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**GLACIAL CHRONOLOGY OF THE SOUTH BYLOT AND SALMON RIVER
LOWLANDS, N.W.T., USING ERRATIC DISPERSAL PATTERNS,
COSMOGENIC DATING, RADIOCARBON DATING AND LICHENOMETRY**

by
Shirley J. McCuaig

A thesis submitted to the Faculty of Science
in partial fulfillment of the requirements
for the degree of Master of Science

Department of Earth Sciences

Carleton University

Ottawa, Ontario

May, 1994

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N.W.T., USING ERRATIC DISPERSAL PATTERNS,
COSMOGENIC DATING, RADIOCARBON DATING AND LICHENOMETRY**

Submitted by Shirley McCuaig, B.Sc.
in partial fulfillment of the requirements for
the degree of Master of Science

Thesis Supervisor

Chair, Department of Earth Sciences

Carleton University
April 1994

Abstract

Various methods were used to determine the timing and extent of glacial events spanning the Pleistocene and Holocene Epochs in the South Bylot and Salmon River Lowlands, N.W.T., Canada. Events that occurred prior to the Late Wisconsinan Substage are recorded in the area, but have been difficult to date with classical dating methods. Cosmogenic dating has provided some of the first pre-Late Wisconsinan absolute dates for the Canadian Arctic.

The dispersal patterns of glacial erratics revealed three distinct zones of glacial incursion, one of native origin (related to local ice bodies) and two of foreign origin (related to continental ice sheets). The clustering of ^{10}Be cosmogenic dates around 60 ka BP, regardless of dispersal zone, indicates that the Eclipse Glaciation is likely to have reached its maximum in Early Wisconsinan (Early Foxe) time. The Sermilik Nonglacial Period began with the retreat of Eclipse ice 60 ka ago, and has continued up to the present in most of the study area. Ice appears to have been restricted to the channels separating Baffin and Bylot Islands during the subsequent Button Glaciation. Holocene (?) native ice retreat may have occurred in some areas, as indicated by lichenometric relative age groupings of 49-52 mm and 55-59 mm.

Cosmogenic dating was used to: 1) date native and foreign glacial events of pre-Late Wisconsinan age in the study area, 2) provide constraints on the timing of glacial events and the extent of ice at the northeastern margin of the Laurentide Ice Sheet, and 3) make inferences about the stability of native ice bodies.

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Chapter 1

Introduction

The glacial history of Arctic Canada is an important key to grasping the dynamics of both extensive regional ice sheets and smaller, areally restricted ice bodies. The resolution of the timing of glacial events, which is crucial to glacial history, has proven to be an onerous task. Absolute dating in Quaternary research has often been confined to the last 50 000 years, which is the limit of the radiocarbon dating technique. Other dating methods, such as amino acid racemization and uranium series dating have met with limited success. An applicable method for dating events beyond the radiocarbon limit is clearly required to accurately identify older glacial events in the Arctic, allowing the placement of such events in their proper context.

A study site was chosen for the thesis in an area where glacial geological research has been limited to reconnaissance level study, which indicated that the glacial terrain is greater than 50 ka in age (Klassen, 1993). Highly contrasting regional bedrock lithologies are sources of abundant glacial erratics, which provide powerful constraints on glacial interpretations and thus make the site ideal for glacial research. The study area, located on the northwest coast of Baffin Island, encompasses the South Bylot and Salmon River Lowlands of Bylot and Baffin Islands, respectively (figure 1.1).

The main objective of this thesis is to provide some insight into the sequence of glacial events in the northwest Baffin Island region, building on reconnaissance work by Klassen (1993), which has provided a stratigraphic framework suggestive of events of pre-Late Wisconsinan age in the area. This study aims to produce a more detailed event chronology spanning the Pleistocene and Holocene Epochs, through the application of several dating methods. The glacial chronology developed from this work will be used to further clarify both past movements of local ice bodies and the history of the Laurentide

Figure 1.1 General location map (after Kliza, 1994). The South Bylot Lowland of Bylot Island and the Salmon River Lowland of Baffin Island, which comprise the thesis area, are shown in dark grey. Land is shown in pale grey; water bodies in white; ice bodies in hatching. The town of Pond Inlet is located on the Salmon River Lowland.

