both the ^{36}Cl ages and the Heinrich event ages, exact correlation is not possible. Phillips *et al.* (1990, 1996) use their ^{36}Cl chronology for the Sierra Nevada to suggest that Heinrich events represent rapid global climatic changes and they suggest that mountain glaciers may be more sensitive to changes in climate than continental icesheets.

In the Wind River Mountains of Wyoming, Gosse *et al.* (1995a) used ^{10}Be measurements on 44 boulders from the type locality of the Pinedale glaciation to determine that the last glacial maximum in this area was reached by 21,700 ^{10}Be years, lasted for 5900 years, and was followed by rapid deglaciation. They used the maximum exposure age (21.7 ka) of 19 boulders from the terminal moraines to date the last glacial maximum. The range of boulder ages on this moraine was 5900 years, which led them to conclude that the boulders were delivered to the moraine over that time period. They ruled out the possibility of differential erosion of the boulders on the basis of glacial polish preserved on some surfaces, indicating that less than 2 mm of erosion has occurred over approximately

20,000 years. Boulders dated from four nested recessional moraines yielded ages that overlapped one another. Gosse *et al.* (1995a) suggest that the recessional moraines were deposited rapidly, as the uncertainty associated with the mean age of the recessional moraines is 900 years, which is beyond the resolution of the ^{10}Be technique to date a single deposit.

Cosmogenic ^{10}Be measurements (Gosse *et al.*, 1995b) from the Inner Titcomb Lakes moraine in the Wind River Mountains of Wyoming have been used by Gosse *et al.* to suggest that the Younger Dryas cooling event (dated 12.8 to 11.5 ka) can be recognized in the western hemisphere and may indeed have been a global event. There evidence comes from nine boulders deposited on the crest of the Inner Titcomb Lakes moraine that yield ^{10}Be ages indicating the moraine was deposited between 11.4 ± 0.5 to 13.8 ± 0.6 ka, thereby overlapping the North Atlantic Younger Dryas. The range of ages on the moraine reflects the uncertainty in rock erosion rate, snow cover, and the ^{10}Be production rate. Minimum ages were calculated using the model of no erosion, no snow cover and the ^{10}Be production rate of Nishiizumi *et al.* (1990), while maximum ages included a modeled effect of snow cover, erosion and the lower ^{10}Be production rate of Clark and Gillespie (1994). While Gosse *et al.* (1995b) suggest that their dating is precise enough to represent a cooling event that took place over a period of 1.3 ka, their data as reported in their abstract have an age range of 2.4 ka; however in their Table 1 (Gosse *et al.*, 1995b, p. 879), the actual range of ages (from 10 not 9 boulders) from the Inner Titcomb Lake moraines is 8.8 ± 0.3 to 14.6 ± 0.4, spanning 5.8 ka. In addition, if the ^{10}Be production rates of Clark and Gillespie (1994) are correct, then 8 of the 10 boulders on the moraine are over 1000 years older than the end of the Younger Dryas event. In addition, it is not known whether the Inner Titcomb Lakes moraine was deposited during a re-advance or whether it represents a late phase of retreat from the glacial maximum. The work of Gosse *et al.* (1995b) provides a good age estimate for the deposition of the Inner Titcomb Lakes moraine that is in general stratigraphic accordance with ^{10}Be ages for samples stratigraphically above and below the moraine. The uncertainty in the ^{10}Be production rate and the large range in ages from the moraine however, make it difficult, if not impossible, to correlate to a global cooling event that spanned 1.3 ka until production rates are better known.

Hallet and Putkonen (1994), in reference to the work of Phillips *et al.* (1990), suggest the possibility that mean exposure ages on younger moraines could exceed those on an older moraine through erosion on the moraine surface, which could remove the oldest boulders and expose fresh boulders over time. For moraines that have eroded over time, boulders on moraines could show a wide range of ages, including ages that are younger than the actual deposition of the moraine, and the mean exposure age can only be considered a minimum. On younger moraines, this process would not greatly affect the calculated exposure age of the moraine. This paper is important in that it incorporates a geomorphic process model into the interpretation of calculated exposure ages, and provides insight into the range of ages that have been reported on individual moraine sequences in the literature. In the case of the "old" Bloody Canyon moraines in the Sierra Nevada, this model may explain the range of ages on a single moraine and the apparent inconsistency of the ^{36}Cl isotope data (Phillips *et al.*, 1990) with accepted geological interpretations, but it is unclear if such a model would also apply to the range of ages measured on the relatively young Pinedale moraines (Gosse *et al.*, 1995a).

<u>Antarctica</u>

Brown *et al.*, (1991b) and Brook *et al.* (1993) used measurements of *in-situ* produced ^{3}He, ^{10}Be, and ^{26}Al to calculate model exposure ages of a suite of moraines in the Arena valley area, adjacent to Taylor Glacier in West Antarctica. The ^{10}Be and ^{26}Al model exposure ages correlate fairly well, ranging in age from 0.117 to 2.1 myr, although the data are noisy. The ^{3}He model exposure ages determined for the same surfaces show significant He loss compared to the ^{10}Be and ^{26}Al dates. Although there is large scatter in the data, the exposure ages of the moraines follow the relative age sequence of the area. Brown *et al.* (1991a) and Brook *et al.* (1993) found that all of the moraines exhibit a broad distribution of exposure ages. Brown *et al.* (1991a) report that the exposure ages on the older moraines are within experimental uncertainty of one another, whereas the younger samples exhibited less isotopic concordance within each moraine. These workers propose that younger moraines in this area may contain some boulders that have prior exposure histories, which could result if boulders were exposed in cliff outcrops prior to incorporation into the glacial system. Brown *et al.* (1991a) and Brook *et al.* (1993) also suggest that the older outliers may represent intermixing of boulders from earlier glacial events if younger glacial advances incorporated boulders from older moraines.

Although the data of Brown *et al.* (1991a) and Brook *et al.* (1993) are noisy and may be affected by inheritance in the samples, they provide the first exposure ages for moraines that previously lacked chronological control. Their studies have correlated cosmogenic ages with accepted glacial chronologies and suggest an expanded Pliocene ice sheet, which is supported by ^{21}Ne data (Staudacher and Allegre, 1991). Brook *et al.* (1993) suggest that significant variability in cosmogenic ages within single deposits may reflect prior soil cover, prior cosmic ray dosing, sample movement, or perhaps differential erosion rates on individual boulders.

Scandinavia

Brook *et al.* (1996) used cosmogenic ^{10}Be and ^{26}Al model exposure ages of four bedrock samples to date weathering zone surfaces in western Norway. A distinct altitudinal boundary exists between relatively unweathered glaciated terrain and summit regions of highly weathered bedrock in the field area. The purpose of this study was to assess the origin of these zones, addressing the debate over whether or not weathering zones indicate former icesheet heights or different englacial basal ice regimes. The preliminary results of this study suggest that the upper, weathered bedrock surface has a model exposure age of >55 ka, whereas the surface below the blockfield boundary has model exposure ages between 26 and 21 ka, and the surface below a lower trimline has model exposure ages between 18 and 13 ka. Brook *et al.* (1996) suggest that their data support the concept that weathering zones in this region indicate former icesheet heights and that these zones are formed from progressively less extensive ice advances, as the model exposure ages become progressively older with increasing altitude. While this is the first report using cosmogenic nuclides to assign ages to relative weathering zones, the data are extremely limited (four samples: one from the above the blockfield boundary, one from above the lowest trimline and two from below the lowest trimline), and so it is not possible to know how well these samples represent the age distribution within each weathering zone.

Canadian Arctic

McCuaig *et al.* (1994) used ^{10}Be and ^{26}Al measurements to estimate the glacial history on southern Bylot Island in northern Baffin Island. The ^{10}Be model ages of all the erratics (native and foreign origin) clustered around 60 ka (authors do not indicate number of samples, or actual age estimates), suggesting that the northeastern margin of the Laurentide icesheet reached its maximum in this area during the early Wisconsinan. The ^{10}Be ages were not consistent with ^{26}Al ages (>2 ma) on the same rock samples, suggesting that the analyses by McCuaig *et al.* (1994) may not be robust. These data represent the only published use of cosmogenic nuclides for dating glacial surfaces in the Canadian Arctic.

Southern margin of Laurentide icesheet

Larsen (1996) sampled both bedrock and boulders from the well-dated terminal moraine of the Laurentide icesheet in New Jersey. This study determined production rates of ^{10}Be and ^{26}Al over the last 21,500 years. Sixteen samples were collected including quartzite and gneiss samples from either large boulders or glacially striated and/or molded bedrock. The production rates calculated by Larsen (1996) for ^{10}Be and ^{26}Al, 5.17±0.15 and 30.40±1.01 atoms g^{-1} yr^{-1}, respectively, suggest that the currently accepted production rates for these isotopes over the late Pleistocene (Nishiizumi *et al.*, 1989) are 10-20% too high. In addition to making another estimate of ^{10}Be and ^{26}Al production rates, Larsen's data demonstrate good isotopic concordance both among different rock types and, more important, between bedrock and boulder erratics from the same glaciated surface. His data show that statistically, for samples from the terminal moraine of the Laurentide icesheet, bedrock and boulder samples may be considered as a single sample population and differences in sample lithology do not affect isotopic abundance. The work of Larsen (1996), Clark *et al.* (1995), and Bierman *et al.* (1995) illustrates that the uncertainty in isotope production rates (10-20%) exceeds greatly that of analytical precision (typically 3-5%) and suggests that more isotopic calibration studies of production rates are needed.

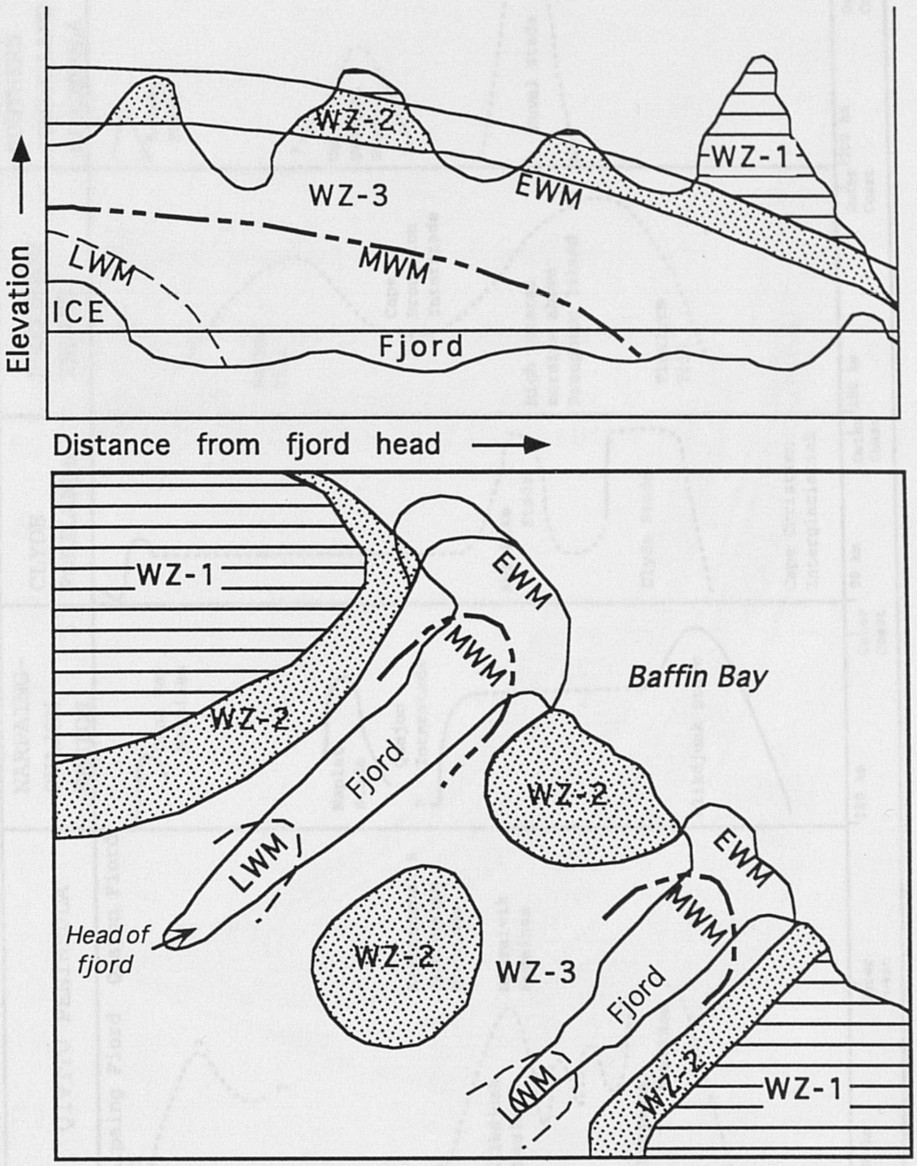

Figure 2.1 A schematic portrayal of the Quaternary glaciation model for eastern Baffin Island and Cumberland Peninsula. The upper diagram shows the vertical distribution of lateral moraines and weathering zones on a longitudinal profile The lower diagram shows the same in planimetric view of highlands dissected by fjords. WZ=weathering zone, EMW=early Wisconsinan moraine, MWM=mid Wisconsinan moraine, LWM=late Wisconsinan moraine. (adapted from Dyke, 1977)

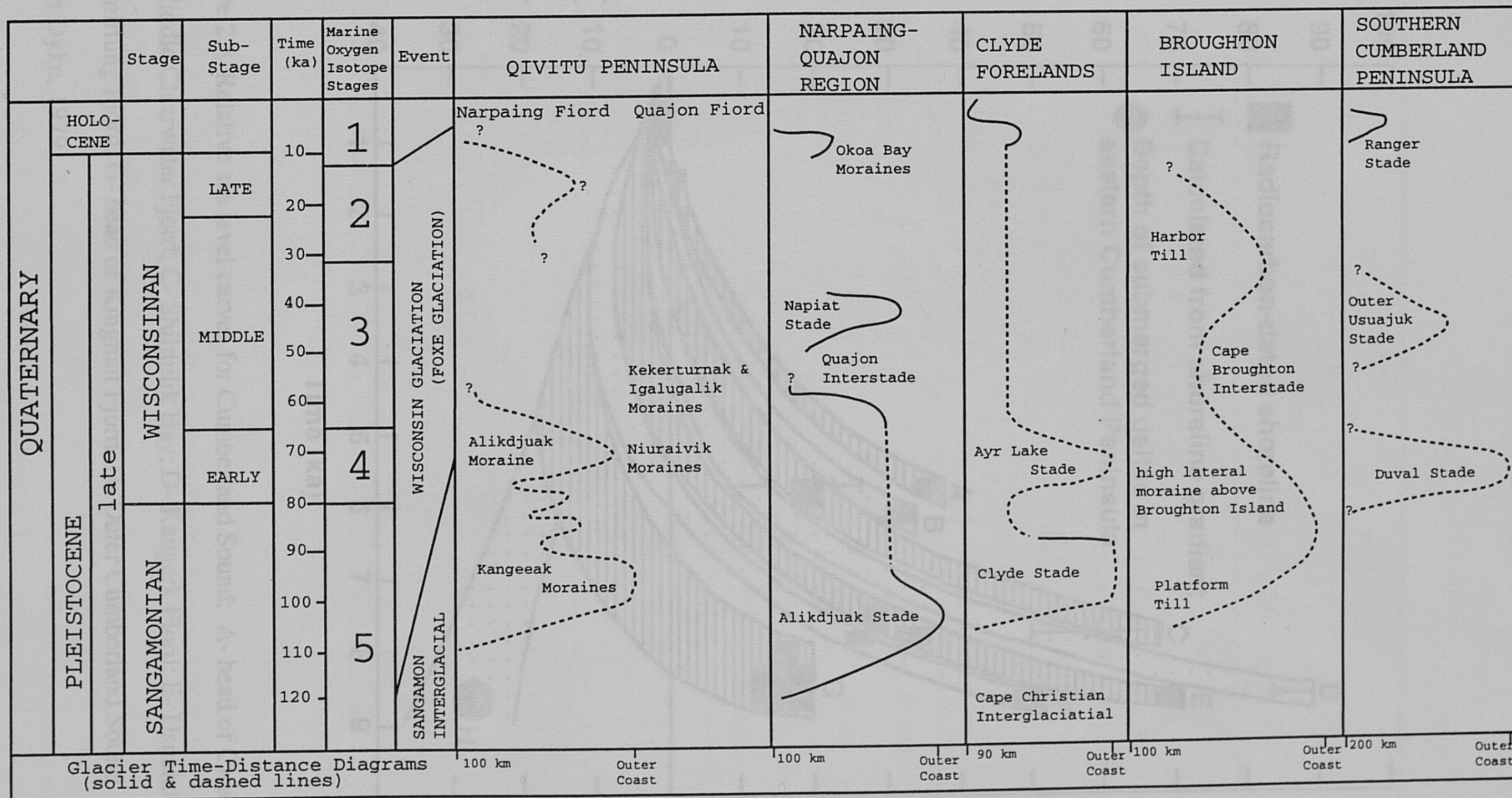

Figure 2.2 Glacial time-distance diagrams showing the correlation of tills, moraines and glacial events on eastern Baffin Island (adapted from Nelson, 1980; Pheasant and Andrews, 1973; Andrews *et al.*, 1973; Miller *et al.*, 1977; Andrews *et al.*, 1976; Dyke, 1977, 1979).

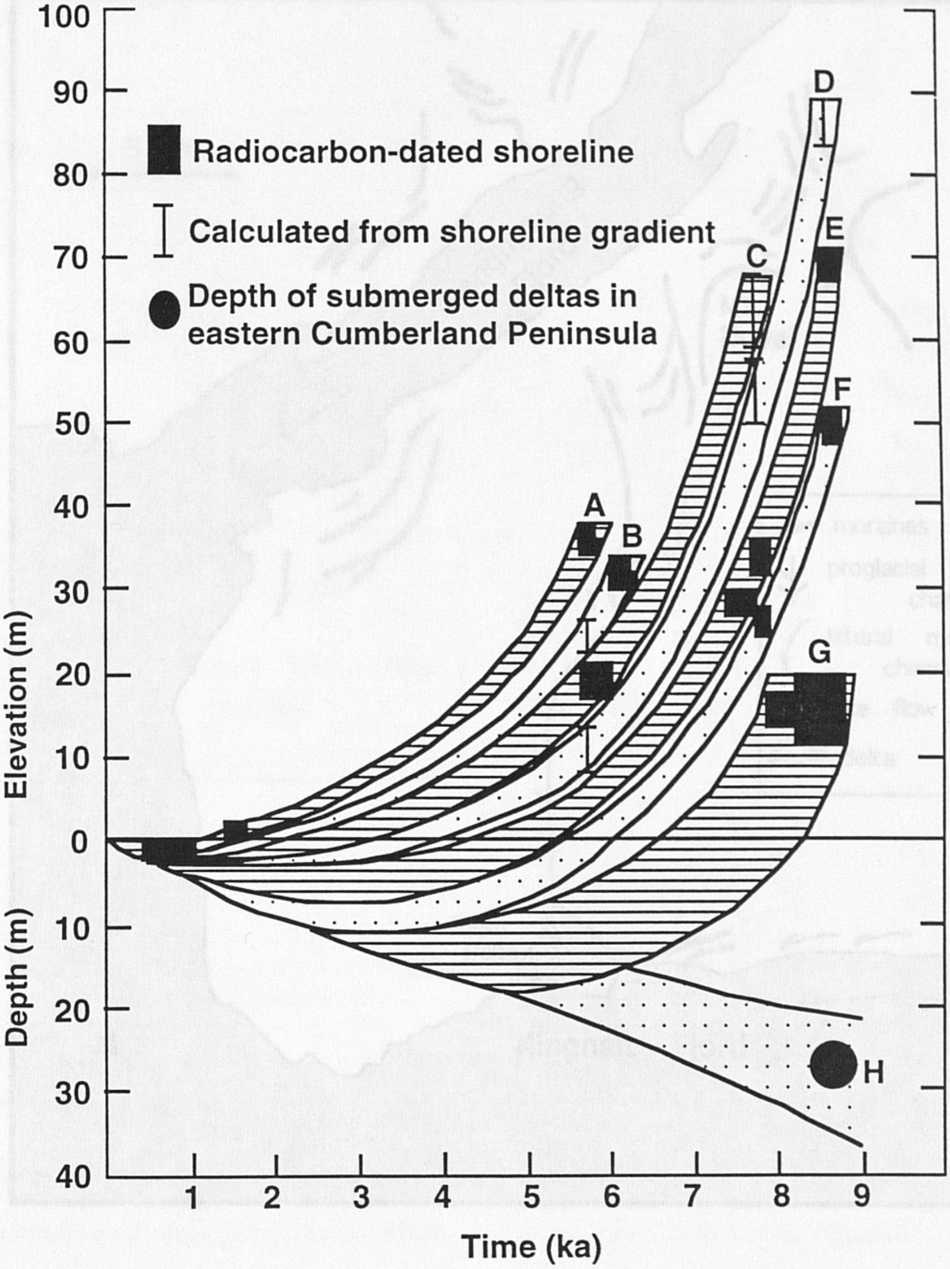

Figure 2.3 Relative sea level curves for Cumberland Sound: A- head of Clearwater Fjord; B- Middle Clearwater Fjord; C- Shilmilik Bay; D- Kangerk Fjord; E- Usualuk Fjord; F- Pangnirtung Fjord; G- head of Kingnait Fjord; H- Outer Cumberland Sound (from Dyke, 1979).