Ice Sheet and its predecessor(s). If possible, glacial response will be related to global climate change.

A second objective is to reveal the absolute ages of glacial events that occurred more than 50 ka ago using the recently developed method of cosmogenic dating, which appears to have considerable potential for Quaternary research. At least three glacial events occurred in the area prior to the Late Wisconsinan Substage (Klassen, 1993), but it is not known whether one, or all, of these events occurred in Early Wisconsinan, Illinoian or pre-Illinoian time. The thesis research will serve as a further test of the cosmogenic method, as well as to potentially produce some of the first absolute dates for glacial events older than 50 ka in the Arctic Archipelago.

The final objective of this study is to provide a more detailed historical framework for future paleoclimatological work. The great antiquity of the area and its unique location with respect to continental glaciation make it an ideal site for micropaleontological, palynological and permafrost studies. To date, little paleoclimate research has been undertaken in the region (Kliza, 1994).

Physiography

The physiography of the study area is quite distinct, with mountainous highlands standing in sharp contrast to rolling lowland plains. The mountainous areas make up one section of the Davis Highlands, which are part of the Davis Region of the Canadian Shield (Bostock, 1970). The lowlands, which comprise the majority of the thesis area, are part of the Arctic Lowlands physiographic unit of flat-lying Late Proterozoic, Paleozoic, and Mesozoic sedimentary rocks (Bostock, 1970). Highland peaks reach a maximum altitude of approximately 2130 m asl, while the lowlands form gently sloping surfaces from sea level to about 500 m asl.

The northwest trending Byam Martin Mountains, which constitute the central part of Bylot Island, are flanked to the southwest by the South Bylot Lowland, which forms a fan-shaped, rolling plain that slopes gently toward Eclipse Sound. Another, smaller, lowland plain borders the northeast side of the mountain belt. The Byam Martin Mountains, with the distinct alpine character of sharp horns, aretes and cirques, are often only visible as nunataks in the Bylot Island ice field.

The Salmon River Lowland is located south of Bylot Island on adjacent Baffin Island. It is topographically similar to the South Bylot Lowland, but it is bounded on the south, east and north sides by a high, deeply dissected, plateau typical of the Davis highlands.

The Byam Martin Mountains support a large population of cirque glaciers from which numerous valley glaciers flow onto the surrounding lowlands. Some of the mountains enclosing the Salmon River Lowland also support glaciers. Part of a chain of ice covered highlands, the Bylot and Baffin Island glaciers are centrally located among the Devon Island, Ellesmere, Penny, and Barnes Ice Caps.

Few of the regional topographic features have formal names that can be found on maps or in the literature. Some features have been given informal Inuktitut epithets by the people of Pond Inlet and these, along with several informal English names proposed by the author, will be used for clarity. Figures 1.2 and 1.3 show the topographic nomenclature of the Bylot Island and Salmon River Lowlands that will be referenced in the thesis.

Climate

The two climatic zones that occur in the study area are directly related to the two physiographic units present. The highland portions of the study area fall within the Oceanic High Arctic Ecoclimatic Region, while the lowlands are considered Mid Arctic (figure 1.4) (Ecoregions Working Group, 1989).

Legend. This legend applies to all following maps, unless stated otherwise.

Figure 1.2. Location map of the South Bylot Lowlands

Figure 1.3. Location map of the Salmon River Lowlands

Figure 1.4. Climatic zones of the study area (after Ecoregions Working Group, 1989). Land is shown in grey; water bodies in white; ice bodies in hatching.

The Oceanic High Arctic highland regions comprise a small part of the thesis area and have extremely cold winters, with mean daily temperatures ranging from -32° to -36° C (Ecoregions Working Group, 1989) from December to March. July and August are the only "warm" months, with mean daily temperatures that climb above 0° C. Annual snowfall is about 100 mm, although it is higher in areas closer to Baffin Bay (Ecoregions Working Group, 1989).