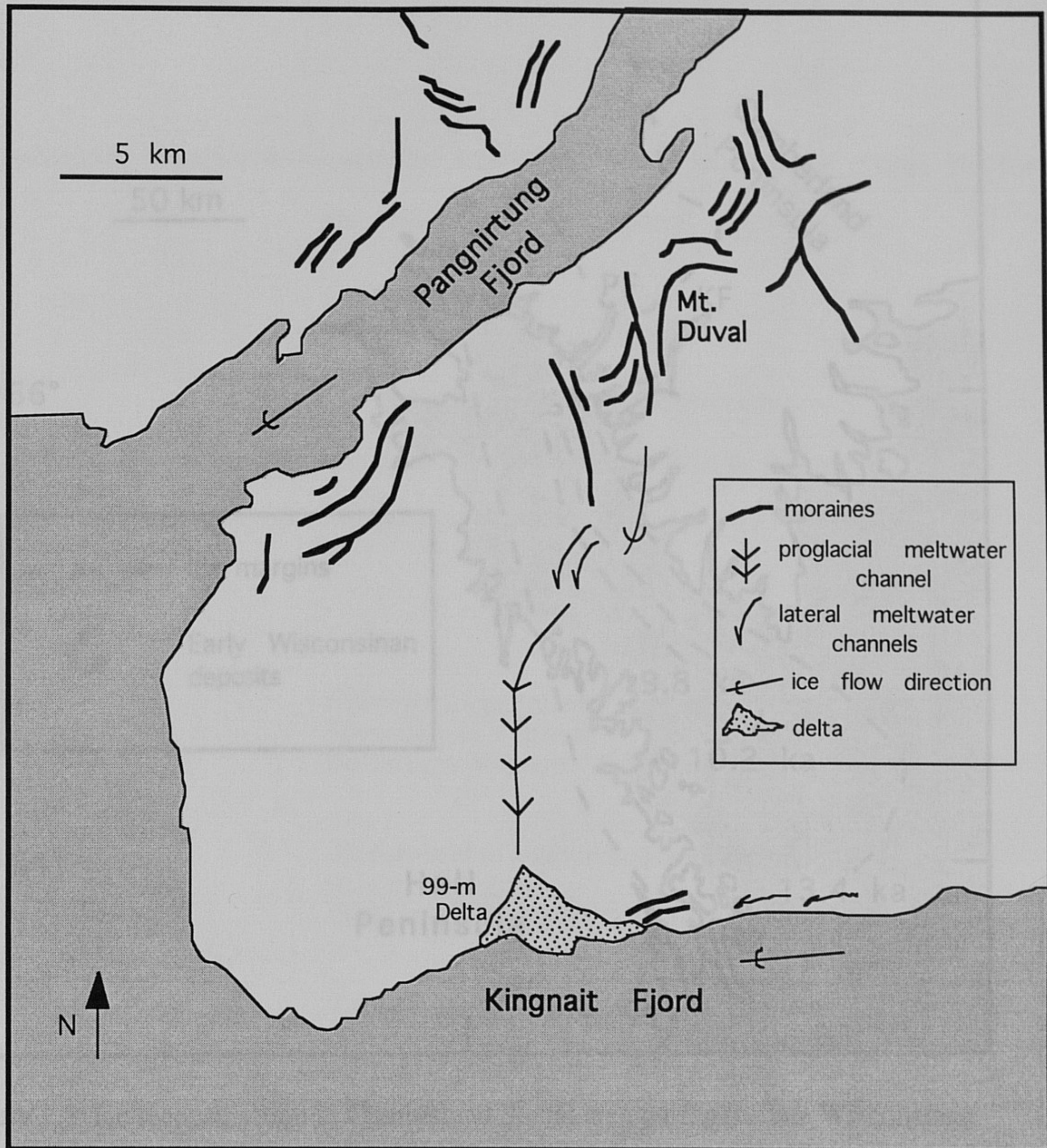

Figure 2.4 Relationship of the Duval moraines to the 99-m marine limit delta in outer Kingnait Fjord (adapted from Dyke, 1979).

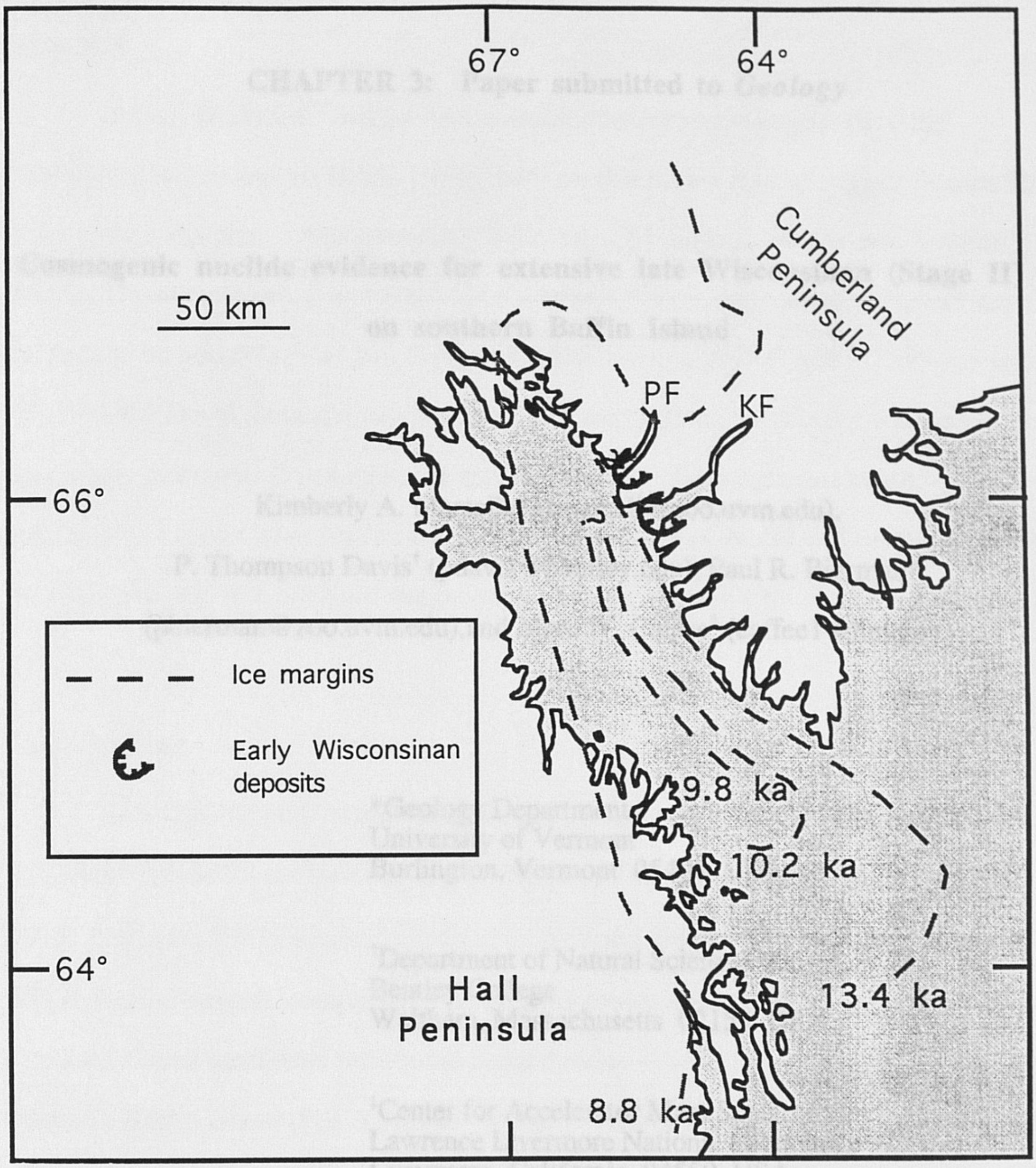

Figure 2.5 Ice reconstruction in Cumberland Sound area during the late Wisconsinan maximum and subsequent deglaciation. Ice margin at 8.8 ka slightly modified from Dyke *et al.* (1982); early Wisconsinan deposits mapped by Dyke (1977, 1979); PF=Pangnirtung Fjord, KF=Kingnait Fjord (adapted from Jennings, 1993).

CHAPTER 3: Paper submitted to *Geology*

Cosmogenic nuclide evidence for extensive late Wisconsinan (Stage II) ice on southern Baffin Island

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Abstract

In-situ produced ^{10}Be and ^{26}Al, measured in over 140 samples from the Pangnirtung Fjord area on Baffin Island, indicate that during marine oxygen isotope Stage II ice extent was greater than previously believed. Cosmogenic nuclide data demonstrate that the Duval moraines, which mark the maximum Wisconsinan ice advance and were thought to be >60,000 years old, were deposited between 25,000 and 9,000 years ago. In addition, the Duval moraines may not mark a distinct relative weathering zone break as previously proposed. Our data imply that climatic and glaciologic conditions were sufficient to support an expanded ice margin in the eastern Canadian arctic during the late Wisconsinan and that full fjord deglaciation occurred rapidly after about 10,000 years ago.

Introduction

Although the vertical and lateral extent of icesheets in the eastern Canadian arctic has been debated for the past century (Coleman, 1920; Flint, 1943; Løken, 1966; Ives, 1978; Andrews, 1987), icesheet configuration during the late Wisconsinan (~24 to 8 kyr BP, or marine oxygen isotope Stage II) remains controversial, primarily because the resultant glacial landforms were never dated directly. In the Pangnirtung Fjord area on southern Baffin Island (Fig. 3.1), 140 ^{10}Be and ^{26}Al analyses now constrain the age of erratic boulders and once-glaciated bedrock outcrops.

The long-standing controversy concerning late Wisconsinan ice extent in the Canadian Arctic is based on different ice reconstructions. One reconstruction, the "maximum ice" model, proposes that an extensive icesheet, with both land- and marine-based domes, existed during the late Wisconsinan (Hughes *et al.*, 1977; Denton and Hughes, 1981; Grosswald, 1984). Other researchers (Miller, 1973; Miller and Dyke, 1974; Andrews, 1975; Boulton, 1979; Dyke and Prest, 1987) support the "minimum ice" model, suggesting that ice was limited to small, land-based icecaps, leaving fjords in the coastal zone of Baffin Island ice-free during the late Wisconsinan.

Weathering zone concept

Much of the chronological debate centers around the interpretation of weathering zones. In arctic fjord regions, three weathering zones (WZ-1, WZ-2, WZ-3) have been defined, increasing in degree of subaerial weathering with altitude (Pheasant, 1971; Pheasant and Andrews, 1973; Boyer and Pheasant, 1974; Locke, 1985). The uppermost zone (WZ-1), primarily high mountain summits and uplands with tors and blockfields, is the most weathered. The lowermost zone (WZ-3), including the valley bottoms and lower fjord walls, is the least weathered (Pheasant and Andrews, 1973). The intermediate zone (WZ-2) is characterized by moderately weathered till, moraines, and bedrock.

The current weathering zone paradigm presumes that this sequence represents a glacial stratigraphic chronology, with the uppermost zone representing the longest exposure (Locke, 1985). The boundaries between weathering zones are usually moraines, assumed to be paleo-ice margins of fjord glaciers, and each lower zone is believed to represent progressively less extensive glaciations during the late Pleistocene (Pheasant, 1971; Boyer and Pheasant, 1974). This three-zone chronology has become a cornerstone for the Quaternary history of the eastern Canadian Arctic, and these weathering zones have been correlated along the coastal fjord region of Baffin Island and Labrador.

Reliable numerical dating of these weathering zones does not exist. Correlations require that weathering zone characteristics are time dependent and are similar in different regions, commonly separated by over 1000 km. An alternative interpretation of weathering zones is based on a glaciological model (Sugden, 1977; Sugden and Watts, 1977). Different glacial regimes could be responsible for the formation of what appear to be distinct, altitudinal weathering zones. Cold-based, non-erosive ice in the uppermost zone may preserve a relict landscape while active, warm-based erosive ice in the lowermost zone sculpts a fresh landscape. An intermediate weathering zone, produced by a transitional basal ice regime (Sugden, 1977; Sugden and Watts, 1977), could separate the uppermost and lowermost zones.

Cosmogenic nuclide dating

As a test of the maximum versus minimum ice models and as an evaluation of the weathering zone concept, we used *in situ*-produced ^{10}Be and ^{26}Al to date directly glacial features on southeastern Cumberland Peninsula, Baffin Island (Fig. 3.1). We focused our fieldwork on the Pangnirtung Fjord area, where WZ-2 and WZ-3 are well exhibited and are separated by the extensive Duval moraines. The Pangnirtung Fjord area was originally mapped by Dyke (1977, 1979), who suggested that the Duval moraines marked the boundary between WZ-2 and WZ-3 and proposed that the moraines were between 60,000 and 100,000 years old (Dyke, 1977, 1979; Dyke *et al.*, 1982).

The Duval moraines are long, linear ridge segments extending about 50 km along both sides of Pangnirtung Fjord. The type locality of these moraines (Fig. 3.2) is just south of Mt. Duval, north of the hamlet of Pangnirtung (Fig. 3.1). Bedrock and boulders in the field area are predominantly Precambrian gneisses and quartz monzonites, both quartz-rich enough for cosmogenic nuclide dating using ^{10}Be and ^{26}Al. Our samples included: (1) six boulders from the type Duval moraines, (2) two samples from a molded bedrock outcrop located just west of the abrupt terminus of the type Duval moraine, one of which was from a prominent, striated quartz vein in that outcrop, (3) two boulders from a raised glaciomarine delta stratigraphically related to the Duval moraines, (4) seven boulders from two recessional moraines (R1 and R2) stratigraphically younger than the Duval moraines, (5) seven boulders from Duval equivalent moraines, and (6) eight samples from WZ-2, stratigraphically older than the Duval moraines (Table 3.1). In addition, more than 100 other samples were collected from molded or striated bedrock, such as the tops of roche moutonnées, and from the tops of large boulders in an 80-km traverse from Cumberland Sound to Summit Lake in Pangnirtung Pass (Fig. 3.1). Only the largest and tallest boulders on moraine crests were sampled to avoid sampling boulders that may have rolled or been buried under snow or till, which would result in younger apparent exposure ages. Samples were collected in cross-transects from the valley floor (WZ-3) up to high tor surfaces (WZ-1).

We used standard methods to process samples and reduce data (Kohl and Nishiizumi, 1992; Bierman and Gillespie, 1995). Isotope production, as a function of altitude and latitude, is scaled for nucleon abundance only (Lal, 1991). For consistency with other published ages, we used the production rates of Nishiizumi *et al.* (1989) to calculate model exposure ages, although some recent findings suggest that these production rates may be 10% to 20% too high (Clark *et al.*, 1995; Bierman *et al.*, 1996a). Using lower production rates would systematically increase our model exposure ages by a similar percentage; however, because our samples are from high latitudes (66° N), nuclide production rates are not time dependent. With the exception of samples from WZ-1, there is no indication that erosion or snow cover has significantly lowered nuclide abundances (annual precipitation is less than 300 mm).

Results

Model exposure ages suggest that the "maximum ice" model is more appropriate than the "minimum ice" model for the Pangnirtung Fjord area, because the Duval moraines were deposited during the late Wisconsinan (Stage II), rather than during the early

Wisconsinan (Stage IV). Of the 140 samples analyzed (Fig. 3.3), only those from the highest tors sampled (WZ-1) have early Wisconsinan minimum-limiting model ages. At least 13 pairs of cosmogenic nuclide analyses demonstrate that the Duval moraines are late Wisconsinan features. These ages have a bi-modal distribution, with one peak in the latest Wisconsinan (9 .0 to 14.8 kyr BP; 11.0±2.2, 1σ, n=8), and one in the early late Wisconsinan (18.2 to 39.5 kyr BP; 25.0±7.3, 1σ, n=5; Fig. 3.4A). The two samples from the molded bedrock outcrop fall into the older age group, with average model exposure ages of 23.5 and 25.5 kyr BP (Fig. 3.4A).

Our data indicate that the ice margin fluctuated near the Duval moraines during the late Wisconsinan (9 to 25 kyr BP). The absence of samples with model exposure ages between 14 and 20 kyr BP suggests either that boulders were not deposited, or that we did not sample boulders deposited during this time. During final ice retreat in Pangnirtung Fjord, two small but distinct recessional moraines (R1, R2; Fig. 3.2) were formed. Data from the recessional moraines (R1=9.5 to 14.4 kyr BP, 11.9±2.0, 1σ, n=3; R2=9.0 to 10.4 kyr BP, 9.5±0.5, 1σ, n=4) suggest that within the resolution of the dating method, the stratigraphically younger recessional moraines formed coevally with the youngest occupation of the Duval moraines (Fig. 3.4A-C). Model exposure ages from the raised glaciomarine delta in Kingnait Fjord, for which the sediment source was the Duval moraines (Fig. 3.1; 9.5 to 13.3 kyr BP, 11.4±1.9, 1σ, n=2), indicate that the delta was abandoned at the same time ice pulled back from the Duval moraines (Fig. 3.4D). The similarity of ages on the recessional moraines, the glaciomarine delta, and the younger cluster of Duval moraines indicates that final ice retreat from the late Wisconsinan maximum position occurred just after 10 kyr BP.

In addition, all samples taken from the floor of Pangnirtung Pass (closer to the ice source) suggest deglaciation was very rapid and complete by 7 to 9 kyr BP. There is no evidence for prolonged stillstands or re-advances of the ice. Bedrock and boulders sampled from ~1100 m asl on Mt. Thor (Fig. 3.1) also have latest Wisconsinan ages (12.5±3.0, 1σ, n=3; Fig. 3.3), which provide a minimum thickness for ice flowing out Pangnirtung Pass and down Pangnirtung Fjord. During the late Wisconsinan maximum, the ice surface must have descended at least 500 m over a distance of about 50 km from Mt. Thor to the Duval moraines, indicating a minimum ice slope angle of 0.57°. Recent work (Jennings, 1993; Jennings *et al.*, 1996) suggests that ice from another source to the west advanced to its maximum position beyond Cumberland Sound by ~11 kyr BP and began its retreat into the sound by 10.2 kyr BP. The ice filling Pangnirtung Fjord most likely merged with ice in Cumberland Sound before both ice masses retreated after 10 kyr BP. Extensive ice filling Pangnirtung Fjord and other fjords on Baffin Island during the late Wisconsinan would provide additional source areas for Heinrich-like events in the North Atlantic region.

Implications

Our data suggest that the weathering zone concept must be modified. Samples from the highly weathered WZ-1 have long (>100 kyr) and complex exposure histories (Bierman *et al.*, 1996b) suggestive of burial during exposure by snow, cold-based ice, till, or loess. There is no distinct geomorphic boundary between WZ-1 and WZ-2 in our field area; however, samples collected from WZ-1 and WZ-2 are isotopically distinct. All seven samples collected from WZ-2 have late Wisconsinan model exposure ages (9 to 22 kyr BP, 12.0±4.5, 1σ, n=7; Fig 4E) and show no evidence for burial. These ages are concordant with the distribution of ages on the Duval moraines and demonstrate that the morphologically distinct Duval moraines do not mark a significant chronological boundary.

It appears that the pattern of ice advances and retreat cannot be described adequately by the classical stratigraphic interpretation of weathering zones. Although WZ-1 is indeed the oldest geomorphic surface, boulders and bedrock in WZ-2 and WZ-3 have similar exposure ages despite being separated physically by the Duval moraines. It appears that rather than marking a temporal boundary, the Duval moraines may mark the margin of erosive warm-based fjord ice juxtaposed with cold-based ice or snow fields covering the highlands (Sugden, 1977; Sugden and Watts, 1977; Nesje *et al.*, 1988). Thus, our data suggest that ice in the Pangnirtung Fjord area on Baffin Island was more extensive in the late Wisconsinan than previously believed (by at least 100 km) and that deglaciation from the Duval maximum occurred in at most a few thousand years. Therefore, other existing chronologies, based on relative dating of weathering zones, such as on eastern Baffin Island where researchers suggest that ice was restricted to the fjord heads during the late Wisconsinan (Miller, 1973; Boyer and Pheasant, 1974; Miller *et al.*, 1977; Nelson, 1980; Brigham, 1983; Mode, 1985), must be re-examined.

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Figure 3.1. Location map of Baffin Island (inset) and Pangnirtung Fjord field area on southeastern Cumberland Peninsula, Baffin Island. Water bodies are gray, Duval moraines are black line segments, and the 99-m asl raised glaciomarine delta is black. Large, solid black arrow indicates large meltwater channel that drained Duval moraines to form delta. CP=Cumberland Peninsula; PF=Pangnirtung Fjord; KF=Kingnait Fjord; P= Pangnirtung hamlet; PP=Pangnirtung Pass; SL=Summit Lake; T=Mt. Thor; D=Mt. Duval.

Figure 3.2. Looking east at type locality of the Duval moraines, separating WZ-2 and WZ-3, recessional moraines, R1 and R2, and Mt. Duval (~700 m). Pangnirtung Fjord in lower left. Field of view in foreground is ~1.5 km.

Figure 3.3. Elevation (m) vs. ^{10}Be model exposure age (kyr BP). Ages on the Duval moraines are contained within ellipse. Dashed line represents idealized break between WZ-2 and WZ-3; note that no samples with model ages >40 kyr BP are found below 600 m. Inset plot of ^{26}Al vs. ^{10}Be model exposure ages (kyr BP) for all samples $\leq$ 40 kyr BP indicates isotopic concordance for most samples.

Figure 3.4. Comparison of ^{26}Al and ^{10}Be model exposure ages (kyr BP) for: (A) Duval moraines, including two bedrock samples, (B) first recessional moraine, R1, (C) second recessional moraine, R2, (D) 99-m raised glaciomarine delta, and (E) WZ-2 (above Duval moraines). Error bars represent a production rate uncertainty of 3% (for elevation) propagated along with analytical uncertainties. Bracketed area represents latest deposition on Duval moraines, recessional moraines, and 99-m delta. Dashed lines represent 1:1 ratio of ^{26}Al and ^{10}Be model ages.

Figure 3.1

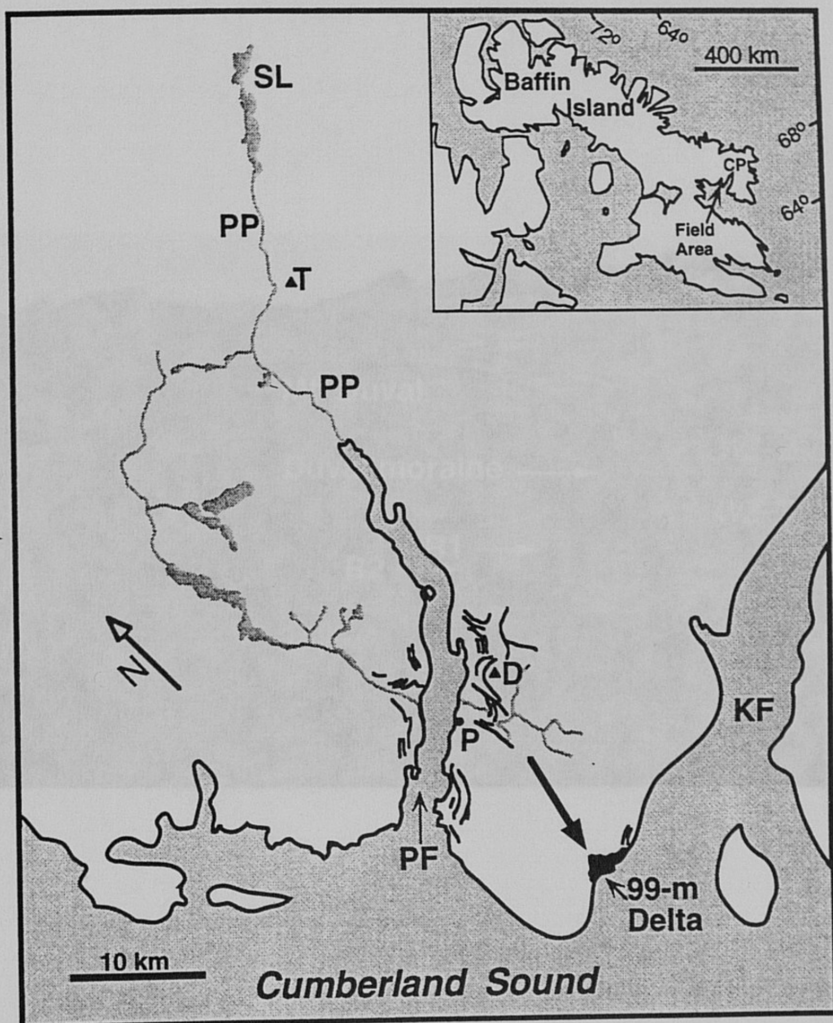

SL
PP
T
PP
N
D
P
KF
PF
99-m
Delta
10 km
Cumberland Sound
72°
64°
400 km
Baffin
Island
68°
CP
64°
Field
Area

Figure 3.2

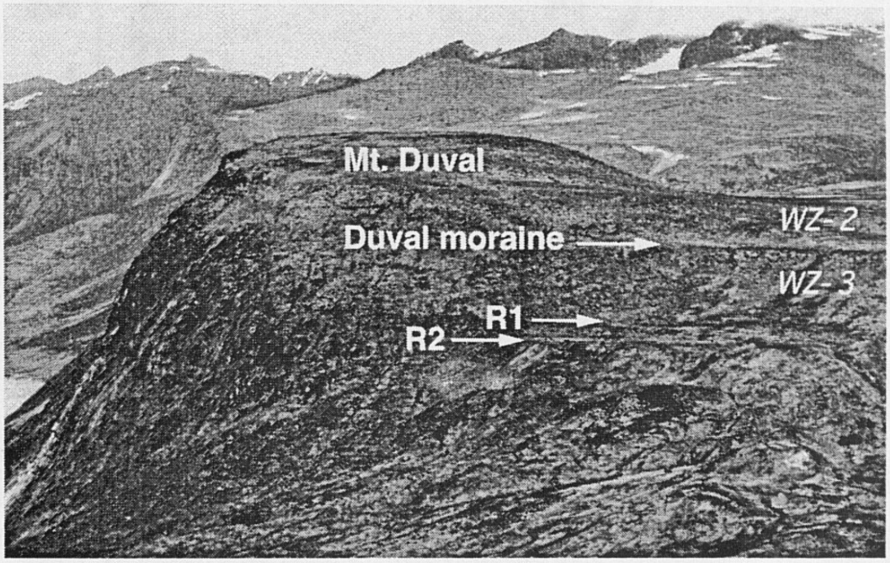


Figure 3.3

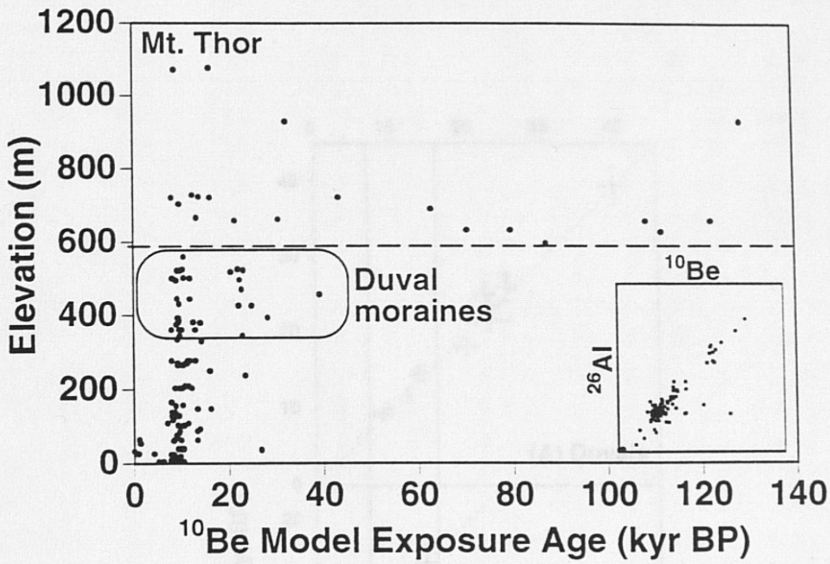
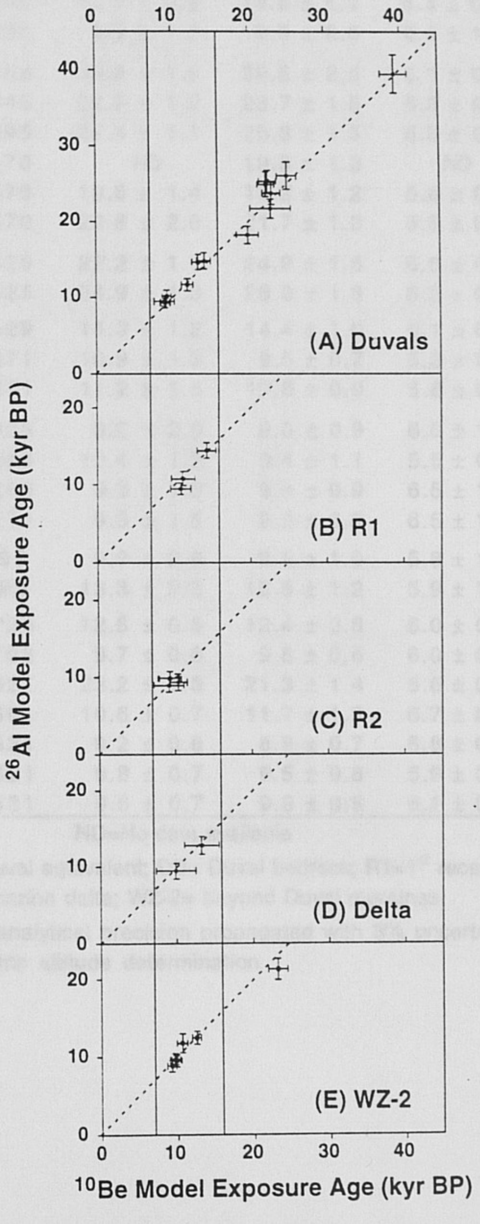

Mt. Thor
Duval moraines
Elevation (m)
1200
1000
800
600
400
200
0
0
20
40
60
80
100
120
140
^{10}Be Model Exposure Age (kyr BP)
^{10}Be
^{26}Al

Figure 3.4



(A) Duvals
(B) R1
(C) R2
(D) Delta
(E) WZ-2
^{26}Al Model Exposure Age (kyr BP)
^{10}Be Model Exposure Age (kyr BP)

Table 3.1. Isotopic data for the Pangnirtung Fjord region (66° N), Baffin Island

Sample	Site†	Elevation (m)	^{10}Be model age (ky)§	^{26}Al model age (ky)§	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-006	TD	500	11.9 ± 0.7	11.7 ± 0.8	6.0 ± 0.5	0.11 ± 0.006	0.68 ± 0.04
KM95-007	TD	493	9.0 ± 0.7	9.3 ± 0.6	6.3 ± 0.6	0.09 ± 0.006	0.54 ± 0.03
KM95-009	DE	665	13.3 ± 0.8	14.6 ± 1.0	6.6 ± 0.5	0.15 ± 0.007	0.99 ± 0.06
KM95-025	DE	445	9.2 ± 0.6	10.2 ± 0.8	6.7 ± 0.6	0.08 ± 0.005	0.55 ± 0.04
KM95-026	DE	345	9.7 ± 0.7	ND	ND	0.08 ± 0.006	ND
KM95-028	DE	375	9.0 ± 0.6	9.4 ± 0.7	6.4 ± 0.6	0.08 ± 0.004	0.49 ± 0.03
KM95-029	DE	380	14.1 ± 0.9	14.8 ± 1.1	6.4 ± 0.6	0.12 ± 0.007	0.78 ± 0.05
KM95-034	TD	495	8.9 ± 1.3	9.5 ± 0.6	6.5 ± 1.0	0.05 ± 0.007	0.35 ± 0.02
KM95-024	DE	455	39.2 ± 1.8	39.5 ± 2.5	6.1 ± 0.4	0.36 ± 0.012	2.18 ± 0.12
KM95-027	DE	345	22.9 ± 1.2	23.7 ± 1.5	6.3 ± 0.4	0.19 ± 0.008	1.21 ± 0.07
KM95-033	TD	495	22.4 ± 1.1	25.3 ± 1.5	6.8 ± 0.4	0.22 ± 0.009	1.47 ± 0.07
KM95-035	TD	470	ND	19.9 ± 1.3	ND	ND	1.13 ± 0.06
KM95-035*	TD	470	19.8 ± 1.4	18.2 ± 1.2	5.6 ± 0.5	0.18 ± 0.012	1.03 ± 0.06
KM95-036	TD	470	22.8 ± 2.6	21.7 ± 1.3	5.8 ± 0.7	0.14 ± 0.015	0.79 ± 0.04
KM95-039	DB	425	22.2 ± 1.1	24.9 ± 1.6	6.8 ± 0.5	0.20 ± 0.008	1.34 ± 0.07
KM95-040	DB	425	24.9 ± 1.3	26.0 ± 1.8	6.3 ± 0.5	0.23 ± 0.010	1.44 ± 0.09
KM95-005	R1	329	14.3 ± 1.2	14.4 ± 1.0	6.1 ± 0.6	0.09 ± 0.007	0.53 ± 0.03
KM95-113	R1	271	10.9 ± 1.3	9.5 ± 0.7	5.3 ± 0.7	0.08 ± 0.009	0.44 ± 0.03
KM95-114	R1	271	11.2 ± 1.5	10.8 ± 0.9	5.8 ± 0.9	0.09 ± 0.012	0.51 ± 0.04
KM95-004	R2	335	9.2 ± 2.0	9.0 ± 0.9	6.0 ± 1.4	0.06 ± 0.012	0.33 ± 0.03
KM95-115	R2	266	10.4 ± 1.2	9.4 ± 1.1	5.5 ± 0.9	0.08 ± 0.009	0.44 ± 0.05
KM95-116	R2	266	9.3 ± 1.3	9.9 ± 0.9	6.5 ± 1.0	0.07 ± 0.010	0.47 ± 0.04
KM95-117	R2	271	9.3 ± 1.5	9.9 ± 0.9	6.5 ± 1.2	0.07 ± 0.011	0.46 ± 0.04
KM95-142	D	99	9.9 ± 2.6	9.5 ± 1.0	5.8 ± 1.6	0.06 ± 0.016	0.37 ± 0.04
KM95-143	D	99	13.3 ± 2.3	12.8 ± 1.2	5.9 ± 1.1	0.09 ± 0.015	0.50 ± 0.04
KM95-010	WZ-2	728	12.5 ± 0.6	12.4 ± 0.8	6.0 ± 0.4	0.15 ± 0.006	0.88 ± 0.05
KM95-011	WZ-2	703	9.7 ± 0.6	9.6 ± 0.6	6.0 ± 0.5	0.11 ± 0.006	0.68 ± 0.04
KM95-101	WZ-2	521	23.2 ± 1.3	21.3 ± 1.4	5.6 ± 0.4	0.23 ± 0.010	1.26 ± 0.07
KM95-102	WZ-2	501	10.6 ± 0.7	11.7 ± 1.2	6.7 ± 0.7	0.10 ± 0.006	0.68 ± 0.06
KM95-103	WZ-2	521	9.2 ± 0.6	8.8 ± 0.7	5.8 ± 0.5	0.09 ± 0.006	0.53 ± 0.04
KM95-104	WZ-2	521	9.8 ± 0.7	9.5 ± 0.8	5.9 ± 0.6	0.10 ± 0.006	0.58 ± 0.05
KM95-105	WZ-2	431	9.6 ± 0.7	9.6 ± 0.8	6.1 ± 0.6	0.09 ± 0.005	0.52 ± 0.04

*Replicate sample ND=No data available

†TD=Type Duval; DE= Duval equivalent; DB= Duval bedrock; R1=1st recessional; R2=2nd recessional; D= 99m-raised glacio-marine delta; WZ-2= beyond Duval moraines

§Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

CHAPTER 4: Paper for submission to *Geological Society of America Bulletin*

Deglacial dynamics and timing, Pangnirtung Fjord and Kolik Valley, Baffin Island, Canada

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Abstract

Paired analyses of *in situ*-produced ^{10}Be and ^{26}Al provide the first direct age estimates for glacial deposits in the Pangnirtung Fjord region, southern Cumberland Peninsula, Baffin Island. The glacial chronology of the area has been revised based on thirty-seven gneissic boulder and six molded gneissic bedrock samples which yield late Wisconsinan exposure ages. Specifically, the prominent Duval moraines, were formed between 25 and 10 ka. The Duval moraines represent a significant ice advance on southern Cumberland Peninsula and have been previously estimated to be early Wisconsinan features (100-60 ka). Because the Duval moraines appear to mark the most extensive ice advance in the area, their age is important for constructing regional glacial chronologies and determining the timing and extent of the last glaciation.

Two recessional moraines, molded and striated bedrock, and glacial erratics on the floor of the Kolik tributary valley indicate that deglaciation occurred rapidly after 10 ka. In addition, boulders from a raised glaciomarine delta stratigraphically related to the Duval morainal position were deposited just prior to the final retreat at 10 ka.

The revised ages of the prominent glacial features in the Pangnirtung Fjord area suggest that an expanded Penny icecap filled the fjord during the late Wisconsinan at the same time as Laurentide ice filled adjacent Cumberland Sound (Jennings *et al.*, 1996).

The revised glacial chronology of the Pangnirtung Fjord area presented here is at odds with the glacial chronologies determined for northern Cumberland Peninsula and the eastern Canadian Arctic. These previous chronologies suggested that regional ice did not extend beyond the head of the fjords during the late Wisconisnan, which has led investigators to propose that the maximum extent of the northeastern margin of the Laurentide icesheet occurred during the early Wisconsinan. A restricted ice margin on Baffin Island during the late Wisconsinan implies that the southern margin of the

Laurentide was out of phase with the northern margin and that the regional climate could not support an expanded ice margin. Our revised chronology of the Pangnirtung region, based on cosmogenic exposure dating, demonstrates that on southeastern Baffin Island, ice extent was not restricted to the fjord heads, that the maximum extent of ice occurred during the late Wisconsinan, and that ice filled the fjord for approximately 15,000 years before rapid retreat at 10 ka. This revised chronology suggests that climatic and glaciologic conditions were quite different during the late Wisconsinan than previously believed.

Introduction

The majority of Arctic Canada was ice covered during much of the Quaternary; however, the vertical and lateral extents of icesheets in the eastern Canadian Arctic have been debated for the past century (Daly, 1902; Coleman, 1920; Odell, 1933; Flint, 1943; Løken, 1966; Ives, 1978; Andrews, 1987). Icesheet configuration during the late Wisconsinan (~24 to 8 ka) remains controversial, primarily because the resultant glacial landforms had never been dated directly, until now. Most of the glacial chronologies in the past were constrained only by relative dating methods and by ^{14}C dating of associated organic material in marine and lacustrine sediments (e.g. Pheasant and Andrews, 1973; Boyer and Pheasant, 1974; Andrews *et al.*, 1975; Birkeland, 1978; Dyke, 1977; Miller *et al.*, 1977; Nelson, 1980)

This study determined exposure ages of glacial features in the eastern Canadian Arctic using *in situ*-produced cosmogenic isotopes, in order to address the long-standing controversy concerning the vertical and lateral extent of the Laurentide icesheet during the Wisconsinan glaciation. The Pangnirtung Fjord area was chosen as the study site for three reasons: (1) the Duval moraines, a major moraine system, are well exposed and appear to separate two weathering zones (WZ-2 and WZ-3), (2) the area lacks adequate numerical dating control, and (3) the predominant lithologies are quartz rich-Precambrian gneisses and quartz monzonites, which provide sufficient material (i.e. quartz) for cosmogenic isotope dating using ^{10}Be and ^{26}Al.

Cosmogenically produced nuclides are quickly becoming a useful tool for constraining landscape erosion rates, as well as exposure ages of landforms. Advances in the fields of analytical chemistry and nuclear physics have allowed quantitative measurement of isotopes including those produced by the interaction of cosmic rays with rock and soil (Elmore and Phillips, 1987). The interaction of cosmic rays with target atoms produces otherwise-rare nuclides in terrestrial material (Davis and Schaeffer, 1955); the ability to measure these isotopes in terrestrial material provides a method for directly dating surface features. Recent studies have shown that the use of *in situ*-produced cosmogenic isotopes provides valuable information for estimating the age of moraines and constraining the timing and extent of alpine and continental glaciation (Phillips *et al.*, 1990, 1996; Brown *et al.*, 1991; Brook *et al.*, 1993, 1996; Gosse *et al.*, 1995a, 1995b) . Cosmogenic exposure dating provides the first method to date directly glacial landforms; in the past the ages for many of these surfaces could only be estimated through relative dating studies and correlation with proxy records, such as those preserved in marine and lacustrine sediments.

We have used measurements of *in situ* cosmogenic ^{10}Be and ^{26}Al to constrain the timing and extent of glaciation on southern Cumberland Peninsula, Baffin Island (Fig. 4.1). In this paper, we discuss the prominent Duval moraines and related features in the Pangnirtung Fjord region. These moraines appear to mark the most significant ice advance on southern Cumberland Peninsula. However, their previously estimated early Wisconsinan age (100-60 ka; Dyke, 1977, 1979) is at odds with terrestrial field evidence suggesting that the area has been more recently covered by ice and recent work (Jennings *et al.*, 1996) that shows evidence for late Wisconsinan till on the floor of adjacent Cumberland Sound.

Our data provide the first direct age estimates of terrestrial glacial features from the northeastern margin of the Laurentide icesheet. These data are significant because Baffin Island has been a major focal point of the debate concerning the vertical and lateral extent of icesheets during late Wisconsinan time. Two different models for the configuration of the northeastern sector of the Laurentide icesheet during the late Wisconsinan (about 24 to 6 ka) have persisted in the literature since the early 1940's (Ives, 1978). One reconstruction proposes that a large, extensive icesheet with both land- and marine-based domes existed during the late Wisconsinan (Hughes *et al.*, 1977; Denton and Hughes, 1981; Grosswald, 1984), constituting what is commonly referred to as the "big ice" or "maximum ice" model (Fig. 4.2, thick solid line). In contrast, many researchers (Andrews, 1975; Dyke and Prest, 1987; Miller, 1973; Miller and Dyke, 1974) believe that ice during the late Wisconsinan glaciation was limited to small, land-based icecaps, leaving the coastal zone of Baffin Island ice-free (Fig. 4.2, thin solid line). This second hypothesis is often referred to as the "small ice" or "minimum ice" model. Baffin Island marks the northeastern margin of the former Laurentide icesheet (Fig. 4.3) and therefore has been the focus of much Quaternary research. The lack of reliable ages for glacial chronologies has contributed to the persistent controversy concerning the late Wisconsinan configuration of the Laurentide icesheet in this region. Numerous researchers have questioned the accuracy of mapped extents of late Wisconsinan ice (e.g. Ives, 1978; Mayewski *et. al.*, 1981) based on the reliability of the chronological methods used for determining these limits.

Field area

Physiography

Cumberland Peninsula, southeastern Baffin Island, is underlain primarily by crystalline rock of the Precambrian Canadian shield, medium to coarse grained Archean gneisses and quartz monzonites (Jackson and Taylor, 1972). Tectonic uplift, following a period of rifting between western Greenland and the eastern coast of Baffin Island, produced the topographic asymmetry of Baffin Island, with the majority of high mountains and deep fjords located along the eastern coast (Dowdeswell and Andrews, 1985). Today the landscape of eastern Baffin Island is dominated by active glaciers including the Barnes and Penny icecaps, as well as thousands of cirque and valley glaciers. Prominent erosional glacial landforms include cirques, horns, aretes, U-shaped and hanging valleys, and deeply incised fjords.

Pangnirtung Fjord, southern Cumberland Peninsula, 3 km wide and 45 km long, opens into Cumberland Sound (Fig. 4.1), which is 80 km wide and 240 km long, with a NW-SE orientation. Cumberland Sound opens into Davis Strait, which is oriented roughly N-S and situated between the eastern coast of Baffin Island and the western coast of Greenland. Pangnirtung Fjord is typically free of ice from late-June to mid-October and has a tidal range of 8-14 m. The fjord walls rise steeply to the mountains, except at the mouths of large river valleys (e.g. the Kolik valley; Fig. 4.4a). Summit heights increase inland, towards Pangnirtung Pass. Pangnirtung Pass, a classic, U-shaped glacial valley occupied by the braided Weasel River, is the head of the fjord (Fig. 4.4b).

The Pangnirtung Fjord region could have been overrun by ice originating from three different ice sources: the northeastern margin of the Laurentide icesheet, radial flow from the Penny Ice Cap, and local ice fed by cirque glaciers. The source of ice in Cumberland Sound during the late Wisconsinan is most likely expansion of the Amadjuak and Hall ice domes overlying the Foxe Basin (Fig. 4.5; Jennings, 1993). Ice filling Pangnirtung Fjord, Kingnait Fjord, and the Kolik valley probably came from an expanded Penny icecap (Penny ice; Fig. 4.5). Lateral moraines, which slope down-fjord, are well exposed along the sidewalls of Pangnirtung Fjord, but no terminal moraines were recognized in the field area. No moraines were observed at the head of Pangnirtung Fjord;

however, such moraines are recognized and mapped as the Cockburn moraines in many other fjord heads along the eastern coast (Andrews and Ives, 1978). These moraines have been dated to between 8 and 9 ka using ^{14}C dating and amino acid racemization analyses on shell deposits, and have previously been interpreted as representing the maximum extent of late Wisconsinan ice on eastern Baffin Island (e.g. Miller and Dyke, 1974; Andrews, 1975; Dyke, 1977, 1979; Andrews and Ives, 1978; Fig. 4.5).

Climate

Cumberland Peninsula has a cold maritime climate. The summer (June-September) mean temperature for Pangnirtung is 5.2 °C; the winter (October-May) mean temperature is -16 °C. Gilbert and Church (1983) report a mean January and February air temperature at Pangnirtung of -27 °C and precipitation through the winter months averaging 13 to 40 mm H_2O per month. Mean annual precipitation is about 400 mm, of which 40-45% falls as rain (Maxwell, 1980). The mean annual temperature and precipitation values are from the observation station at Pangnirtung, 13 m asl. Our field area includes sites up to 1000 m asl, which have higher annual precipitation (500 mm), with up to 80% being snow (Maxwell, 1980). The current glaciation level (average elevation between mountains that maintain an ice mass and those that are ice free) on southern Cumberland Peninsula ranges from 1000 to 1200 m asl (Jacobs *et al.*, 1985).