The lowlands, which comprise the majority of the thesis area, are something of an oasis in an otherwise High Arctic region. The low lying plains are sheltered by the surrounding highlands, allowing Mid Arctic climate to extend into the area. The South Bylot and Salmon River Lowlands experience short summers in July and August, where mean monthly temperatures are about 5° C (Environment Canada, Atmospheric Environment Service, 1982), although warmer days are not uncommon. Winters are long and cold, but not as cold as those in the ice-covered highlands. Mean daily temperatures in the town of Pond Inlet from December to March range from -28.4° to -32.7° (Environment Canada, Atmospheric Environment Service, 1982). Precipitation is less than 200 mm/a (Environment Canada, Atmospheric Environment Service, 1982), falling mainly as snow, which remains on the ground from September to June (Ecoregions Working Group, 1989).

Vegetation

The development of biomass in the study area is a reflection of local climate, although other factors, such as substrate type and periglacial activity, are certainly important. The ice-free parts of the highland areas are sparsely vegetated, probably owing to the combination of harsh climate and the lack of stable growth surfaces due to the high degree of frost shattering. In contrast, the lowland oasis is relatively lush.

Mid Arctic tundra vegetation, consisting of grasses, mosses, sedges, lichen and flowers, is well developed in the lowland regions (plate 1.1 and 1.2). Only 10 species make up the bulk of the regional flora, but there are representatives of many species actually present (Drury, 1962).

Floral distribution is highly dependent on microclimate. Flowers commonly occur on sunny south-facing slopes, with the exception of Arctic bell heather, which grows preferentially on shaded north-facing slopes (Drury, 1962). Sedges and mosses proliferate in poorly drained localities, such as valley floors. Lichen species dominate in areas exposed to strong winds, while lee slopes tend to be moss covered (Drury, 1962). A striking example of this relationship was seen near Aktineq Glacier, on the windswept crest of Etajaguaq Mountain. Lichen were observed on wind exposed boulder surfaces, whereas moss formed a triangular tundra mat in the lee of the same rocks, defining the limit of wind protection (plate 1.3). No vegetation was evident between the boulders, which are glacial erratics, because the poorly lithified sandstone bedrock is currently undergoing slow deflation by the driving winds.

Common flowering plants noted by the author during the course of field work include Arctic poppy (*Papaver lapponicum*), Arctic avens (*Dryas integrifolia*), Arctic bell heather (*Cassiope tetragona*) and a purple flower, possibly purple saxifrage (*Saxifraga oppositifolia*). Mountain sorrel (*Oxyria digyna*) and nodding bladder campion (*Melandrium apetalum ssp. arcticum*) were less abundant. Dwarf willow (*Salix arctica*) is common in wetter areas.

The South Bylot Lowland has been judged to be "rich in calciphile [species]" (Drury, 1962), in spite of a lack of calcareous bedrock. The source of calcareous nutrients for calciphilic plant species is probably related to glacially transported carbonates, although Cretaceous arkosic concretions cannot be ruled out. Field work in the South Bylot Lowland has often shown, however, that far-travelled carbonates are much more numerous than the calcareous concretions of local provenance.

Periglacial Features

The study region exhibits pervasive evidence of periglacial activity. Ice wedge polygons, sorted circles, mud boils, soil stripes, and solifluction were observed at various locations in the study area. Frost-shattered metamorphic bedrock characterizes vegetation-free parts of the highlands. The depth and movement of the active layer is, in fact, one of the controlling factors in floral distribution (Drury, 1962). Maximum depth to the frost table in summer ranges from 25 cm in organic terrain to 125 cm in some vegetation-free areas (Drury, 1962).

The South Bylot and Salmon River Lowlands are replete with lakes and ponds. The outwash plains of Ujarasukjuit and Ujarasukjulik Glaciers exhibit pronounced kame and kettle topography which is particularly conspicuous near the snouts of the two glaciers (figure 1.2). Kettle lakes also characterize the recessional moraine zone of the now nonexistent Utuk Glacier, which must have been a piedmont lobe issuing from Utuk Lake Valley (figure 1.3). Kettle lakes can also be seen in a wide band fronting the South Bylot coastline, but the hummocky topography of the belt is significantly more subdued in relief than that of the glacial outwash plains.

Eolian Features

Strong regional winds and abundant poorly lithified sandstone deposits combine to produce localized eolian features. As mentioned earlier, eolian activity is an important controlling factor in plant distribution (Drury, 1962), with the important exception of lichen, which are little affected. Sand blasting by such eolian activity is, however, restricted to areas close to loose sediment sources. Since wind carries sand in an air layer about 1 m thick above the ground surface, the sand particles impact the ground surface often, in a process called saltation. In arid areas, sand grains can travel long distances in

this manner. However, in the tundra environment of the study area, the clasts become wet upon impact with the tundra, which renders them too heavy and too cohesive to be transported by wind. The tundra thus acts as a sponge, absorbing eolian clasts, so that eolian landscapes are only observed in areas proximal to the source of eolian debris.

Several examples of such eolian terrain were encountered in the study area. Ventifacts were observed on the modern beach near the mouth of Aktineq Creek. Cobble and pebble pavements are common in wind buffeted areas, such as mountain tops and beaches. Etajaguaq Mountain, for example, has an interesting lag deposit, consisting of 1 cm diameter spheres and smaller fragments of lithified Cretaceous-Tertiary sandstone resting on unlithified sands of the same formation (plate 1.4 and 1.5). Just east of the outwash plain of Aktineq Glacier, calcareous concretions of Cretaceous-Tertiary bedrock were observed resting on the same unlithified sandy bedrock. The interesting feature in this locality, however, is that the surrounding pebble pavement consists of clasts of foreign as well as native provenance.

Eolian activity may have been more important in the past, when vegetation was not so well established. Although some sandy sediments around Aktineq Glacier may be eolian in origin, no unequivocal loess deposits were found.

Chapter 2

Previous Work

Ice Sheet Configuration

Several models have been proposed for the development and decay of the Late Wisconsinan Laurentide Ice Sheet, but the configuration of its Early Wisconsinan predecessor(s) remains poorly resolved. The difficulty is that later glaciations tend to erase evidence of former ones, and historically there has been lack of suitable, versatile dating methods that apply to features of Early Wisconsinan or greater age. The thesis area is one of the few localities in which it is possible to identify, and thus possibly date, events of pre-Late Wisconsinan age, as the evidence has not been obliterated by the advance of Late Wisconsinan ice.

Vincent and Prest (1987) provide a model of "maximum Post-Sangamonian *sensu stricto*" Laurentide ice extent that may actually represent the Early Wisconsinan configuration (figure 2.1). The authors have cautioned, however, that in the absence of an adequate dating method, it can be argued that the glacial maximum occurred either before, during, or after the Early Wisconsinan Substage. Regardless of its age, the pre-Late Wisconsinan Laurentide Ice Sheet appears to have been more extensive than the Late Wisconsinan version.

The configuration of Laurentide ice during the Middle Wisconsinan Substage is similarly uncertain due to the problems inherent in Quaternary dating methods. Although radiocarbon dates of bone, peat or wood are reasonably accurate up to 50 ka, dates on shells, which are in many cases the only organic substances available for dating, become less reliable beyond 20 ka. As a result, three hypothetical Middle Wisconsinan ice configurations are proposed by Dredge and Thorleifson (1987) (figures 2.2, 2.3 and 2.4). Hypothesis I shows a limited Middle Wisconsinan retreat, with extensive ice in the Arctic;

Figure 2.1. Maximum Post-Sangamonian *sensu stricto* configuration of the Laurentide Ice Sheet, which may represent the early Wisconsin configuration (after Vincent and Prest, 1987). The ice sheet is shown in white; land in grey; water in dark grey.