Previous glacial chronologies

Existing glacial chronologies for the eastern coast of Baffin Island (Fig. 4.6) suggest a model of progressively less extensive ice advances during the last glaciation, with the most extensive glaciation occurring in the early Wisconsinan (ca. 120 to 60 ka). The outermost limit of late Wisconsinan (ca. 8 ka) ice is mapped roughly parallel to the fjord heads of the eastern coast of Baffin Island and well inland from the head of Cumberland Sound. The 8 ka limit is based on ^{14}C ages from glaciomarine deposits related to an extensive moraine system at the head of fjords on the eastern coast (Miller and Dyke, 1974). The development of glacial chronologies in the eastern Canadian Arctic has relied primarily on morphostratigraphic studies of moraines and shoreline features that have been relatively dated using criteria such as: development of felsenmeer and tors (e.g. Boyer and Pheasant, 1974), surface boulder and bedrock weathering stages (e.g. Dyke, 1977), stages of soil development (e.g. Birkeland, 1978), degree of hornblende etching (e.g. Locke, 1979), and depth of soil oxidation (e.g. Nelson, 1980).

The emphasis on developing relative weathering criteria and the recognition of three "weathering zones" along the entire northeastern margin of the Laurentide icesheet from Newfoundland to the northern Queen Elizabeth Islands (Ives, 1978) led to the view of Wisconsinan ice extent that centers on the development of the weathering zone concept (Boyer and Pheasant, 1974). In the coastal mountains, three altitudinally distinct weathering zones have been defined, from highest to lowest, weathering zones 1, 2, and 3 (WZ-1, WZ-2, and WZ-3; Fig. 4.7). The uppermost zone (WZ-1), primarily high mountain summits and uplands with tors and blockfields, is the most weathered. The lowermost zone (WZ-3), including the valley bottoms and lower fjord walls, is the least weathered (Pheasant and Andrews, 1973). The intermediate zone (WZ-2) is characterized by moderately weathered till, moraines, and bedrock.

In general, weathering zones are units of the land surface that can be distinguished based on the amount of subaerial weathering that they have incurred. The modern view of weathering zones focuses on these units as altitudinally separated stratigraphic zones that record different lengths of exposure time over which distinct weathering characteristics have formed (Dyke, 1977). The boundaries between these zones are thought, by most, to reflect the former margins of fjord glaciers (Boyer and Pheasant, 1974) and the upper limit of WZ-3 is commonly believed to represent the maximum extent of ice during the Wisconsinan glaciation (Ives, 1978). Moraine systems within WZ-3 on eastern Baffin Island have been divided further; early Wisconsinan moraines (EWM), mid Wisconsinan moraines (MWM), and late Wisconsinan moraines (LWM; Fig. 4.7).

In the eastern Canadian Arctic, weathering zone studies have focused on relatively small areas; however, researchers have presumed that weathering zones can be correlated over long distances. In correlating these units over distances of thousands of kilometers, researchers must make two assumptions: (1) in a particular area, each weathering zone represents subaerial weathering over a specific period of time, and (2) similar degrees of weathering seen in different regions represent weathering over a similar period of time. Such assumptions presume that the effects of variations in lithology, microclimate, or distance from ice source, are negligible.

An alternative hypothesis to the inference that altitudinal zones of increased weathering reflect increasing exposure time is based on a glaciological, rather than stratigraphic, interpretation of observed weathering zones (Sugden, 1977; Sugden and Watts, 1977). Sugden and Watts suggest that differing glacial thermal regimes could be responsible for the observed weathering zones. If cold-based ice persisted on mountain tops while warm-based ice flowed through the fjords and valleys, the morphology of the lowermost weathering zone (WZ-3) would have been shaped by active ice, accounting for the relatively fresh appearance of rocks in this zone. Cold-based ice in the altitudinally highest zone (WZ-1) would have altered the landscape minimally, preserving the pre-last glaciation landscape. The intermediate zone (WZ-2) represents a zone of transitional basal ice conditions, described by Sugden (1977) as a "warm-freezing" zone, where ice is at the pressure melting point but there is also some regelation. Sugden and Watts (1977) suggest that as ice passes from a warm-based regime (as in WZ-3) to a cold-based regime (as in WZ-1) erratics may be incorporated into the cold-based ice.

Interpretations of weathering zones as stratigraphic units led to the model of a restricted late Wisconsinan configuration of the Laurentide icesheet, the "small ice" model (Fig. 4.2, thin solid line). In contrast, interpretations of the weathering zones as glaciologic units, which have been subaerially exposed for similar periods of time, leads to a model in which the Laurentide icesheet is expanded during the late Wisconsinan, the "big ice" model (Fig. 4.2, thick solid line). In both cases, individual observer biases and opinions seem to influence the interpretation of field data (Ives, 1978). Observer bias is difficult to avoid in relative weathering studies, as many of the criteria are qualitative. In addition, quantifying rates of weathering is difficult. For instance, questions such as how long does it take for weathering pits or tors to form, and what factors control the development of these features remain unanswered.

Previous studies in Pangnirtung Fjord

The Pangnirtung Fjord area on southern Cumberland Peninsula was originally mapped by Dyke (1977) as part of an expansive study of Cumberland Peninsula. Using aerial photographs, field relationships, and relative weathering criteria to investigate the relationship of moraines and other glacial features, Dyke proposed that during the late Wisconsinan, ice extent was restricted to the fjord heads and inland. Dyke proposed that the prominent Duval moraines, which appear to mark the most extensive ice position in the field area, formed during the early Wisconsinan; however, no direct measure of the age of the moraines or glacial features was possible at this time.

Dyke's (1977, 1979) study on Cumberland Peninsula is one of the most comprehensive relative weathering studies conducted. Based on relative age dating, Dyke (1977,1979) proposed that the Duval moraines altitudinally separate WZ-3 from WZ-2. Dyke asserted that the Duval moraines date between 100 and 60 ka based on amino acid data from a single disarticulated mollusk fragment on a nearby, stratigraphically related glaciomarine delta. The Ranger moraines, located at some fjord heads, are believed to delimit the maximum extent of late Wisconsinan (ca. 8000 B.P.) ice in this area, however no moraines have been identified at the head of Pangnirtung Fjord.

Based on relative weathering breaks observed in the field, Dyke (1979) defined three stadial periods, which is consistent with chronologies of Andrews and Miller (1972) and Miller *et al.* (1977) for the east coast of Baffin Island following the last interglacial. The Duval Stade represents the most extensive ice advance, which culminated in the formation of the Duval moraines. Local ice filled Pangirtung and Kingnait Fjords, while Laurentide ice was believed to have filled the inner third of Cumberland Sound; relative sea level was about 100 m above present during the Duval Stade. The Duval Stade is correlated with the Alikdjuak Stade on northern Cumberland Peninsula, which is dated around 100 to 160 ka. During the Outer Usualuk Stade, ice retreat and coastal emergence had raised relative sea level to 108-120 m asl and moraines of this stade were believed to have formed about 40 ka. During the Ranger Stade, Laurentide ice at the head of Cumberland Sound is believed to have coalesced with the western lobe of Penny ice; deglaciation from this position began about 8.7 ka based on correlation with the Cockburn/Baffinland Stade on eastern coast of Baffin Island. Dyke's (1977, 1979) glacial chronology is correlated with the glacial chronologies proposed for the northern side of Cumberland Peninsula, where better exposures of shell-rich glaciomarine sections allows abundant ^{14}C dating and amino acid analyses (Miller, 1973; Boyer and Pheasant, 1974; Nelson, 1980; Brigham, 1983), as well as on eastern Baffin Island (Fig. 4.4, Southern Cumberland Peninsula).

Since the work of Dyke (1977, 1979), some researchers (e.g. Lemmen *et al.*, 1988; Davis *et al.*, 1992; Jennings, 1993) have hypothesized that late Wisconsinan ice may have been more extensive in the Pangnirtung Fjord area than previously mapped and that the

Duval moraines may be late Wisconsinan features. The cosmogenic nuclide data presented in this paper, including 11 measurements from Duval moraines, support the hypothesis that late Wisconsinan ice was more extensive than originally mapped by Dyke (1977). Using the late Wisconsinan (25.5 to 9.2 ka) ages we have calculated for the Duval moraines and the ice margin reconstructions for Cumberland Sound (Jennings, 1993; Jennings *et al.* 1996), we present a revised glacial chronology for the Pangnirtung Fjord area (Fig. 4.6).

Methods

Field Methods

Most samples were collected from the Pangnirtung Fjord area during the summer of 1995. Using sledge hammers and chisels, we collected samples from glacially molded or striated bedrock and from the tops of large boulders. In total, 139 samples were collected during the 1995 field season, of which 56 are from bedrock and 83 are from boulders. All samples are of gneiss, quartz monzonite, or quartz veins, many of which were extremely resistant and difficult to sample. Each sample site was located using topographic maps (1:50,000 and 1:250,000 scales) and air photos (1:15,000 and 1:60,000 scales). A portable GPS was used to obtain latitude and longitude data (± 30 m) and less precise elevation data (± 90 m) at each site. Altimeters were used in conjunction with barometric data loggers to obtain more precise elevation data (± 5 to 10 m). The elevation data were adjusted by comparing the logged barometric data with barometric data from the Pangnirtung Airport, which resulted in corrected elevations accurate to ± 30 meters, or a 3% uncertainty when calculating model exposure ages. Geomorphic descriptions of each sample site included the dip, height, length, width, thickness, exposure geometry, and maximum relief of the sampled surface.

Samples were chosen on the basis of three criteria: (1) the likelihood that the site was covered by neither till nor snow since deglaciation, (2) the likelihood that the sample had not moved since deglaciation, and (3) the likelihood that there had been little or no erosion of the surface since deglaciation. In most of the field area, till cover is very thin (< 20 cm) and the maximum annual precipitation is < 400 mm, minimizing potential for till and snow cover to affect the calculated model exposure ages. We sampled only the tallest and largest boulders, usually from moraine crests, and far from any slopes to ensure that the samples had neither rolled nor moved into their current orientations. The sampled surfaces varied from unweathered striated bedrock, to weathered boulders with maximum crystal relief of 3-4 cm. Samples were collected from tors on the weathered uplands in order to estimate the rate of long-term erosion. Such samples were significantly more weathered than any collected from lower along the fjords and valley floors, and thus calculated model ages are interpreted as minimum limits only.

Laboratory Methods

Samples were prepared at the University of Vermont following standard methods (Kohl and Nishiizumi, 1992) with some modifications. Samples were crushed, ground, and sieved, retaining the 250-710 μm fraction. Samples were then sonicated at 65° C for 12 hours in a solution of 4% HF-4% HNO_3 to remove clays and meteoric ^{10}Be, rinsed, and then sonicated for 12 hours in a solution of 6N HCl to remove Fe.

Initial acid etching improves the efficiency of density separation for which we used acetylene tetrabromide, typically isolating > 80% quartz. Quartz mineral separates were sonicated in a 1% HF-1% HNO_3 solution at 65° C for 8 hours. The samples were then rinsed with de-ionized (DI) water, re-sonicated for 14 hours, and again for 24 hours, using fresh acid each time. After the final acid treatment, samples were rinsed well in DI water.

Sonication produced nearly pure quartz containing less than 200 ppm Al, Fe, and Ti . A fraction of each sample was dissolved and analyzed by ICP to determine major cation abundance. Samples containing >200 ppm of Al, Fe, or Ti were subjected to another cycle of etching in $HF-HNO_3$, and HCl. All of the samples were then subjected to one final etching in $HF-HNO_3$ for 12-14 hours.

Samples were processed in batches of seven along with a laboratory processing blank of Be and Al carrier that was subjected to the remainder of the sample process. After weighing (20-40 g) quartz, 200-300 μg of Be carrier was added, and the sample was dissolved in 48% HF. No Al carrier was added to any of the samples.

After the quartz was dissolved, HF was evaporated. Samples were re-dissolved in 20 ml 1.2N HCl, two aliquots of the sample were taken for ICP measurement of Al and Be, and the remainder of the sample was brought to pH 4 using dilute NH_4OH, and centrifuged. At pH 4, Fe and Ti precipitate and Be and Al remain in solution. The supernatant was brought to pH 9, precipitating Be and Al. The precipitate was dissolved in HCl and dried down. The sample was then re-dissolved four times in 0.5 ml $HClO_4$. After final dissolution, the sample was dried and dissolved in 1 ml 1.2 N HCl. Be and Al fractions were isolated using cation exchange. BeOH and $Al(OH)_3$ were precipitated with NH_4OH, dried, heated to produce oxides, mixed with pure Ag, and packed into targets for analysis at the Lawrence Livermore National Laboratory (LLNL) Center for Accelerator Mass Spectrometry (CAMS).

<u>Model exposure age calculation</u>

Measured isotopic ratios were compared with LLNL CAMS laboratory standards and corrected for background ratios measured in laboratory process blanks. The internal LLNL standards compare with standards made by Nishiizumi and Zurich (ETH) at 1%

(Southon, personal communication, 1997). Blank corrections were made by subtracting the measured isotopic ratio for the process blanks associated with each batch. In addition, $^{10}Be/^{9}Be$ ratios are corrected for boron interference. The blank ratios for $^{10}Be/^{9}Be$ ranged from 2.6×10^{-14} to 4.0×10^{-14}, while the $^{26}Al/^{27}Al$ blank ratios ranged from 4.0×10^{-15} to 6.0×10^{-15}. Measured isotopic ratios for samples (after blank correction) ranged from 9.2×10^{-14} to 3.5×10^{-12} for $^{10}Be/^{9}Be$ and 1.9×10^{-14} to 4.1×10^{-12} for $^{26}Al/^{27}Al$. In the younger samples, the blank correction is significant, which increases the uncertainty associated with these samples.

Calculated abundances were corrected for sample thickness, latitude, and altitude after Lal (1991); no corrections were made for sample geometry because none of the samples collected were from surface edges, none were significantly shielded by horizon geometry, and none dipped more than 40° (a 5% correction). Assuming samples have had little erosion and no pre-depositional cosmogenic exposure, model exposure ages were calculated using equation 1, where, N is the measured isotopic abundance after background correction (atm g^{-1}), P is the isotopic production rate (atm g^{-1} y^{-1}), λ is the decay constant for unstable isotopes (y^{-1}), and t represents the time of exposure (y):

$$t = \frac{-\ln\left(1-\frac{N\lambda}{P}\right)}{\lambda} \qquad (1)$$

We used the production rates of Nishiizumi *et al.* (1989) to calculate model exposure ages. However, the production rates of ^{10}Be and ^{26}Al remain uncertain (Clark *et al.*, 1995; Larsen, 1996). If the production rates of Nishiizumi *et al.* (1989) are over-estimated by 10-20%, our model exposure ages would systematically increase by a similar percentage; however, because our samples are from high latitudes (66° N), nuclide production rates are not time dependent (Lal, 1991).

Data and Discussion

Our data suggest that late Wisconsinan ice inundated Pangnirtung Fjord and adjacent Kingnait Fjord until about 10,000 years ago. The existing chronology of the region indicates that the area remained ice free during the late Wisconsinan (Dyke, 1979); therefore, our re-interpretation of the glacial history represents a significant difference in the extent of ice during this time period. Here we present a subset of our exposure age data that focuses on: (1) the type-Duval moraine and the areas stratigraphically above and below the moraine, (2) the 99-m asl glaciomarine delta on the northern side of Kingnait Fjord, and (3) the Kolik tributary valley region, including regions above and below the Duval moraine limit. The model exposure ages we have calculated are compared with earlier work of other researchers in these three main areas.

Each sample was analyzed for both ^{10}Be and ^{26}Al; mean ages were calculated using the average of the Be and Al model exposure ages, unless one of the isotopic measurements was considered unreliable (based on low Be yield or low accelerator beam current). The uncertainty associated with each exposure age represents the propagation of laboratory uncertainties, along with a 3% uncertainty associated with our barometric altitude determinations. We calculated mean exposure ages for each sample because the Be and Al model exposure ages are well correlated (Fig. 4.8; r^2=0.98).

<u>Type-Duval moraines</u>

The Duval moraines represent a significant advance of ice on southern Cumberland Peninsula, so the age of these features has important climatic significance. Although mapped as early Wisconsinan features (Dyke *et al.*, 1982), the Duval moraines were never directly dated. Our dating demonstrates that these features formed during the late Wisconsinan. On the Duval moraine, ^{10}Be and ^{26}Al ages are well correlated (r^2=0.97) for individual boulders. However, the interpretation of ice cover based on the age distribution of our samples is complex.

The Duval moraines and recessional moraines below the Duval moraines are dissected by a large erosion gully that feeds into the Duval River (Fig. 4.9). The type-Duval moraine is single crested on the up-fjord side of the Duval River, but becomes distinctly double crested on the down-fjord side (Fig. 4.9). In the area surrounding the type locality of the Duval moraine, we collected 15 samples for exposure dating (Fig. 4.9; Table 4.1): (1) six samples from boulders on the type-Duval moraine, (2) two samples from molded bedrock southwest of the abrupt terminus of the moraine, (3) three boulder samples from the upper recessional moraine, stratigraphically below the Duval moraine, and (4) four boulder samples from the lower recessional moraine, stratigraphically the lowest moraine preserved in the type-Duval moraine area. In addition, we discuss five boulder samples from three Duval equivalent moraines on both sides of Pangnirtung Fjord (Table 4.2) and ten boulder samples from the broad, upland area beyond (stratigraphically above) the type-Duval moraines (WZ-2, Table 4.3).

Type-Duval moraines--The isotopic data show that the type-Duval moraines were deposited during the late Wisconsinan. There is a bi-modal distribution of boulder ages on the Duval moraines, with one mode in the early late Wisconsinan (21.8±2.2, 1σ, n=3), and the other mode in the latest Wisconsinan (10.1±1.5, 1σ, n=3). The two modes represent distinct sample populations (t-test; 95% confidence). Two of the three younger ages are from boulders sampled on the inner ridge crest where the type-Duval moraine is double-crested (Fig. 4.9). The older samples occur on the outer crest and the farthest down-fjord segment of the type-Duval moraine (Fig. 4.9). Although the cosmogenic ages yield two distinct boulder populations on the moraine, the samples do not physically appear to be different;

all are of the same gneissic lithology, have a similar size range, and have a similar degree of surface weathering (Fig. 4.10).

For individual boulders, the ^{10}Be and ^{26}Al ages are concordant (Table 4.1; r^2=0.97) and a replicate analysis made on sample KM95-35 (Table 4.1) indicates that the measurements are reproducible. Samples KM95-39 and 40, from the glacially molded bedrock adjacent to the type-Duval moraine terminus to the southeast (Fig. 4.9), are from the same outcrop. KM95-39 is from the surface of the weathered gneissic bedrock and KM95-40 is from a polished, striated quartz vein that stands only 1-2 mm in relief on the bedrock surface (Fig. 4.11). The similarity of the model exposure ages of these samples (Table 4.1) suggests that surface erosion of the weathered gneiss is negligible for the late Wisconsinan time period. In addition, the samples from the molded bedrock (KM95-39 and 95-40) cluster in the older age group of boulders on the type-Duval moraine with an average age of 24.5 ka. Thus, the age distribution of boulders on the Duval moraine most likely represents a geomorphic process. We consider three possible alternatives: (1) the age distribution may represent a long-term erosional process of the moraine surface that has removed older boulders, exposing boulders that were previously below the surface (Hallet and Putkonen, 1994), (2) the boulders with lower exposure ages have a history of snow, ice, or till cover, or surface erosion (Brook *et al.*, 1993), or (3) the age distribution represents the time period over which boulders were delivered to the moraine (Gosse *et al.*, 1995). Alternative 1 is rejected because it is based on a model only applies to old (>100 ka) moraines and one would not expect to see the distinct bi-modal distribution in our ages based on this model. Alternative 2 would require that the three younger boulders experienced a different history of snow, ice, or till cover, or surface erosion than the three older boulders. This scenario does not seem feasible, as the boulders are all of the same lithology, size range, and all are located on the moraine crest.

We favor alternative 3, which from our cosmogenic data indicates that ice filled Pangnirtung Fjord to the height of the Duval moraines from about 24 ka (the oldest average age on the moraine) until about 9 ka (the youngest average age on the moraine), implying that the moraine formed over a period of 15,000 years. Because we cannot be sure that we sampled the earliest and most recent boulders deposited on the moraine, the exposure ages and age range represent minimum limits; however, the end of deposition on the Duval moraines is constrained by the oldest recessional moraine and the stratigraphically related glaciomarine delta, which are discussed later. Our data indicate that the ice margin fluctuated near the Duval moraines during the late Wisconsinan, about 24 to 9 ka. The geomorphic distribution of the boulders on the moraine (Fig. 4.9) suggests that the initial advance may have been slightly more extensive than later advances, as represented by outer and inner moraine crests, respectively.

Five boulders sampled from three separate Duval equivalent moraine segments on both sides of Pangnirtung Fjord exhibit a similar age distribution to those on the type-Duval moraine (Table 4.2), as four of the five exposure ages fall between 14.5 and 9.2 ka, with one older outlier at 39.4 ka. It is difficult to compare the ages among these moraines because we sampled no more than two boulders from any one segment; however, all these data support the presence of extensive late Wisconsinan ice filling Pangnirtung Fjord until about 9 ka. Based on the age distribution of boulders on all Duval moraines sampled, 10 of 11 samples are within the range 24 to 9 ka, with only one sample older than 24 ka (KM95-24; 39.4 ka), suggesting that this sample may have some significant isotopic inheritance.

<u>Duval recessional moraines</u>--Although the type-Duval moraine and its equivalents are recognizable and extensive along both sides of Pangnirtung Fjord, major recessional moraines were observed only below the type-Duval moraines. Two distinct, nested

recessional moraines were sampled (Fig. 4.12); the first recessional moraine (R1) is about 600 meters north and 200 meters downslope of the type-Duval moraine. The second recessional moraine (R2) is another 150 meters further north and between 10-30 meters downslope of R1 (Fig. 4.9). These moraines are more distinctive on the down-fjord side of the Duval River and both are double-crested, similar to the type-Duval moraine.

The average exposure ages of both recessional moraines are indistinguishable within one standard deviation from the younger mode of type-Duval moraine ages. The mean age for R2 (9.5±0.5 ka, 1σ, n= 4) is younger than the mean age for R1 (11.9±2.0 ka, 1σ, n=3), consistent with their stratigraphic relationship. However, the mean age of R1 (11.9 ka) is older than the younger mode of the type-Duval moraine (10.1 ka), and all of the ages from R1 and R2 appear to overlap with the youngest ages on the type-Duval and equivalent moraines. None of these three sample populations can be separated at 95% confidence. These data suggest that the recessional moraines were deposited soon after ice retreat from the type-Duval position.

The uncertainty associated with the mean age for both recessional moraines, 10.5±1.8 ka (1σ, n=7), approaches the resolution of both ^{10}Be and ^{26}Al for dating a single deposit, as most of the single-isotope model exposure ages have an analytical uncertainty of 1 to 3 ka during the late Wisconsinan time period. The average age of the four boulder samples from R2 are tightly clustered; in contrast, the three boulder samples from R1 show a wider scatter of average ages, with the age of the youngest boulder on R1 resembling the average exposure age of R2 (Table 4.1). The average exposure age of R1 appears to be an overestimate and it may be that boulders on R1 and R2 retain some prior exposure history or that the younger recessional moraines contain boulders that are intermixed with boulders from earlier glacial events (after Brown *et al.*, 1991). The exposure age chronology for the recessional moraines suggests that they were deposited rapidly after initial retreat from the type-Duval moraine position; similar rapid deposition of recessional moraines has been documented by Gosse *et al.* (1995) for mountain glaciers in the western United States.

<u>Weathered areas beyond the Duval moraine limit</u>--The weathered areas beyond the Duval morainal limit (WZ-2) present a glacial geologic conundrum. Based on the weathering zone concept (Dyke, 1977, 1979), samples from these areas should yield exposure ages greater than those of the Duval moraines. However, with the exception of one boulder from the summit of Mt. Duval, high tors from above the Pangnirtung hamlet (Bierman *et al.*, in review), and the high tors in the Kolik valley region, exposure ages from WZ-2 overlap exposure ages of samples from the Duval moraines.

Sample KM95-3 (689 m), from the summit of Mt. Duval (the topographic high in this area), has a much older exposure age than the samples stratigraphically below it (model ^{10}Be and ^{26}Al ages of 65.8±2.7 and 59.0±3.6 ka; Table 4.3). This sample suggests that the summit of Mt. Duval was either a nunatak, or was covered by cold-based, non-erosive ice during the late Wisconsinan ice advance.

At 21 to 24 ka, the ice appears to have filled Pangnirtung Fjord to the height of the Duval moraines; however, samples KM95-100 through 104 (Fig. 4.9; Fig 4.13), from the broad region beyond the type-Duval moraine limit, yield exposure ages ranging from 22.3 to 9.0 ka (Table 4.3; r^2=0.85), suggesting that they were covered by glacial ice at this time also. Samples KM95-10 (728 m) and KM95-11 (703 m) are higher than KM95-3, however, these two samples are from the broad, upland plateau beyond the Duval equivalent moraines on the northeastern side of Mt. Duval, rather than on an isolated topographic high. The exposure ages of these samples range from 12.5 to 9.7 ka (Table 4.3).

None of the samples from the weathered areas beyond the Duval morainal limit show evidence for long-term burial (i.e. significant discordance between the ^{10}Be and ^{26}Al ages or $^{26}Al/^{10}Be<<6.0$; Table 4.3). All of the exposure ages are concordant with the distribution of ages on the Duval moraines and demonstrate that the morphologically distinct Duval moraines do not mark a significant boundary in nuclide abundances, nor by inference, an important chronological boundary.

The tors and erratics from the highest regions above the upland plateau surface have nuclide abundances indicating that they were most likely covered by cold-based ice or snow fields, or buried by till, during the Wisconsinan (Bierman *et al.*, in review), while the intermediate upland plateau areas have exposure ages with mean ages indistinguishable from the means of exposure ages of the Duval moraines. One interpretation for the distribution of late Wisconsinan age boulders in the weathered zone beyond the Duval moraines is that the Duval moraines actually represent a medial moraine. The weathered zone beyond the type-Duval moraine is located in a topographic swale that could explain the position of a medial moraine at this location. Prior to retreat of ice in Pangnirtung Fjord after 10 ka large amounts of meltwater were necessary to form the glaciomarine delta on the northern side of Kingnait Fjord, which would not be inconsistent with ice sources on both sides of the type-Duval moraines. However, medial moraines are rarely preserved, so this explanation for the similarity of exposure ages across the type-Duval moraine is rejected.

An alternative explanation for the age distribution of boulders in the weathered zone beyond the Duval moraine limit is based on the model of Sugden (1977) and Sugden and Watts (1977), which suggests that altitudinal zones of increased weathering reflect differing glacial thermal regimes. The presence of warm-based ice in fjords and cold-based ice on the highlands between fjords may indicate that the intermediate upland plateau represents a transitional ice zone, similar to situations suggested by Sugden and Watts (1977). Our exposure age data fit this model, as the nuclide abundances from the uppermost zone (WZ-1) suggest the presence of cold-based ice, while the exposure ages for the Duval moraines indicate the presence of warm-based ice in Pangnirtung Fjord. The model of Sugden and

Watts (1977) explains the morphology of the intermediate weathered (WZ-2) zone based on a zone of transitional basal ice conditions, where erratics may be incorporated into the cold-based ice as ice passes from a warm-based regime to a cold-based regime. Our exposure age data support this model, as the erratics are of similar model exposure ages as the boulders deposited on the Duval moraines. If the erratics were incorporated into the ice from the region of warm-based ice (i.e. the Duval moraines), one would expect that these erratics would have nuclide abundances similar to those from the moraine.

It appears that rather than marking a temporal boundary, the Duval moraines may mark the margin of erosive warm-based fjord ice juxtaposed with cold-based ice or snow fields covering the highlands (Sugden, 1977; Sugden and Watts, 1977). Such a scenario is similar to one of four suggested, but then rejected, by Nesje *et al.* (1988) for the areas in Norway adjacent to the Scandinavian icesheet. Our data suggest that the well documented weathered zones in the Canadian Arctic (e.g. Boyer and Pheasant, 1974; Dyke, 1977; Ives, 1978; Nelson, 1980; Locke, 1985) may have resulted from differing glacial regimes, rather than different lengths of exposure time, which has major implications for the way in which these weathering zones have been previously interpreted.

<u>99-m Delta</u>

The glacial chronology for southern Cumberland Peninsula proposed by Dyke (1977, 1979) is dependent on the estimated age of the raised glaciomarine delta located on the northern side of Kingnait Fjord. This large uplifted glaciomarine delta (Fig. 4.14) extends up to 99-m asl on the south side of the foreland between Pangnirtung and Kingnait Fjords. The sediment for the delta originated from the Duval moraines and related outwash surface, as evidenced by a large meltwater channel (8 km long) and a series of smaller meltwater channels that grade proximally to the type-Duval moraines in Pangnirtung Fjord

(Fig. 4.15; Dyke, 1977). The stratigraphic relationship indicates that the delta must have formed as ice stood at the type-Duval moraines. Two lateral moraine segments along Kingnait Fjord slope down to the marine limit on the eastern side of the deposit (Fig. 4.15; Dyke, 1977), indicating that ice filled Kingnait Fjord prior to formation of the delta. Based on exposure ages from islands near the mouth of Kingnait Fjord (Davis *et al.*, in review), these lateral moraine segments must be roughly equivalent in age to the Duval moraines.

An early Wisconsinan age for the 99-m delta is primarily based on reconstructions of isostatic uplift in the Cumberland Sound region. Late Wisconsinan ice margins were believed to be well inland of the head of Cumberland Sound, represented by a marine limit of about 55 m asl at Pangnirtung, radiocarbon dated about 8.8 ka (^{14}C years; Fig. 4.16; Dyke, 1977, 1979). Dyke (1977, 1979) interpreted higher marine limits, such as that represented by the 99-m delta to be early Wisconsinan in age. However, one reworked shell fragment was collected from a gully incised into silts interpreted as bottomset sediments of the delta and analyzed for amino acid racemization (Dyke, 1977). The amino acid analysis of this shell fragment was found to be similar to that of shell fragments from the type locality of the Quajon Interstade on northern Cumberland Peninsula, which had been dated at >29,000 ^{14}C yr BP, and at 59,000 ± 18,000 yr BP by the Uranium-series methods (Andrews *et al.*, 1975). This 99-m delta sample was the only shell material from the Cumberland Sound area that was interpreted to be pre-Holocene in age (Dyke, 1979) and provided the only chronological data from which to estimate the age of the Duval moraines.

We collected six samples associated with the 99-m delta for exposure age dating (Fig. 4.18; Table 4.4). Two samples were collected from two large boulders a few meters beyond the delta surface on the right-lateral side, two samples were collected from polished bedrock beyond the right-lateral side of the delta, and two samples were collected from two large boulders on the surface of the delta (Fig. 4.17). Of these six samples, only the two boulders on the delta directly indicate the age of formation of the delta and, by association, the age of the Duval moraines. A reasonable question is, how did these large boulders get deposited on the surface of the delta? The most likely source of these boulders is ice rafting. The mean exposure age of the four samples from slightly above the delta reflects deglaciation of Kingnait Fjord and provides a constraint for the maximum age of the delta.

The average exposure age for the four samples above the delta (Fig. 4.18), 10.6 ka, suggests that the delta is less than about 11 ka, similar to the average age of the two boulders on the delta, 11.4 ka (Table 4.4). If ice withdrew from the Kingnait Fjord area at about 10.6 ka and ice stood at the Duval position until 10.1 ka, then the delta must have formed between about 10.6 and 10.1 ka. Of the two boulders on the delta surface (KM95-142 and KM95-143), KM95-142 has concordant Be and Al exposure ages (9.9±2.6; 9.5±1.0 ka) consistent with the above interpretation. Sample KM95-143 also has concordant Be and Al exposure ages (13.3±2.3; 12.8±1.2 ka), although these ages are older than other exposure ages that constrain final retreat from the Duval moraines. Perhaps KM95-143 carried with it some prior exposure when it was deposited by ice-rafting onto the surface of the delta.

The age of the delta is well constrained between about 11 and 10 ka. The delta must be younger than the boulder and bedrock samples in Kingnait Fjord and must predate the retreat of ice from the Duval moraines in Pangnirtung Fjord. Conversely, exposure ages from the 99-m asl glaciomarine delta provide additional age constraint for deposition of the Duval moraines during the late Wisconsinan, specifically, less than 11 ka. The relative sea level curve for Pangnirtung Fjord determined by Dyke (1979) was extended up to the 99-m asl delta, indicating that if the delta did indeed form in the late Wisconsinan its age based on the uplift curves would be about 9.3 ka (Fig. 4.16).

Kolik Valley

The Duval moraines and associated glaciomarine delta provide evidence for ice filling Pangnirtung Fjord during the late Wisconsinan. Exposure age data from tributary valleys to Pangnirtung Fjord provide further information for delimiting the timing and extent of glaciation in the field area. It was not clear to Dyke (1977, 1979) whether ice draining out the fjords was accompanied by ice filling the tributary valleys, such as the Kolik valley on the northern side of Pangnirtung Fjord (Fig. 4.19). Our data, however, provide the first direct age estimates for landforms in the Kolik valley and suggest that this tributary valley was also filled with ice until about 11 ka.

The braided Kolik River flows through a U-shaped valley (Fig. 4.4a), which is fed by meltwater of glaciers, including cirque glaciers and outlet glaciers of the Penny icecap. The Kolik valley is oriented roughly north-south, with drainage to the south into Pangnirtung Fjord. In the Kolik valley, we identified an esker system that grades into ice-contact deltas about 4 to 5 km upstream of the river mouth along Pangnirtung Fjord (Fig. 4.19, Fig. 4.20). This esker is a steep-sided, 6 km long, anastamosing, sinuous ridge composed of well-sorted silts, sand, and gravel that is clearly identifiable in the field and on aerial photographs. The deltas were formed into a series of lakes impounded by ice in Pangnirtung Fjord and were fed by the esker as ice retreated out of the Kolik valley to the north.

Two glacially molded and polished bedrock exposures in the valley bottom were sampled about 6 km upstream from the lowermost esker/delta complex, as were two large boulders resting upon the bedrock. In addition to sampling the valley bottom, we collected samples from two different tributary valleys in the upper drainage basin of the Kolik River: the Ukalik Lake valley and the Amarok Lake valley. Both Ukalik Lake and Amarok Lake are beyond (above) the Duval moraine limit as mapped by Dyke (1979); samples collected from tors on a ridge above Amarok Lake represent the highest samples reported in this paper.

The samples collected farthest upstream (closest to the ice source) in the Kolik River valley are from the shores of Tasikutaaq Lake, studied by Lemmen *et al.* (1988), whose work focused on lacustrine sediment records. The Ukalik and Amarok Lake valleys have been studied by Wolfe (1994; 1996 a,b) and Wolfe and Haertling (1996), whose work also focused on lacustrine sediment records.

Tasikutaaq Lake--Lemmen *et al.* (1988) analyzed the Holocene sediment and diatom record from Tasikutaaq Lake, a large glacially fed lake located in the Kolik valley (Fig. 4.19). From numerous sediment cores, they recognized four distinct stages in the depositional record. The first and oldest stage is represented by an inorganic sandy silt unit, interpreted to be ice-proximal sediment from glacial ice either extending into, or near, the south basin of the lake. Laminated silts, which overly the sandy silt unit, are interpreted to have been deposited as ice retreated up the Kolik valley and the environment changed from ice-proximal to ice-distal. During the deposition of most of this unit, the lake remained biologically unproductive, until the upper 2 cm of the unit, where the diatom assemblage reflects species capable of living in extreme arctic environments, typical during ice retreat.

A massive depositional unit, overlying the laminated silts, represents the formation of a large lake in the upper drainage basin as the ice retreated further up valley, providing a sediment filter for sediments input into Tasikutaaq Lake. A bulk organic radiocarbon age of 7580 ± 140 BP from the base of this unit reflects the age of the formation of the upper lake. The final depositional stage recognized by Lemmen *et al.* (1988) consists of a zone of increasingly well defined laminations, reflecting increasing rates of sedimentation related to Neoglaciation (radiocarbon dated as beginning about 3480 ± 70 BP).

Dyke (1977, 1979) and Dyke *et al.* (1982) mapped the late Wisconsinan limit of ice more than 60 km west of Tasikutaaq Lake, and estimated that the prominent moraine in the upper drainage area was mid-Wisconsinan in age, implying that glacial ice did not occupy the Kolik valley during the late Wisconsinan. Lemmen *et al.*(1988) postulated that ice retreated from the upper drainage basin of Tasikutaaq Lake during late Wisconsinan time, and therefore the late Wisconisnan ice limit must be much more extensive in the Kolik valley than proposed by Dyke *et al.* (1982). However, Lemmen *et al.*(1988) did not suggest that ice completely filled the Kolik valley during the late Wisconsinan.

Bedrock samples collected in 1994 adjacent to Tasikutaaq Lake yielded concordant ^{26}Al and ^{10}Be ages of 12.0±2.7 and 12.3±2.9 ka (Marsella *et al.*, 1995), supporting the hypothesis of Lemmen *et al.* (1988) that late Wisconsinan ice extended further down the Kolik valley than proposed by Dyke *et al.* (1982). The samples described next from further downstream in the lower Kolik River valley indicate that ice was even more extensive during the late Wisconsinan than Lemmen *et al.* (1988) suggest.

<u>Lower Kolik River valley</u>--Four samples were collected in the lower Kolik valley, downstream of Tasikutaaq Lake (Fig. 4.19). Two samples were collected from glacially molded bedrock and two samples from large boulders resting upon the bedrock exposures (Fig. 4.21). These samples have a mean age of 11.0 ka (indistinguishable from the delta abandonment in outer Kingnait Fjord and the younger Duval moraine ages in Pangnirtung Fjord), indicating that the Kolik valley was filled with ice during the late Wisconsinan. The esker-delta complex downstream (Fig. 4.19) of these four samples is interpreted to have formed around this time also; however, there were no boulders or bedrock to sample from these features except for one, low-lying boulder on the esker (KM95-30) which yielded non-concordant, but young, ^{10}Be and ^{26}Al model exposure ages, 16.0±3.7 and 9.1±1.3 ka, respectively, suggesting that this sample is unreliable for dating the age of the esker.

Amarok and Ukalik Lake valleys--The upland regions of the Kolik valley are composed of surfaces that are extremely weathered, characterized by bedrock tors surrounded by a grussified surface, yet preserve a record of prior glaciation. The large bedrock tors exhibit weathering features such as micro- and macro-pitting and surface disaggregation. Fresh cobble erratics were found on tors in juxtaposition to the felsenmeer and grus surrounding the landforms. At elevations similar to those where tors are found (up to 1000 m asl), molded bedrock forms covered by perched boulders are also commonly seen, providing further evidence for previous glacial cover. Dyke (1979) and Wolfe (1996a) interpreted the upland areas surrounding these valleys as not being glaciated since the pre-Wisconsinan, but remaining ice-free, as nunataks during the entire Wisconsinan.

The highest samples collected in the Kolik valley area are KM95-16, from a bedrock tor, and KM95-17, a fresh cobble erratic (both 928 m asl) located above the Amarok Lake valley. KM95-16 has minimum limiting ^{10}Be and ^{26}Al ages of 128.7±5.5 and 100.1±5.5 ka, respectively. The significant discordance in the Be and Al model exposure ages and the $^{26}Al/^{10}Be$ measured ratio of 4.7±0.2 (Table 4.5) suggest that this sample has had a complex burial and exposure history (Bierman *et al.*, in review). Similar to the highest areas in Pangnirtung Fjord, this region is suggested to have been covered by cold-based ice and/or extensive snow fields during the Wisconsinan. The age of the cobble provides some insight into the history of the tor. Sugden and Watts (1977) have suggested that for parts of eastern Baffin Island erratics may have been incorporated into cold-based ice as the ice passed from warm-based to a cold-based basal regime. If KM95-17 is a tillstone, the ^{10}Be and ^{26}Al limiting model exposure ages of 32.3±1.4 and 28.6±1.8 ka (Table 4.5), respectively, suggest that cold-based ice covered the tor some time before 30 ka. Thus, this exposure age estimate should also be considered a minimum, as the measured $^{26}Al/^{10}Be$ ratio for KM95-17 is 5.4±0.3, indicating the cobble may have been buried or shielded from cosmic rays during or after exposure.

A single large boulder, KM95-18, sampled from the Amarok Lake valley floor has concordant ^{10}Be and ^{26}Al ages of 13.9±0.7 and 13.2±0.9 ka, respectively. This sample appears to indicate that ice retreated from the Amarok Lake valley before 13.5 ka, which is just prior to ice retreat in the Kolik valley. If this exposure age is representative of the entire valley floor, it would suggest that the local cirque glaciers retreated just prior to the main retreat of Penny and Laurentide Ice through the Kolik valley and Pangnirtung Fjord.

Ukalik Lake, south of Amarok Lake, is a steeply sloping basin 5 km down basin from a small cirque (Fig. 4.19). Work by Wolfe (1994;1996a,b) suggests that both the Amarok and Ukalik Lake basins were formed prior to the last glacial cycle and served as biological refugia during the late Wisconsinan. Four AMS ^{14}C ages on bryophyte remains from sediment cores suggest that mosses were extant on the uplands between 17 and 38 ka. However, radiocarbon dating organic materials in lacustrine environments greater than 20,000 years old is problematic (Abbott and Stafford, 1996).

We collected three samples from the bedrock cirque lip of Ukalik Lake; one boulder, KM95-19, and two bedrock samples, KM95-20 and 21 (Fig. 4.19). Samples KM95-19 and 20 are from the same surface (Fig. 4.22); the boulder is a fresh erratic and the bedrock is striated and polished. KM95-21 is from one of the many stoss-lee bedrock surfaces observed just above lake level. The average ^{10}Be and ^{26}Al age of the boulder, KM95-19 is 22.7 ka, whereas both bedrock surfaces have average ages of 11.5 ka. These cosmogenic ages suggest the presence of ice in the Ukalik Lake basin during the late Wisconsinan (~22-11.5 ka), which would suggest that the radiocarbon ages of Wolfe are too old. The moss ages are from small samples (<50 mg) of bryophytes deposited in silty sediments. There is no local source of carbonate in the field area to suggest that the samples have incorporated "old" carbon; however, there is the potential that the mosses were re-deposited from older sediments (Wolfe, 1996a).

If we accept Wolfe's (1996a) radiocarbon ages, then the Ukalik Lake basin would have formed from glaciation prior to the mid-Wisconsinan (> 38 ka), after which time it may have been covered by till or snow fields. The discordance between our boulder (22 ka) and bedrock (11.5 ka) ages could be interpreted as the result of the bedrock having been covered by snow or till, with the top of the boulder protruding from this cover, thereby experiencing more exposure. However, thick till cover is unlikely on a bedrock lip, and extended thick snow cover even less likely in such an exposed location.

It is difficult to resolve which of the age estimates (direct dating using cosmogenic isotopes or radiocarbon dating the lacustrine sediment record) is more reasonable for the Ukalik Lake basin, based on the limited number of samples in this area from each study. Of the four moss ages reported by Wolfe (1996a), only one is from the Ukalik Lake sediment record, the limiting 38 ka age. The remaining three moss ages are from the Amarok Lake sediment record and range from 17 to 20 ka. If the single AMS age from Ukalik Lake comes from a moss that has been redeposited from older sediment, then the remaining moss ages could suggest that the mosses were extant on the uplands only since 20 ka. The Amarok Lake basin is about 200 m higher than the Ukalik Lake basin and it may be possible that the local cirques responded differently during the same period, leaving the possibility that the Ukalik Lake basin may have been ice covered during the late Wisconsinan, as our cosmogenic age estimates suggest. There are no unequivocal data to demonstrate that the cirque glaciers were expanded during the late Wisconsinan, or that the entire upland region was ice free during the late Wisconsinan.

Conclusions

Numerous cosmogenic isotope measurements suggest that ice retreated from Pangnirtung Fjord about 10,000 years ago and allow us to revise prior estimates of the timing and extent of glaciation in the Pangnirtung Fjord area. The Duval moraines were deposited during a long-lived (~15,000 years) ice advance in the late Wisconsinan. Extensive ice filled Pangnirtung Fjord and advanced to the Duval morainal position at least 22,000 years ago. At about 10 ka, ice retreated from the Duval moraines, and just prior to 10 ka it appears that ice also retreated from the mouth of Kingnait Fjord. A large glaciomarine delta at 99 m asl formed along Kingnait Fjord between 10.9 and 10.1 ka, as meltwater and sediment drained across the foreland from the Duval morainal position. Following the deposition of this delta, the ice appears to have retreated rapidly, producing at least two recessional moraines before retreating from Pangnirtung Fjord. Ice filling the Kolik tributary valley appears to have retreated at about the same time as the fjord ice.

Extensive ice, filling Pangnirtung Fjord, represents an expansion of the Penny Icecap; ice from Pangnirtung Fjord may have merged with expanded Laurentide ice in adjacent Cumberland Sound (Jennings, 1993; Jennings *et al.*, 1996). The observation that none of the Duval moraines slope down towards the fjord bottom suggests that the two ice masses merged. The ice in Cumberland Sound advanced beyond the sound around 11 ka, but had retreated into the sound by about 10.2 ka (Jennings *et al.*, 1996). The ice retreat from the Cumberland Sound area may have initiated the rapid retreat of ice in Pangnirtung and possibly Kingnait Fjord by causing rapid downdraw and ice calving at the fjord mouths.