Figure 2.2. Middle Wisconsinan configuration of the Laurentide Ice Sheet – Hypothesis I (after Dredge and Thorleifson, 1987). The ice sheet is shown in white; land in grey; water in dark grey.

Figure 2.3. Middle Wisconsinan configuration of the Laurentide Ice Sheet – Hypothesis II (after Dredge and Thorleifson, 1987). The ice sheet is shown in white, land in grey; water in dark grey.

Figure 2.4. Middle Wisconsinan configuration of the Laurentide Ice Sheet – Hypothesis III (after Dredge and Thorleifson, 1987). The ice sheet is shown in white; land in grey; water in dark grey.

hypothesis II shows a significant amount of retreat, leaving several isolated ice caps; and hypothesis III is similar to hypothesis I, but incorporates an "embayment" in Hudson Bay in order to account for possible open water conditions at this time.

The uncertainties in the Early and Middle Wisconsinan ice sheet reconstructions serve to emphasize how serious the lack of older finite dates can be to the study of glacial geology.

Figure 2.5 illustrates the Late Wisconsinan ice configuration proposed by Dyke and Prest (1987), including the Cordilleran ice complex. The boundaries of the ice sheet of this time period are much better defined than those of its predecessors because it falls within the datable range of the radiocarbon method.

A comparison of figure 2.1 and figure 2.5 shows that the study area was glaciated in the past, possibly in Early Wisconsinan time, but was left unglaciated in the Late Wisconsinan. The thesis area, one of the few localities in Canada where an ancient glacial signature has been preserved, thus provides a rare window into Canada's pre-Late Wisconsinan past. Pinpointing the age of the continental-scale glacial event that affected the area will have important implications for the ice sheet portrayed in figure 2.1.

Bedrock Geology

The regional pattern of distinctive bedrock lithologies places important constraints on the interpretation of both ice flow directions and areal ice extents. The study area was affected by both continental-scale glaciations (figures 2.1 to 2.5) and glaciations related to local ice bodies. For clarity, the continental-scale glaciations will be referred to as foreign events, while the local glaciations will be referred to as native, after the terminology of Klassen (1993). Native ice bodies are still present in the region today (figure 1.1). In this study, the easily determined provenance of glacial erratics is used to differentiate between areas affected by foreign glaciations and areas affected by native events. Erratics of

Figure 2.5. Maximum Late Wisconsinan configuration of the Laurentide and Cordilleran Ice Sheets (after Dyke and Prest, 1987). The ice sheets are shown in white; land in grey; water in dark grey.

lithologies that do not match the local bedrock conclusively indicate ice flow into the area from distant bedrock sources and hence are reliable indicators of the former presence of foreign ice bodies.

A map of bedrock geology is shown in figure 2.6 (from Klassen, 1993), which provides a broad overview of possible sources of glacial erratics. The geological evolution of the area began in Precambrian time and is characterized by rifting, which began in the Proterozoic and was reactivated in Early Paleozoic and in Mesozoic time.

The Precambrian core of Bylot Island and most of Baffin Island comprises metasedimentary and metavolcanic rocks of Archean and Aphebian age. These felsic to intermediate, amphibolite to granulite grade gneisses are crosscut in places by granite and pegmatite dykes (Jackson and Davidson, 1975). The highlands of the study area are composed almost entirely of these lithologies. The Mary River Iron Formation also formed in Archean-Aphebian time, but only outcrops in a small area of Baffin Island, about 150 km south of the South Bylot Lowland.