The glacial chronology we present here does not agree with the previous glacial chronology of the area (Dyke *et al.*, 1982), or with other glacial chronologies on northern Cumberland Peninsula; however, ours is the first chronology that utilizes direct dating of the preserved glacial features. The previous chronologies have been based on reconstructions of isostatic uplift, inference with radiocarbon and amino acid racemization ages from glacial proxy records, and through numerous weathering zone studies. These weathering zone studies have assumed that distinct altitudinal zones of subaerially weathered land surfaces represent former icesheet heights and therefore record progressively less extensive glaciations. Based on these weathering zone studies and sediment proxy records, the extent of the late Wisconsinan ice margin was believed to be restricted to the heads of fjords along the eastern coast of Baffin Island. The revised chronology presented in this paper not only indicates that the late Wisconsinan margin was more extensive than previously believed, but also provides evidence suggesting that observed weathering zones were produced through differing basal glacial regimes, and the morphologically distinct Duval moraines do not mark a significant chronological boundary. Our study demonstrates the utility of cosmogenic isotopes for dating glacial features produced during the late Wisconsinan, provides evidence for an expanded northeastern margin of the Laurentide icesheet, and implies that while warm-based ice actively eroded the fjord and valleys, cold-based ice may have persisted in regions above 700 m asl.

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Figure captions

Figure 4.1 Location map of Baffin Island (inset) and Pangnirtung Fjord field area on southeastern Cumberland Peninsula, Baffin Island. CP=Cumberland Peninsula; PF=Pangnirtung Fjord; KF=Kingnait Fjord; P= Pangnirtung hamlet; D=Mt. Duval; DR= Duval River; KR=Kolik River; TL=Tasikutaaq Lake; AL=Amarok Lake; UL=Ukalik Lake; PP=Pangnirtung Pass; 99-m delta labeled.

Figure 4.2 Overview map of northern North America showing proposed maximum and minimum extents of Laurentide icesheet. The maximum extent (thick solid line) has been proposed as early Wisconsinan and the minimum extent (thin solid line) has been proposed as late Wisconsinan. Cumberland Peninsula, Baffin Island labeled CP (adapted from Vincent and Prest, 1987).

Figure 4.3 Proposed 11 ka reconstruction for the Laurentide icesheet (adapted from Dyke and Prest, 1987).

Figure 4.4 (a) The Kolik River valley and mouth of the braided Kolik River on the northern side of Pangnirtung Fjord. (b) Looking northeast at the head of Pangnirtung Fjord and the south end of Pangnirtung Pass, a classic U-shaped valley.

Figure 4.5 Ice divides for the northeastern margin of the Laurentide icesheet, Baffin Island (adapted from Dyke and Prest, 1987).

Figure 4.6 Glacial time-distance diagrams showing the correlation of tills, moraines, and glacial events on eastern Baffin Island (adapted from Nelson, 1980; Pheasant and Andrews, 1973; Miller *et al.*, 1977; Andrews *et al.*, 1976; Dyke, 1977, 1979).

Figure 4.7 A schematic portrayal of the Quaternary glaciation model for eastern Baffin Island, including Cumberland Peninsula. The upper diagram shows the vertical distribution of lateral moraines and weathering zones on a longitudinal profile. The lower diagram shows the same in planimetric view of highlands dissected by fjords. WZ=weathering zone, EWM=early Wisconsinan moraine, MWM=mid Wisconsinan moraine, LWM=late Wisconsinan moraine (adapted from Dyke, 1977).

Figure 4.8 ^{26}Al vs. ^{10}Be model exposure ages (ka) for samples reported, r^2=0.98 Dashed lines represent 1:1 ratio of ^{26}Al and ^{10}Be model ages.

Figure 4.9 Sample site map for the type-Duval moraine samples and some WZ-2 samples; numbers represent KM95- sample locations, summarized in Table 4.1. DE = Duval moraine equivalents. Gray solid lines = 100 m contours.

Figure 4.10 Two gneissic boulder samples from the type-Duval moraine: (a) KM95-7 from the inner crest of the moraine; average exposure age=9.2 ka; 1.7 m person for scale. (b) KM95-33 from the outer crest of the moraine; average exposure age=23.8; 1.6 m person for scale. Fig. 4.9 shows the spatial relationship of these two samples, which are of similar size, surface weathering, and lithology.

Figure 4.11 Striated quartz vein (KM95-40) in glacially molded bedrock outcrop (KM95-39) southwest of the type-Duval moraine; lowering of the bedrock surface is a few mm; 15-cm long chisel for scale.

Figure 4.12 Type-Duval recessional moraines R1 and R2. Gneissic boulder samples on (a) R1 and (b) R2; data summarized in Table 4.1.

Figure 4.13 Samples KM95-103 and KM95-104 from the weathered upland surface beyond the Duval moraine limit; data summarized in Table 4.3; 1.8 m person for scale.

Figure 4.14 Looking south towards Cumberland Sound across the raised 99-m asl glaciomarine delta along outer Kingnait Fjord. Two 1.8 m people can be seen on the top surface of the delta.

Figure 4.15 Relationship of the Duval moraines to the raised 99-m marine limit delta along outer Kingnait Fjord (adapted from Dyke, 1979).

Figure 4.16 Relative sea level curves for Cumberland Sound: A- head of Clearwater Fjord; B- Middle Clearwater Fjord; C- Shilmilik Bay; D- Kangerk Fjord; E- Usualuk Fjord; F-Pangnirtung Fjord; G- head of Kingnait Fjord; H- Outer Cumberland Sound (from Dyke, 1979). Gray shaded area is an extension of the Pangnirtung curve to the raised 99-m delta; dotted line indicates the intersection of this surface with the time axis at about 9.3 ka.

Figure 4.17 Spatial distribution of sample sites associated with the raised 99-m glaciomarine delta. Diagram is a schematic, planimetric view. The raised 99-m delta is shown on Figures 4.1 and 4.15.

Figure 4.18 Samples KM95-138 to 141 from beyond the raised 99-m delta surface along outer Kingnait Fjord; data summarized in Table 4.4.

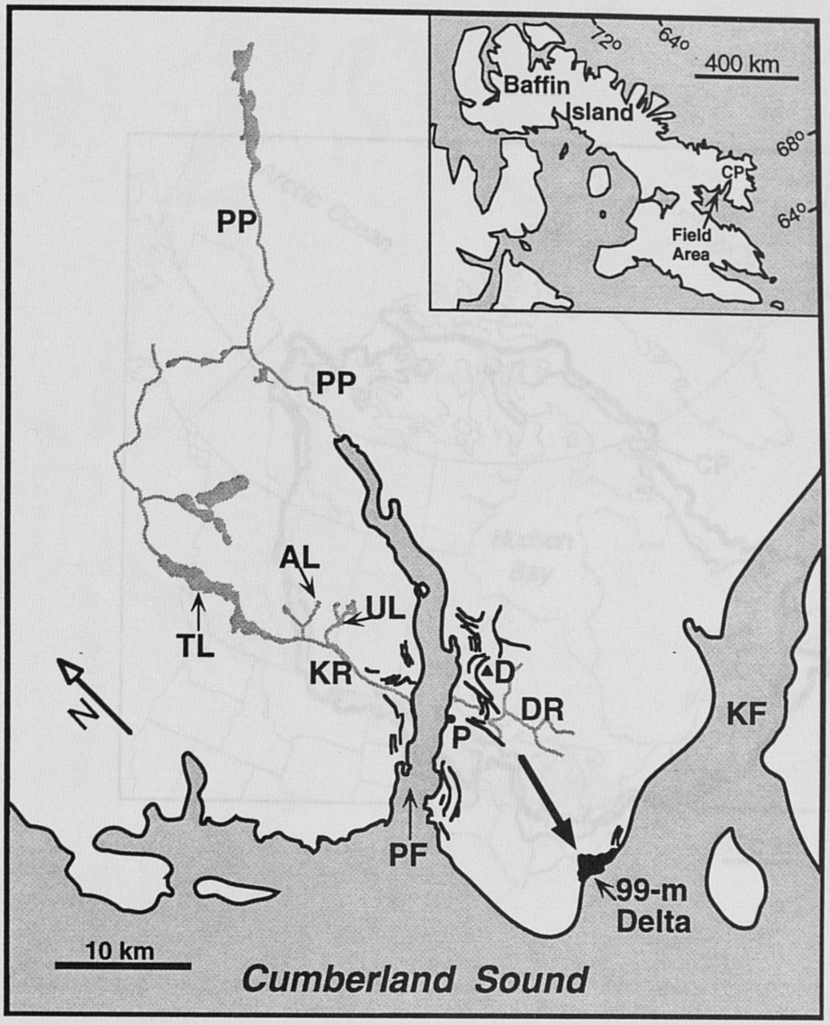
Figure 4.19 Map showing Kolik valley sample sites: *TL* indicates the location of the cosmogenic age reported by Marsella *et al.*, 1995; numbers indicate KM95- sample locations. Gray solid lines = 300 m contour intervals.

Figure 4.20 (a) Looking south, towards Pangnirtung Fjord, down the surface of the esker in the lower Kolik River valley. (b) Looking south at the ice-contact delta.

Figure 4.21 (a) Glacially molded bedrock (KM95-15). (b) Erratic boulders (KM95-12) from the same location in the Kolik River valley; data summarized in Table 4.5.

Figure 4.22 Looking northeast at Ukalik Lake, showing the steep-walled basin and the location of KM95-19 and 20; data summarized in Table 4.5; 1.8 m person for scale.

Figure 4.1



Baffin
Island
72°
64°
400 km
68°
64°
CP
Field
Area
PP
PP
AL
UL
TL
KR
D
DR
P
PF
KF
99-m
Delta
10 km
Cumberland Sound
N

Figure 4.2

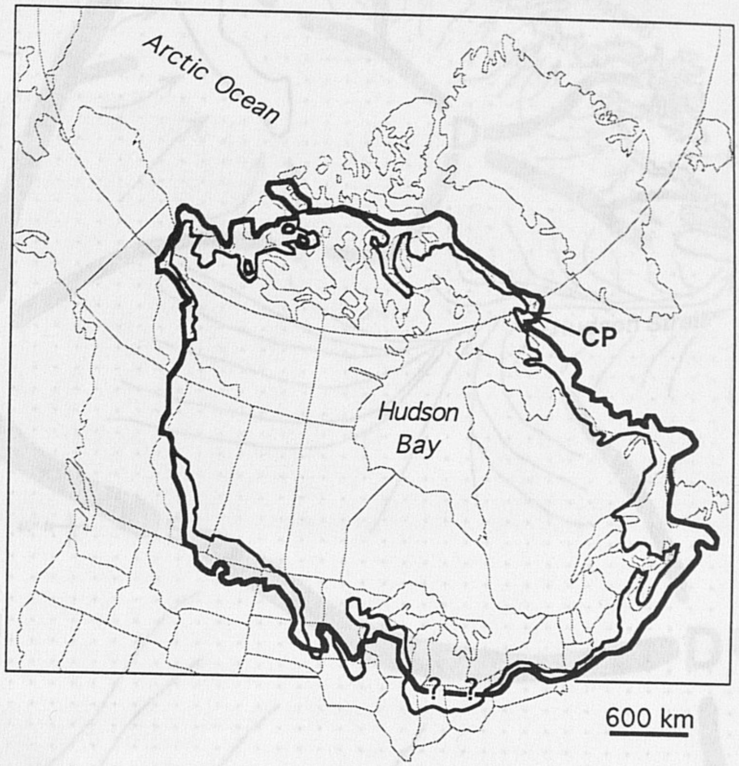
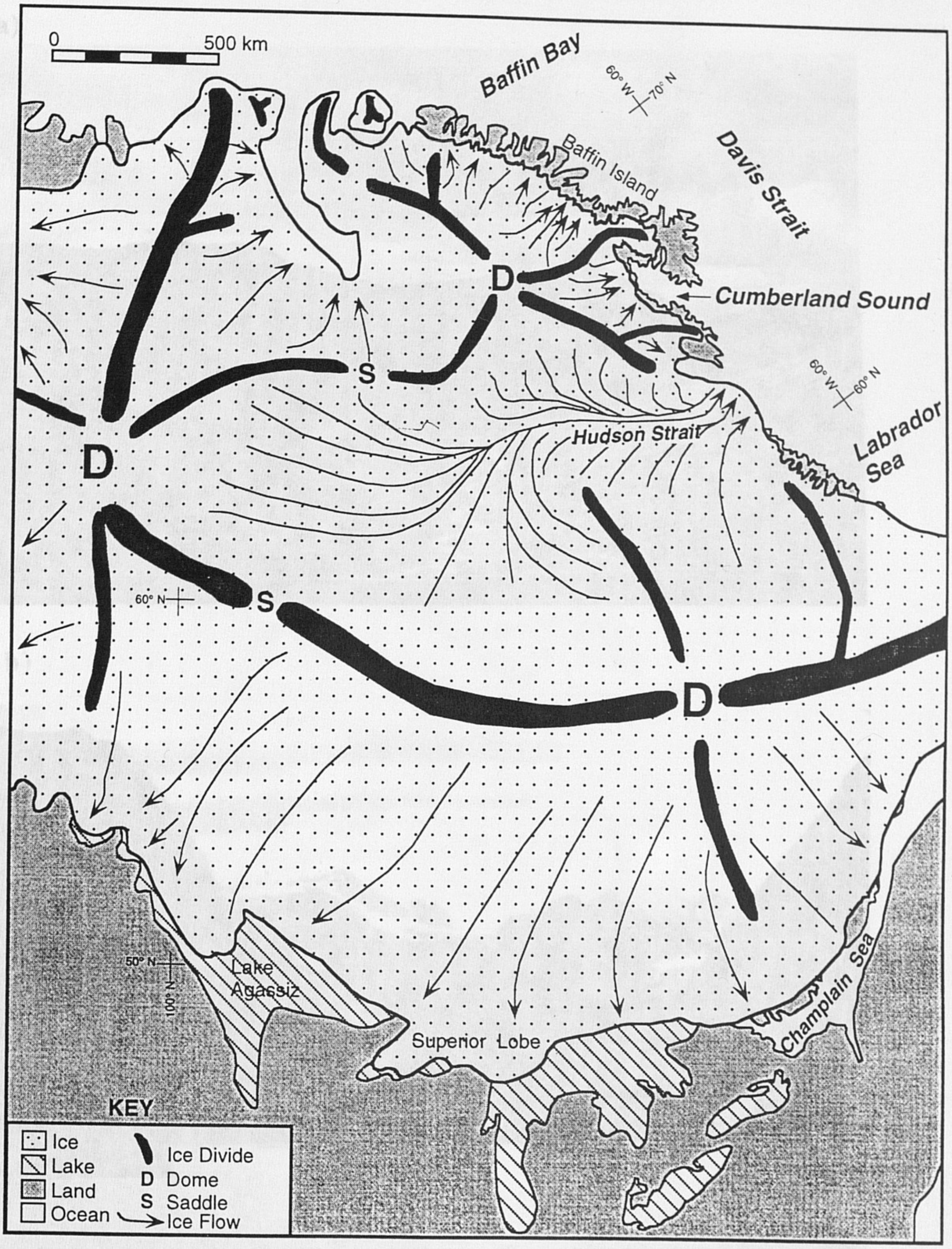


Figure 4.3



0
500 km
Baffin Bay
60° W
70° N
Baffin Island
Davis Strait
D
Cumberland Sound
S
60° W
60° N
Hudson Strait
Labrador Sea
D
60° N
S
D
50° N
100° N
Lake Agassiz
Superior Lobe
Champlain Sea
KEY
Ice
Lake
Land
Ocean
Ice Divide
D Dome
S Saddle
Ice Flow

Figure 4.4

a)

b)

Figure 4.5

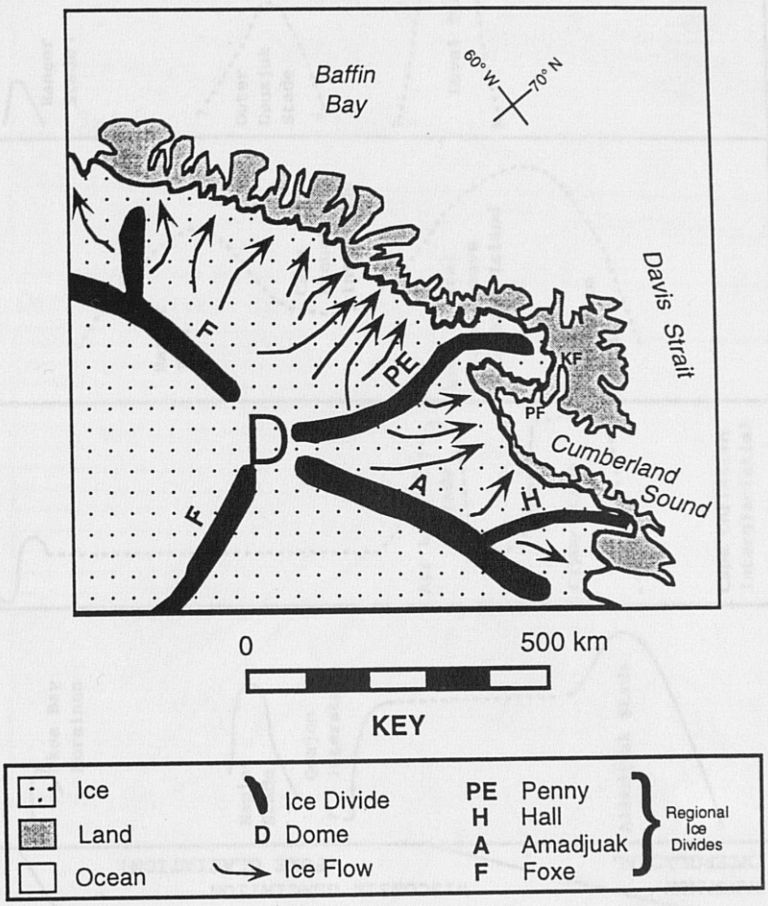
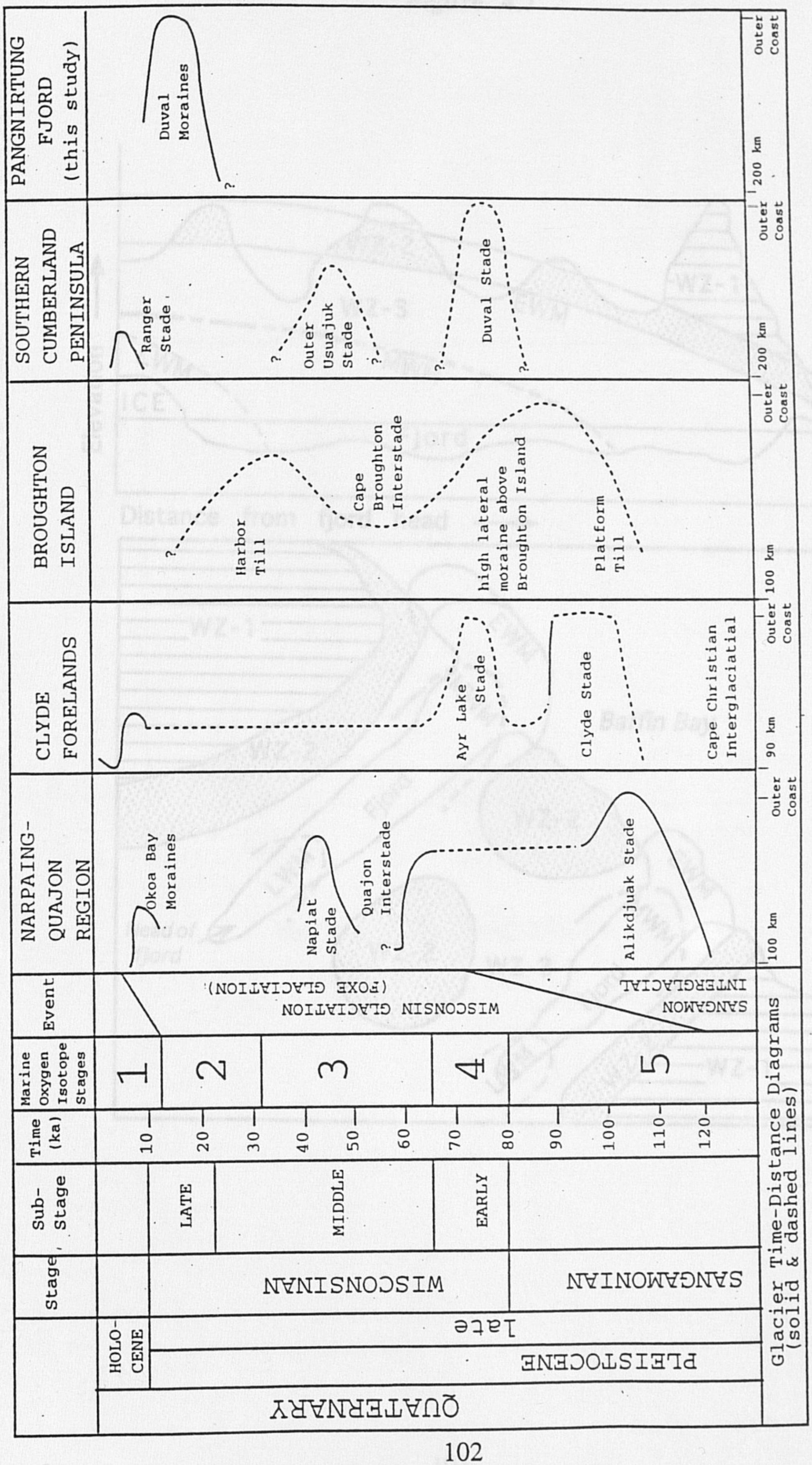

Baffin
Bay
60° W
70° N
Davis Strait
Cumberland
Sound
KF
PF
F
PE
D
A
H
F
0
500 km
KEY
Ice
Land
Ocean
Ice Divide
D Dome
Ice Flow
PE Penny
H Hall
A Amadjuak
F Foxe
Regional Ice Divides

Figure 4.6



QUATERNARY
HOLO-CENE
PLEISTOCENE
late
Stage
WISCONSINAN
SANGAMONIAN
Sub-Stage
LATE
MIDDLE
EARLY
Time (ka)
10
20
30
40
50
60
70
80
90
100
110
120
Marine Oxygen Isotope Stages
1
2
3
4
5
Event
WISCONSIN GLACIATION (FOXE GLACIATION)
SANGAMON INTERGLACIAL
NARPAING-QUAJON REGION
Okoa Bay Moraines
Napiat Stade
Quajon Interstade
Alikdjuak Stade
100 km
Outer Coast
CLYDE FORELANDS
Ayr Lake Stade
Clyde Stade
Cape Christian Interglaciatial
90 km
Outer Coast
BROUGHTON ISLAND
Harbor Till
Cape Broughton Interstade
high lateral moraine above Broughton Island
Platform Till
100 km
Outer Coast
SOUTHERN CUMBERLAND PENINSULA
Ranger Stade
Outer Usuajuk Stade
Duval Stade
200 km
Outer Coast
PANGNIRTUNG FJORD (this study)
Duval Moraines
200 km
Outer Coast
Glacier Time-Distance Diagrams (solid & dashed lines)