Northwest trending grabens produced by extensional tectonics in Neohelikian (Proterozoic) and Early Paleozoic time created depositional basins, in which several suites of sedimentary rock were deposited (Jackson and Davidson, 1975). The Nauyat basalts of Borden Peninsula were emplaced along the major faults of the rift zone. Sedimentary units of Proterozoic age outcrop in western and northern Bylot Island, and also form an extensive belt extending from Admiralty Inlet to the area just southeast of Milne Inlet. This sedimentary package records both a transgressive and a regressive event. The former is represented by the Adams Sound redbed and intertidal sequence, as well as the Arctic Bay fining upward sequence of sandstone, siltstone, shale and carbonates. The carbonates are well developed on Baffin Island, but poorly developed on North Bylot Island, and can be stromatolitic, but this is rare. The regressive sequences of the Proterozoic sedimentary succession are assigned to the Society Cliffs and Victor Bay Formations, and consist of shallow marine dolomite, shale, gypsum, conglomerate, and fluvial redbeds. Diabase dykes

Figure 2.6. Geologic map of northern Baffin Island region, including geological provinces (inset) (from Klassen, 1993).

of the Franklin Dyke Swarm were emplaced when faults were reactivated after the Proterozoic strata were laid down (Jackson and Davidson, 1975).

The Silurian and Ordovician limestones of Baffin Island and Brodeur Peninsula are fossiliferous, reflecting the biological radiation of the Paleozoic Era. Such fossiliferous carbonates do not outcrop in the study area, hence erratics of this group make excellent glacial indicators, along with the Proterozoic sedimentary group.

Renewed early Cretaceous to Tertiary rifting formed a new graben called the Eclipse trough, in which molasse deposits were deposited unconformably over the Precambrian basement (Miall *et al.*, 1980). These immature Cretaceous to Eocene arenites and mudstones form the lowlands of the thesis area, and are composed mainly of poorly lithified fluvial, beach and deltaic deposits, with some minor marine mudstone sequences. The fluvial Hassel Formation of this group is coal bearing (Miall *et al.*, 1980).

For purposes of erratic dispersal study, four main rock types in the Bylot and north Baffin region are defined as follows: 1) Precambrian igneous and high grade metamorphic basement; 2) poorly lithified Cretaceous-Tertiary strata; 3) Proterozoic sedimentary strata and basalts and 4) Paleozoic sedimentary strata. 1) and 2) are considered to be of native provenance, while 3) and 4) are classified as foreign in origin.

Quaternary Geology

Glacial Deposits and Features

Quaternary deposits in the area are thin and discontinuous, often occurring only as sparse boulder cover, suggesting glaciation by cold-based ice (Klassen, 1993). Till occurs as a veneer (<2 m thick), and in the form of subdued ancient moraines or prominent, ice-cored, recent moraines. Eskers are rare, although some esker-like ridges have been observed emerging from abandoned conduits in retreating ice fronts at Aktineq and

"Stagnation" Glaciers. An esker-like ridge was also observed protruding from a glacier-related icing in front of "Fountain" Glacier (Elver, 1994).

Gravelly sandurs are common in valleys draining modern glaciers. Raised glaciomarine deltas can be identified at river mouths on both lowlands, including the location of the town of Pond Inlet, which is situated on one of a series of ice-contact marine deltas just east of the mouth of Salmon River. The deltas are often rich in marine pelecypods, such as *Hiatella arctica*, but such shells are rarely found *in situ*. Raised beaches formed during Holocene isostatic rebound characterize coastal regions (plate 2.1).

Relict meltwater channels control much of the modern drainage, and are particularly noticeable in the higher altitude parts of the lowlands. A striking series of meltwater channels can be seen in a valley just east of Aktineq Glacier (plate 2.2). As mentioned earlier, eolian features such as lag deposits and ventifacts are also common, in spite of the lack of loess.

Regional Glacial History

Previous work in the regional context of the study area argues for limited Late Wisconsinan ice, but extensive pre-Late Wisconsinan ice.

England and Bradley (1978) have proposed that Late Wisconsinan ice was extremely limited in the northeast Ellesmere Island - northwest Greenland area, and may also have been limited in the rest of Greenland, North America and the Arctic Ocean, although extensive ice cover probably existed in these areas more than 35 ka ago. These authors suggest that the reason the Arctic was not heavily glaciated in the Late Wisconsinan Substage is that a period of extreme cold at this time caused expansion of the mid-latitude portions of the Laurentide Ice Sheet, but resulted in Arctic recession due to the aridity associated with extensively frozen polar seas.