Figure 4.7

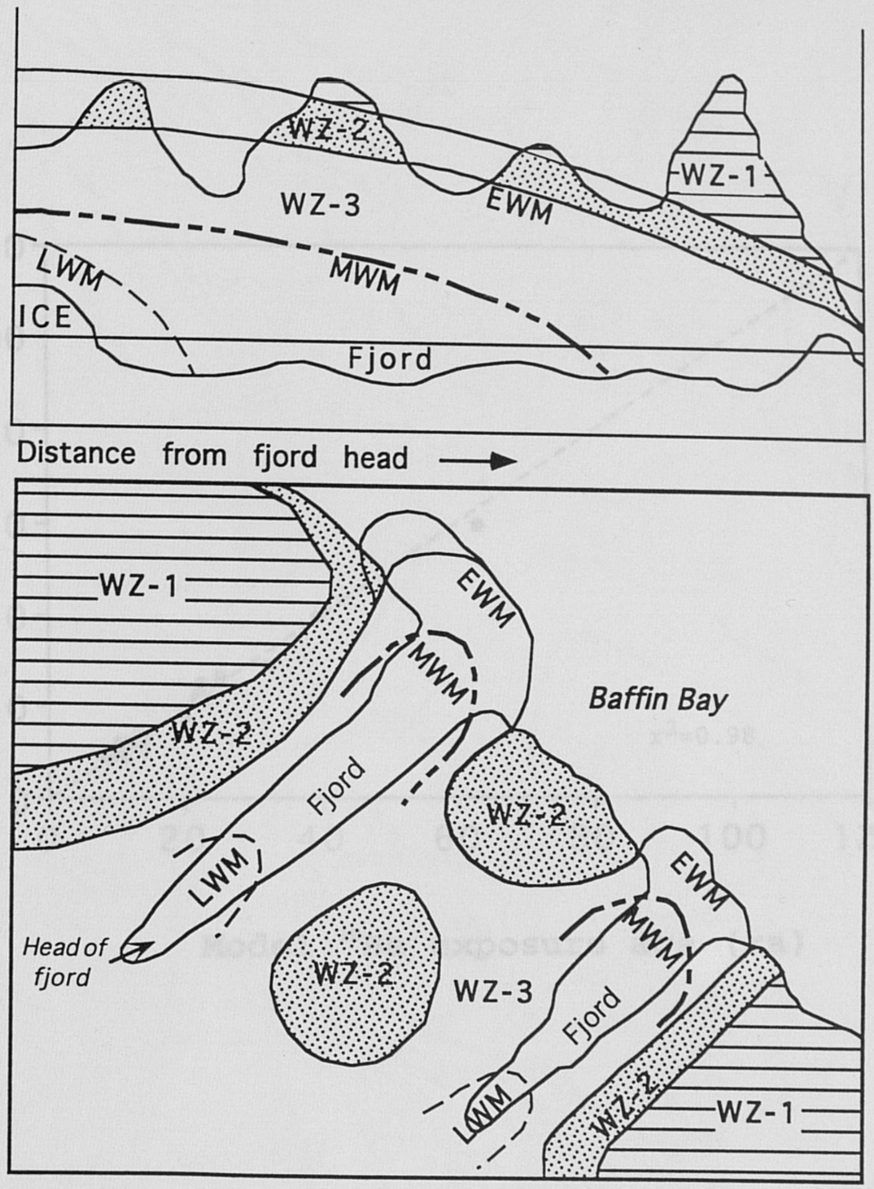

WZ-2
WZ-1
WZ-3
EWM
Elevation
LWM
MWM
ICE
Fjord
Distance from fjord head
WZ-1
EWM
MWM
Baffin Bay
WZ-2
Fjord
WZ-2
LWM
EWM
MWM
Head of fjord
WZ-2
WZ-3
Fjord
LWM
WZ-2
WZ-1

Figure 4.8

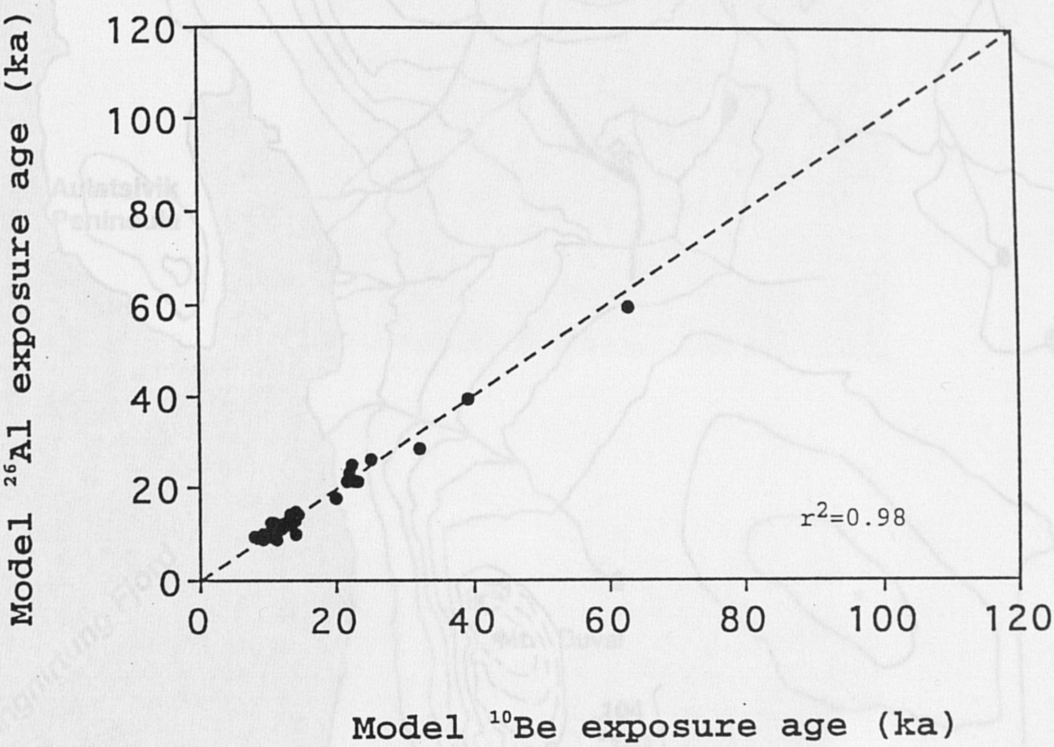
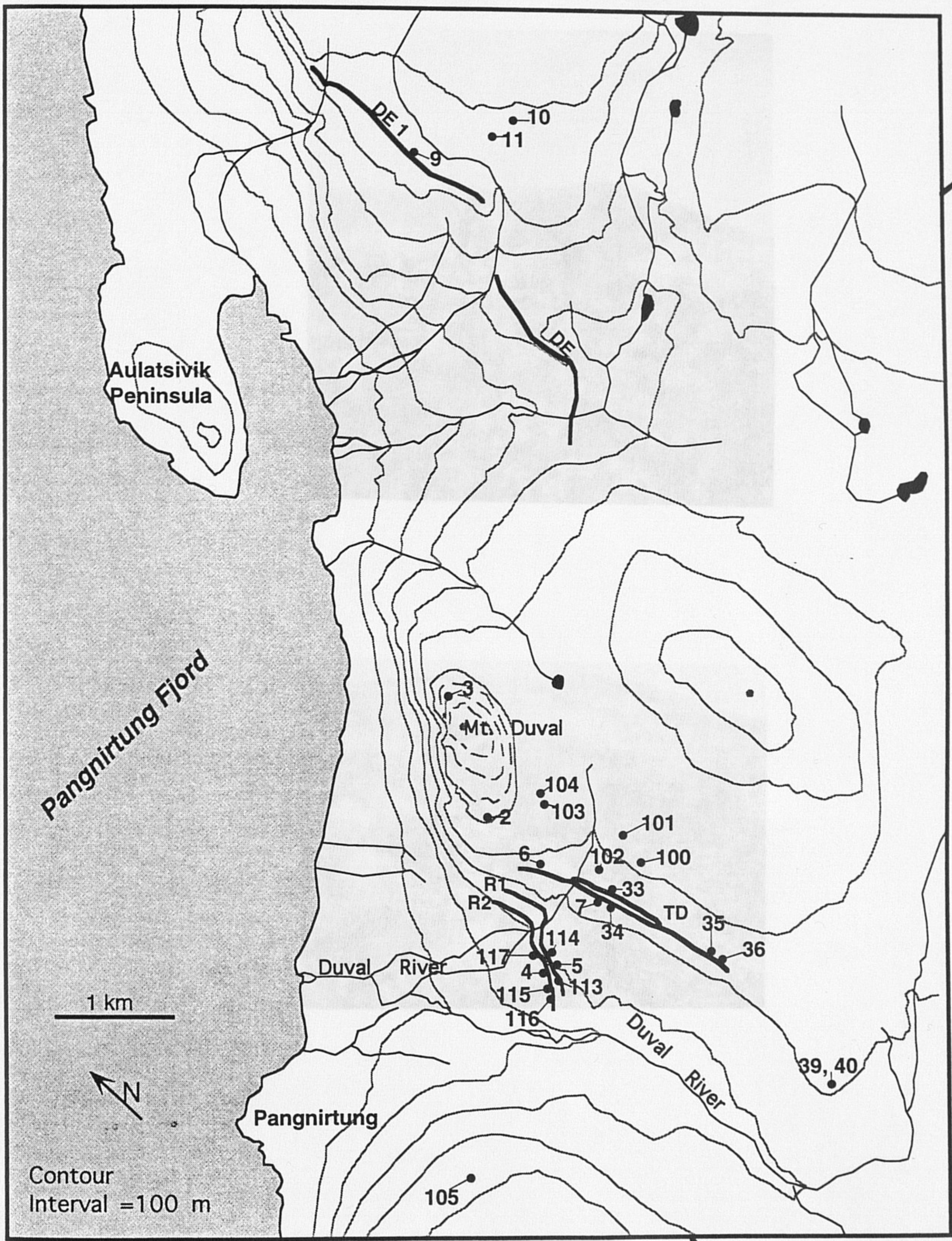

Model ^{26}Al exposure age (ka)
120
100
80
60
40
20
0
0
20
40
60
80
100
120
r^2=0.98
Model ^{10}Be exposure age (ka)

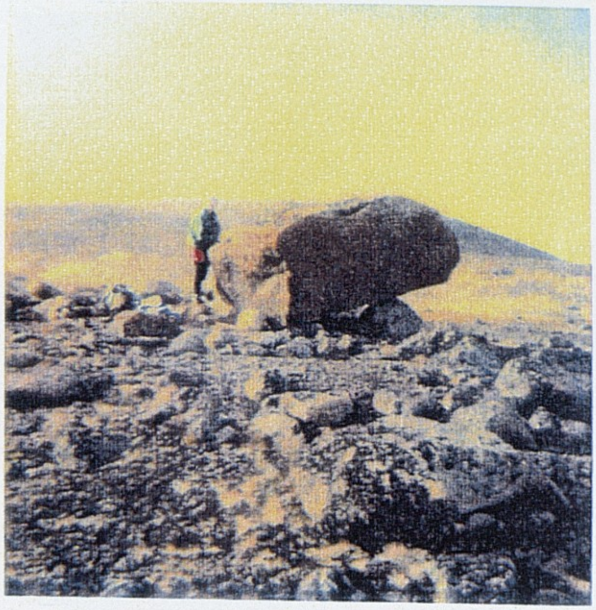
Figure 4.9



DE 1
9
10
11
DE
66° 10' N
Aulatsivik
Peninsula
Pangnirtung Fjord
3
Mt. Duval
104
103
2
101
102
100
6
R1
R2
33
7
34
TD
35
36
114
117
5
4
113
115
116
Duval River
Duval River
1 km
N
39, 40
Pangnirtung
Contour
Interval =100 m
105
65° 40' W

Figure 4.10

a)



b)

Figure 4.11

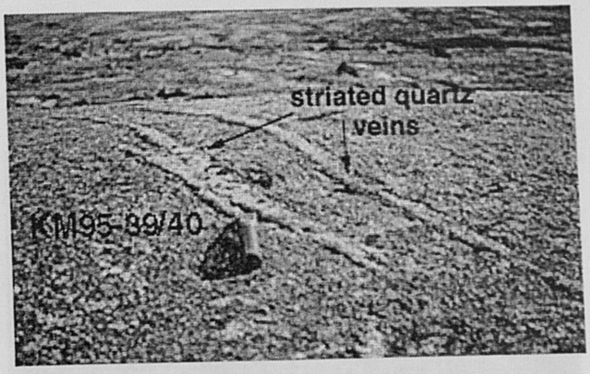
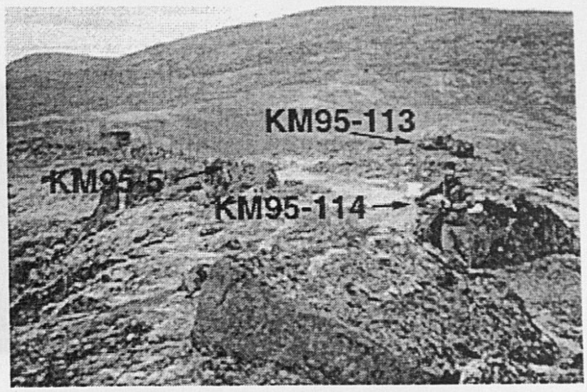


Figure 4.12

a)



b)

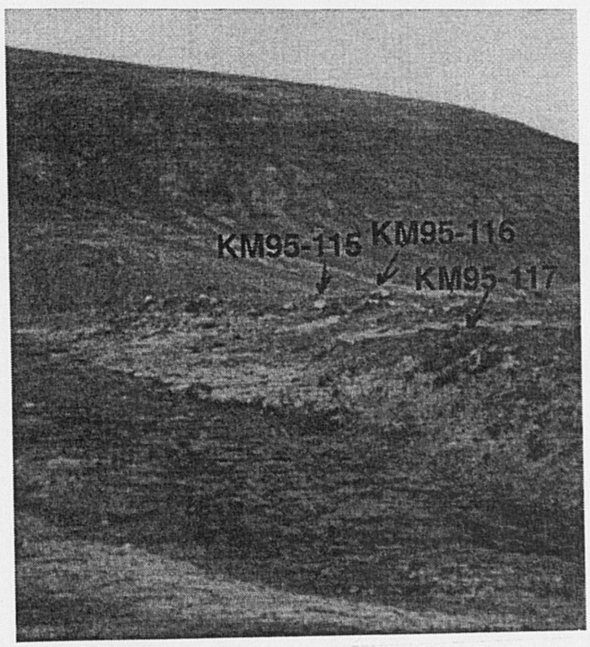


Figure 4.13

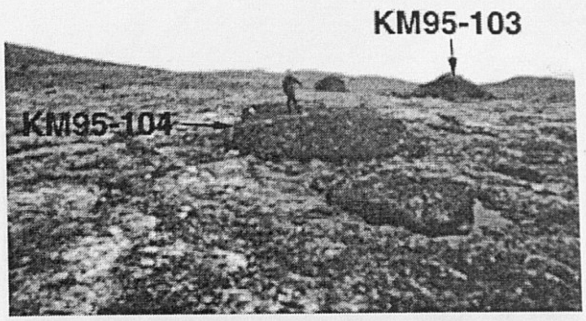


Figure 4.14

Figure 4.15

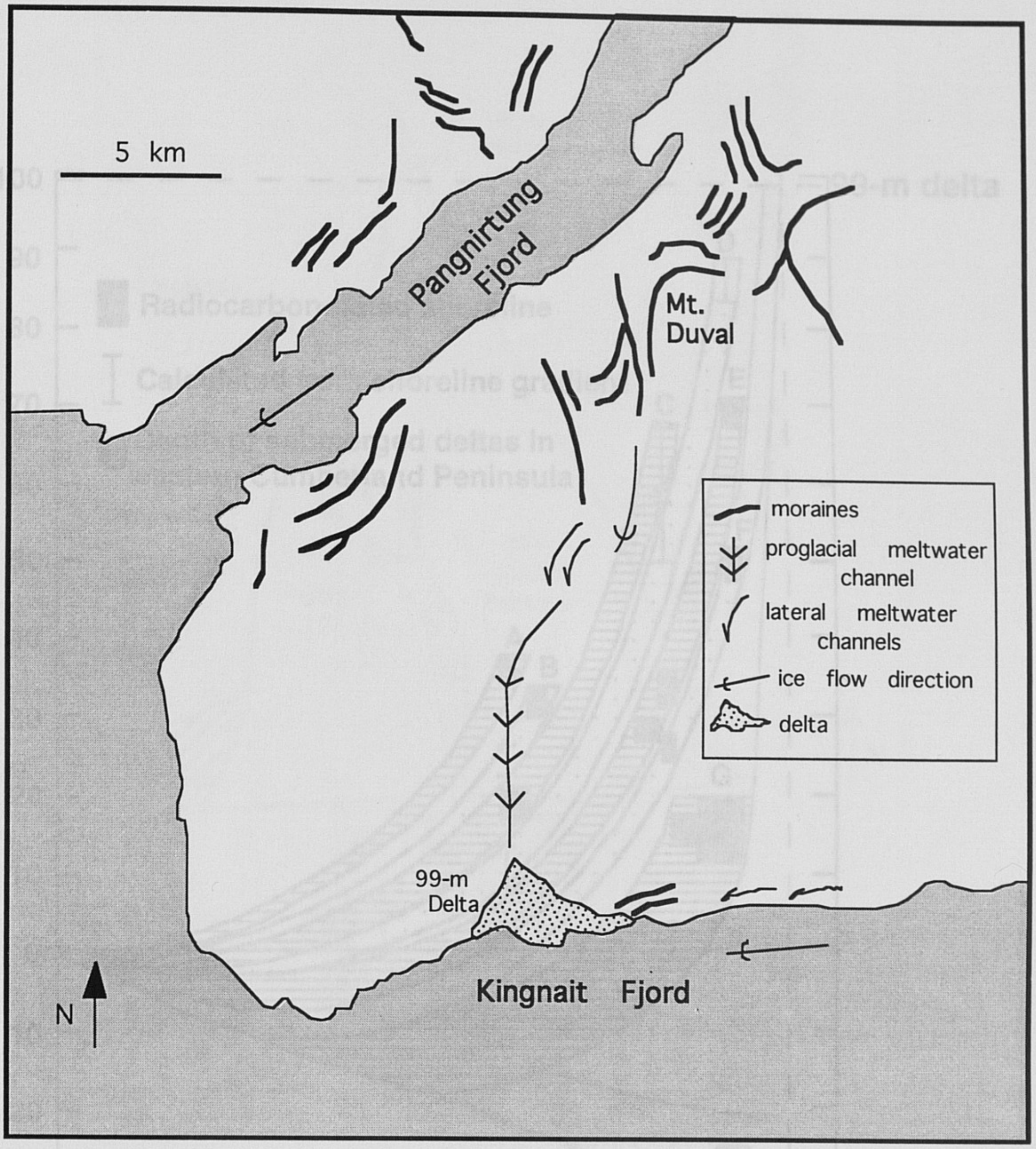
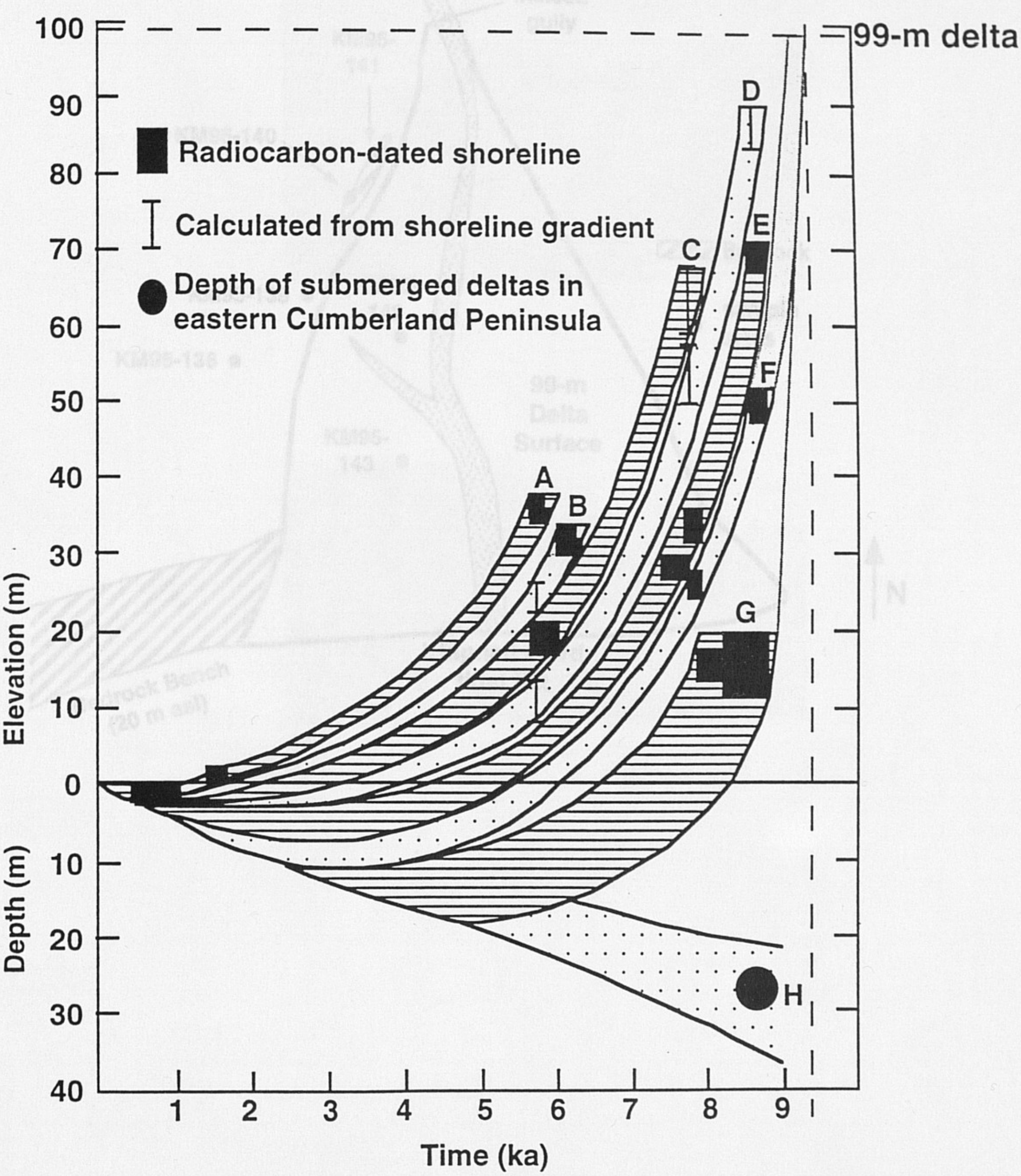

5 km
Pangnirtung Fjord
Mt. Duval
moraines
proglacial meltwater channel
lateral meltwater channels
ice flow direction
delta
99-m Delta
Kingnait Fjord
N

Figure 4.16



99-m delta
Radiocarbon-dated shoreline
Calculated from shoreline gradient
Depth of submerged deltas in eastern Cumberland Peninsula
A
B
C
D
E
F
G
H
Elevation (m)
Depth (m)
Time (ka)
100
90
80
70
60
50
40
30
20
10
0
10
20
30
40
1
2
3
4
5
6
7
8
9

Figure 4.17

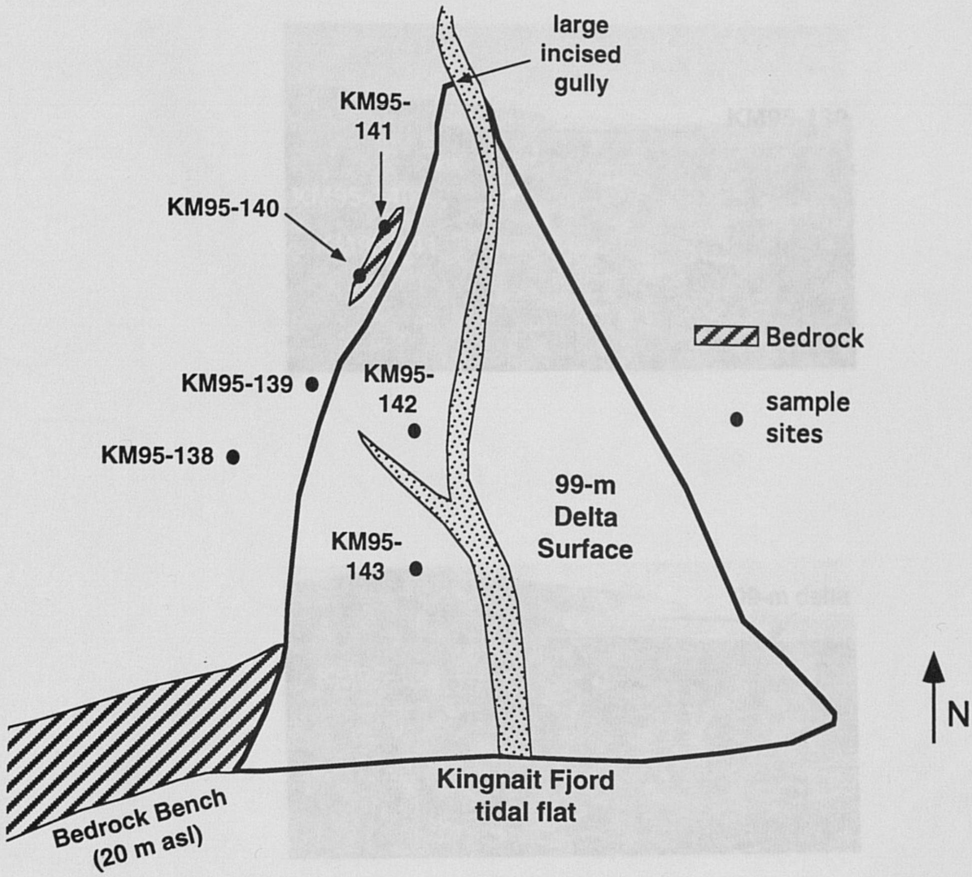

large
incised
gully
KM95-
141
KM95-140
Bedrock
KM95-139
KM95-
142
sample
sites
KM95-138
99-m
Delta
Surface
KM95-
143
N
Kingnait Fjord
tidal flat
Bedrock Bench
(20 m asl)

Figure 4.18

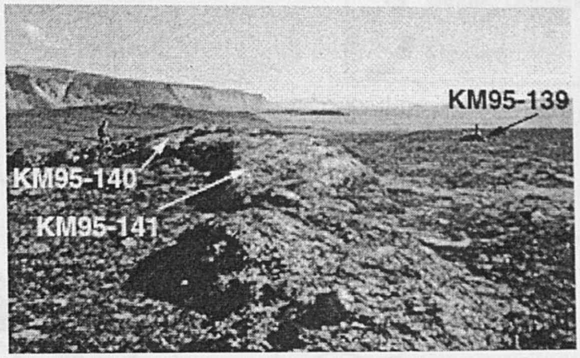


Figure 4.19

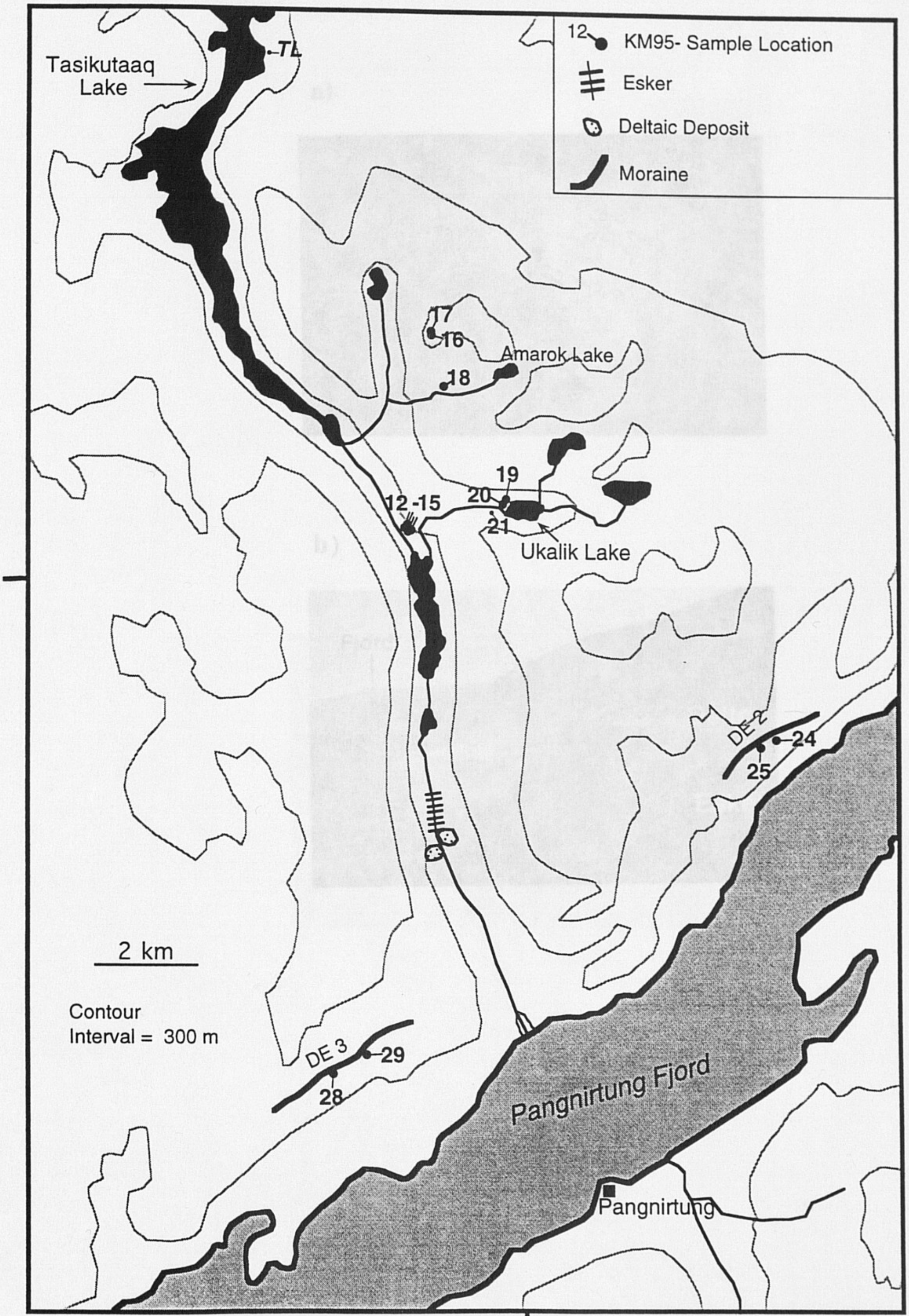
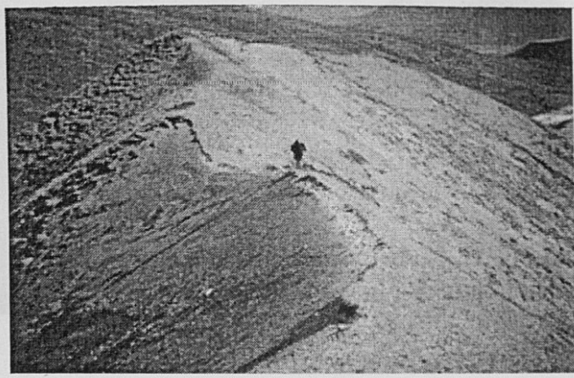

12 KM95- Sample Location
Esker
Deltaic Deposit
Moraine
Tasikutaaq Lake
TL
17
16
Amarok Lake
18
19
20
12 -15
21
Ukalik Lake
66° 15' N
DE 2
24
25
2 km
Contour Interval = 300 m
DE 3
29
28
Pangnirtung Fjord
Pangnirtung
65° 45' W

Figure 4.20

a)



b)

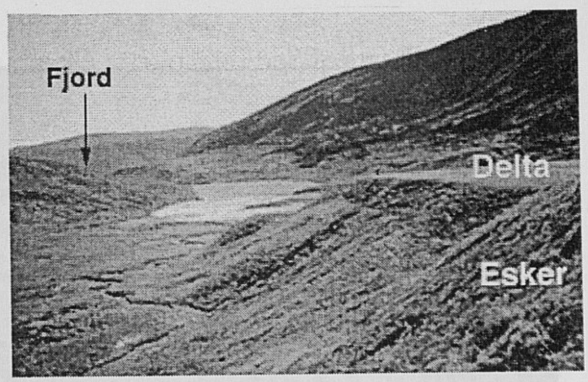


Figure 4.21

a)

b)

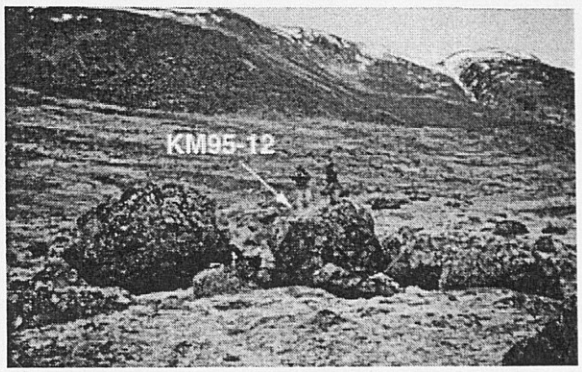


Figure 4.22

Table 4.1. Isotopic data for the type Duval region (66° 8' N, 65° 39' W), Baffin Island

Sample	Site†	Sample^ Type	Elevation (m)	^{10}Be model age (ky)§	^{26}Al model age (ky)§	Average age (ky)	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-006	TD	Bldr	500	11.9 ± 0.7	11.7 ± 0.8	11.8	6.0 ± 0.5	0.11 ± 0.006	0.68 ± 0.04
KM95-007	TD	Bldr	493	9.0 ± 0.7	9.3 ± 0.6	9.2	6.3 ± 0.6	0.09 ± 0.006	0.54 ± 0.03
KM95-034	TD	Bldr	495	8.9 ± 1.3	9.5 ± 0.6	9.2	6.5 ± 1.0	0.05 ± 0.007	0.35 ± 0.02
						MEAN= 10.1			
KM95-033	TD	Bldr	495	22.4 ± 1.1	25.3 ± 1.5	23.8	6.8 ± 0.4	0.22 ± 0.009	1.47 ± 0.07
KM95-035	TD	Bldr	470	ND	19.9 ± 1.3		ND	ND	1.13 ± 0.06
KM95-035*	TD	Bldr	470	19.8 ± 1.4	18.2 ± 1.2	19.3	5.6 ± 0.5	0.18 ± 0.012	1.03 ± 0.06
KM95-036	TD	Bldr	470	22.8 ± 2.6	21.7 ± 1.3	22.3	5.8 ± 0.7	0.14 ± 0.015	0.79 ± 0.04
						MEAN= 21.8			
KM95-039	DB	Bed	425	22.2 ± 1.1	24.9 ± 1.6	23.5	6.8 ± 0.5	0.20 ± 0.008	1.34 ± 0.07
KM95-040	DB	Bed	425	24.9 ± 1.3	26.0 ± 1.8	25.5	6.3 ± 0.5	0.23 ± 0.010	1.44 ± 0.09
						MEAN= 24.5			
KM95-005	R1	Bldr	329	14.3 ± 1.2	14.4 ± 1.0	14.3	6.1 ± 0.6	0.09 ± 0.007	0.53 ± 0.03
KM95-113	R1	Bldr	271	10.9 ± 1.3	9.5 ± 0.7	10.2	5.3 ± 0.7	0.08 ± 0.009	0.44 ± 0.03
KM95-114	R1	Bldr	271	11.2 ± 1.5	10.8 ± 0.9	11.0	5.8 ± 0.9	0.09 ± 0.012	0.51 ± 0.04
						MEAN= 11.9			
KM95-004	R2	Bldr	300	9.5 ± 2.1	9.3 ± 0.9	9.4	6.0 ± 1.4	0.06 ± 0.012	0.33 ± 0.03
KM95-115	R2	Bldr	266	10.4 ± 1.2	9.4 ± 1.1	9.9	5.5 ± 0.9	0.08 ± 0.009	0.44 ± 0.05
KM95-116	R2	Bldr	266	9.3 ± 1.3	9.9 ± 0.9	9.6	6.5 ± 1.0	0.07 ± 0.010	0.47 ± 0.04
KM95-117	R2	Bldr	271	9.3 ± 1.5	9.9 ± 0.9	9.6	6.5 ± 1.2	0.07 ± 0.011	0.46 ± 0.04
						MEAN= 9.6			

*Replicate sample, average = 3

ND=No data available

†TD=Type Duval; DE= Duval equivalent; DB= Duval bedrock; R1=1st recessional; R2=2nd recessional

^ Bldr= boulder, Bed= glacially molded bedrock

§Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

Table 4.2. Isotopic data for Duval moraine equivalents, Baffin Island

Sample	Site†	Elevation (m)	^{10}Be model age (ky)§	^{26}Al model age (ky)§	Average age (ky)	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-009	DE-1	665	13.3 ± 0.8	14.6 ± 1.0	14.0	6.6 ± 0.5	0.15 ± 0.007	0.99 ± 0.06
KM95-024	DE-2	455	39.2 ± 1.8	39.5 ± 2.5	39.4	6.1 ± 0.4	0.36 ± 0.012	2.18 ± 0.12
KM95-025	DE-2	445	9.2 ± 0.6	10.2 ± 0.8	9.7	6.7 ± 0.6	0.08 ± 0.005	0.55 ± 0.04
KM95-028	DE-3	375	9.0 ± 0.6	9.4 ± 0.7	9.2	6.4 ± 0.6	0.08 ± 0.004	0.49 ± 0.03
KM95-029	DE-3	380	14.1 ± 0.9	14.8 ± 1.1	14.5	6.4 ± 0.6	0.12 ± 0.007	0.78 ± 0.05
					***MEAN= 11.8**			

†DE= Duval moraine equivalents; all samples are boulders; DE-1 labeled on Fig. 4.9; DE-2, DE-3 labeled on Fig. 4.19

§Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

*averaged without KM95-024, including KM95-024, MEAN=17.3 ky

Table 4.3. Isotopic data for samples stratigraphically beyond the Duval limit, Baffin Island

Sample*	Site†	Sample Type	Elevation* (m)	^{10}Be model age (ky)§	^{26}Al model age (ky)§	Average age (ky)	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-010	WZ-2	Bldr	728	12.5 ± 0.6	12.4 ± 0.8	12.5	6.0 ± 0.4	0.15 ± 0.006	0.88 ± 0.05
KM95-011	WZ-2	Bldr	703	9.7 ± 0.6	9.6 ± 0.6	9.7	6.0 ± 0.5	0.11 ± 0.006	0.68 ± 0.04
KM95-003	WZ-2	Bldr	689	65.8 ± 2.7	59.0 ± 3.6	62.4	5.7 ± 0.3	0.71 ± 0.021	4.04 ± 0.20
KM95-002	WZ-2	Bldr	657	21.4 ± 1.0	21.2 ± 1.6	21.3	6.0 ± 0.5	0.23 ± 0.009	1.41 ± 0.09
KM95-101	WZ-2	Bldr	521	23.2 ± 1.3	21.3 ± 1.4	22.3	5.6 ± 0.4	0.23 ± 0.010	1.26 ± 0.07
KM95-103	WZ-2	Bldr	521	9.2 ± 0.6	8.8 ± 0.7	9.0	5.8 ± 0.5	0.09 ± 0.006	0.53 ± 0.04
KM95-104	WZ-2	Bldr	521	9.8 ± 0.7	9.5 ± 0.8	9.7	5.9 ± 0.6	0.10 ± 0.006	0.58 ± 0.05
KM95-100**	WZ-2	Bldr	516	12.1 ± 2.9	11.2 ± 0.8	11.6	5.6 ± 0.3	0.12 ± 0.015	0.67 ± 0.04
KM95-102	WZ-2	Bldr	501	10.6 ± 0.7	11.7 ± 1.2	11.2	6.7 ± 0.7	0.10 ± 0.006	0.68 ± 0.06
KM95-105	WZ-2	Bldr	431	9.6 ± 0.7	9.6 ± 0.8	9.6	6.1 ± 0.6	0.09 ± 0.005	0.52 ± 0.04

*Samples are presented in altitudinally descending order

†WZ-2=stratigraphically beyond Duval moraines

**Be and Al measurements from two different aliquots of sample KM95-100

§Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

Table 4.4. Isotopic data for the 99-m delta, Kingnait Fjord (65° 58' N, 65° 44' W), Baffin Island

Sample	Site†	Sample^ Type	Elevation (m)	^{10}Be model age (ky)§	^{26}Al model age (ky)§	Average age (ky)	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-138	Above	Bldr	103	13.9 ± 1.2	10.4 ± 0.8	12.2	4.6 ± 0.5	0.09 ± 0.007	0.42 ± 0.03
KM95-139	Above	Bldr	103	9.3 ± 1.8	9.7 ± 1.1	9.5	6.3 ± 1.4	0.06 ± 0.012	0.39 ± 0.04
KM95-140	Above	Bed*	103	10.5 ± 3.1	10.0 ± 0.7	10.3	5.8 ± 1.7	0.07 ± 0.020	0.40 ± 0.03
KM95-141	Above	Bed*	103	10.5 ± 0.7	10.5 ± 0.8	10.5	6.1 ± 0.6	0.07 ± 0.004	0.43 ± 0.03
					Mean=	**10.6**			
KM95-142	On	Bldr	99	9.9 ± 2.6	9.5 ± 1.0	9.7	5.8 ± 1.6	0.06 ± 0.016	0.37 ± 0.04
KM95-143	On	Bldr	99	13.3 ± 2.3	12.8 ± 1.2	13.0	5.9 ± 1.1	0.09 ± 0.015	0.50 ± 0.04
					Mean=	**11.4**			

†Above=samples stratigraphically above the delta surface, On=samples on delta surface

^Bldr= boulder, Bed= molded bedrock

*same outcrop

§Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

Table 4.5 Isotopic data for the Kolik River region (66°12' N, 65° 48' W), Baffin Island

Sample	Site[†]	Sample^ Type	Elevation (m)	^{10}Be model age (ky)[§]	^{26}Al model age (ky)[§]	Average age (ky)	^{26}Al/^{10}Be	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})
KM95-012	KV	Bldr	280	11.8 ± 0.9	12.1 ± 0.9	12.0	6.2 ± 0.6	0.09 ± 0.007	0.58 ± 0.04
KM95-013	KV	Bldr	280	11.3 ± 1.8	9.0 ± 0.7	10.2	4.8 ± 0.8	0.09 ± 0.014	0.43 ± 0.03
KM95-014	KV	Bed	280	13.0 ± 1.9	13.2 ± 1.0	13.1	6.2 ± 1.0	0.10 ± 0.015	0.64 ± 0.04
KM95-015	KV	Bed	280	8.0 ± 1.2	9.5 ± 0.9	8.8	7.2 ± 1.2	0.06 ± 0.009	0.45 ± 0.04
						Mean= 11.0			
KM95-030	KE	Bldr	25	16.0 ± 3.7	9.1 ± 1.3	12.5	3.5 ± 0.9	0.12 ± 0.028	0.42 ± 0.06
KM95-016	AL	Tor	928	128.7 ± 5.5	100.1 ± 5.7	114.4	4.7 ± 0.2	1.79 ± 0.051	8.34 ± 0.37
KM95-017	AL	Cobble	928	32.3 ± 1.4	28.6 ± 1.8	30.4	5.4 ± 0.3	0.43 ± 0.014	2.28 ± 0.12
KM95-018	AL	Bldr	723	13.9 ± 0.7	13.2 ± 0.9	13.5	5.8 ± 0.4	0.17 ± 0.007	0.96 ± 0.06
KM95-019	UL	Bldr	524	22.0 ± 1.0	23.5 ± 1.5	22.7	6.5 ± 0.4	0.21 ± 0.008	1.38 ± 0.07
KM95-020	UL	Bed	524	10.4 ± 0.6	12.6 ± 0.8	11.5	7.4 ± 0.6	0.10 ± 0.005	0.76 ± 0.04
KM95-021	UL	Bed	559	10.6 ± 0.7	12.3 ± 0.9	11.5	7.0 ± 0.6	0.11 ± 0.006	0.75 ± 0.05

[†]KV=Kolik valley; AL= Amarok Lake valley; UL= Uqalik Lake valley; KE=Kolik esker

^ Bldr= boulder, Bed= glacially molded bedrock, Tor=weathered bedrock, cobble=fresh erratic

[§]Uncertainty represents analytical precision propagated with 3% uncertainty in production rates resulting from barometric altitude determination

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APPENDIX A: Elevation corrections

Elevation data were collected in the field using a hand held altimeter; because altimeters are sensitive to changes in barometric pressure we also used two portable barometric loggers. One of the loggers was kept stationary at the Parks Canada garage in Pangnirtung, which is 34 m asl. The other logger was carried with us in the field. Altimeter readings were taken for each sample site and the date and time of the readings were recorded. The altimeter was originally set to 75 meters at the Pangnirtung campground, which is 5 m asl, to avoid recording negative altitude reading while in the field. In order to correct the altitudes measured in the field, we used the data from our stationary logger, as well as barometric pressure measurements taken at the Pangnirtung airport. The Pangnirtung airport records discrete barometric data, which we smoothed by using a running 5 point average. The correction equation used was:

Altimeter elevation (measured in the field) + [(34 m - 5 pt. running average) - 70 m]

The 34 m value represents the position of the portable barometric logger, whereas the 70 m value represents the difference between the altitude of the Pangnirtung campground and the altitude to which we originally set the altimeters.

Using the correction, some elevation data at known locations were not accurately corrected, such as the 99-m delta, which had a corrected altitude of 63 m asl, the greatest error noted. Therefore, I used a 30-m uncertainty for the elevation data, which results in a propagated model exposure age uncertainty of 3%.