On Baffin Island, Wisconsinan glacial events are assigned to the Foxe Glaciation (Andrews *et al.*, 1984). One of the major ice divides of the Laurentide Ice Sheet was the

Foxe Dome, which was either centered over the west coast of Baffin Island or over adjacent Foxe basin (Andrews and Miller, 1979). Because this important ice divide was the source of Wisconsinan glacial outflow in the eastern Arctic, events related to it are referred to as being of Foxe (Wisconsinan) age. In the Bylot Island area, Klassen (1993) found evidence of ice flow from the south in pre-Late Wisconsinan time, thus, if the foreign events of the thesis area are Early Wisconsinan (Early Foxe) in age, the Foxe Dome is likely to have been their source area. If, however, foreign events are older, it follows that a divide similar to the Foxe Dome must have existed in pre-Wisconsinan (pre-Foxe) time.

Wisconsinan (Foxe) history on Baffin Island generally records three glacial events, each one successively less extensive than the former. The maximum glacial extent on Baffin Island is thought to have been achieved prior to Late Wisconsinan time (Miller and Dyke, 1974, Andrews *et al.*, 1984, Hodgson and Haselton, 1974), and may have been reached prior to the Early Wisconsinan Stage. This event, named the Ayr Lake Stade of Foxe glaciation by Andrews *et al.* (1984), dates within marine isotope stages 4 and 5, with its glacial maximum occurring sometime between 110 and 70 ka (Andrews *et al.*, 1985). Similar amino acid ratios both above and below Ayr Lake tills imply that the maximum lasted approximately 10 ka (Fillon *et al.*, 1981). The Quajon Interstade, with a uranium series age of >69 ka, followed the Ayr Lake advance (Szabo *et al.*, 1981).

Andrews *et al.* (1984) propose that a second [unnamed] glacial maximum may have occurred sometime between 80 and 40 ka. The Iron Strand event of Labrador, dated at approximately 40 ka, is thought to indicate deglaciation after this event (Andrews *et al.*, 1984).

The final, time transgressive event was Late Wisconsinan to early Holocene in age. It occurred from 11 to 10 ka (Baffinland Stade) in southeastern Baffin Island and from 9 to 8 ka (Cockburn Substage, Miller and Dyke, 1974, Andrews and Ives, 1978) in parts of northern Baffin Island, but this event did not extend into the thesis area. Minor

advances of receding local ice sheets also occurred at this time in the south Baffin Island area. This in-phase advance of both local and Laurentide ice bodies suggests that the Cockburn event reflects climatic change (Stravers *et al.*, 1992). The glacial events of Baffin Island are summarized in table 2.1.

The Inugsuin regression, the slow isostatic rebound of Baffin Island following complete deglaciation, took place in the Holocene from 8.5 to 3 ka (Andrews *et al.*, 1984). Deglaciation of the northeastern margin of the Laurentide Ice Sheet therefore lags that of the southern edge by several thousand years, but correlates well with other events in the Arctic (Andrews *et al.*, 1984).

In the northwest Baffin region, Hodgson and Haselton (1974), also found evidence for three successively less extensive glacial phases. Many of the native glaciers and ice fields appear to have recently achieved their maximum extent since the Late Wisconsinan, since native ice is either at or retreating from 30 m high ice-cored moraines (Hodgson and Haselton, 1974).

Borehole oxygen isotope data from the Agassiz, Devon and Greenland Ice caps suggest that the Laurentide glaciations began at 115 ka (Sangamonian Stage), and terminated when significant warming took place at 10 ka, with the coldest period 15-30 ka ago (the Late Wisconsinan maximum occurred at 18 ka (Dyke and Prest, 1987)) and the full glacial condition occurring at 70 ka BP (Early Wisconsinan Substage) (Koerner and Fisher, 1985). This independent glaciological data reflects activities of the entire Laurentide Ice Sheet and so may not reflect the localized glacial history of Baffin Island (Koerner and Fisher, 1985).

Local Glacial History

A reconnaissance study of the glacial geology of Bylot Island was undertaken by Klassen (1981, 1993, Klassen and Shilts, 1987). Erratic distribution data, radiocarbon

ka BP	Stade	Interstade
8	Cockburn	
10	Baffinland	
20		
30		
40		Iron Strand
50		
60	?	
70		Quajon
80		
90	Ayr Lake	
100		
110		
		Cape Broughton

Table 2.1. Approximate timing of glacial events, Baffin Island (after Andrews *et al.*, 1984, Andrews *et al.*, 1985). The older glacial stades are shown as approximate age ranges.

dates and amino acid racemization data were used to determine the approximate timing and extent of several events, which are summarized in table 2.2.

Amino acids are organic compounds found in all living organisms, which have at least one amino group (NH_2) in their molecular structure (Bradley, 1985). These molecules have a central carbon atom, which is bonded to an amino group, a carboxylic acid group, a hydrocarbon group, and a hydrogen atom, so the compound is asymmetric, or chiral. Chiral compounds can occur in two mirror image forms (isomers), differing from each other physically only in the way that the isomers rotate plane polarized light. The two forms are designated as D (dextro) and L (Laevo), depending on the sense of this rotation. Only the L form is found in living animals. However, the D form is the naturally stable configuration, and upon death, the amino acids within an organism convert to this form in a process called racemization (Bradley, 1985).

The amino acid L-isoleucine has two chiral carbon atoms in its structure, so the potential configurations are more complex, yet it produces only one isomer: D-alloisoleucine. This very slow conversion process is technically called epimerization, but it is basically the same process as racemization (Bradley, 1985). The ratio of D-alloisoleucine to L-isoleucine (alle:Ile ratio) can be measured on a gas chromatograph to determine the time elapsed since death. The reconnaissance work by Klassen (1993) measured only the "free" (unbound) amino acids because "total" ratios are less comparable to results from other areas, due to varying sample preparation techniques.

A problem inherent in the amino acid method is that racemization is dependent on temperature, so the temperature history of the sample must be known accurately in order to obtain an absolute date. Since temperature history cannot be established with certainty in the North Baffin area, the method may only be useful for relative dating, and correlation with amino acid data from other localities can be difficult. In fact, temperature inversions due to glacial events may limit the usefulness of the method even for relative dating purposes. Contamination and leaching are also potential problems. In spite of these

Radiocarbon Age (years BP)	Amino Acid Ratio	Glacial Stade		Interstade
		Native	Foreign	
120	0.00	Neoglacial		
	0.05			
	0.10			
9500	0.15		Cape Hatt	
	0.20		Button	
>43 000	0.25			Sermilik
70-110 000?	0.30	Aktineq Phase	Eclipse	
	0.35			
	0.40			Pond Inlet
	0.45			
	0.50			
	0.55			
	0.60			Salmon River
	0.65			
	0.70		Baffin	
	0.75			
	0.80			
	?	Bylot		

Table 2.2. Glacial events in the Bylot Island-northern Baffin region (after Klassen, 1993). The chronologic sequence was obtained using amino acid ratios and stratigraphic relationships. Some of the younger events have been dated by the radiocarbon method.

problems, amino acid analyses can provide useful stratigraphic information where none was previously available. The use of amino acid racemization on marine pelecypods, in conjunction with radiocarbon dates and other stratigraphic evidence, has been an instrumental factor in determining the order, if not all the ages, of events in the Bylot - North Baffin region, and as such has provided an excellent base on which to build.

Klassen has identified four, rather than three foreign events in the area, along with several native events. It appears then, that the three Foxe events could be represented, and possibly a pre-Foxe event, implying that the glacial history chronicled in the area represents a significant period of time. The following is a brief summary of the glacial events described in the reconnaissance studies of Klassen (1981, 1993).

1) The earliest recognizable glacial event appears to be an advance of native glaciers, referred to as the Bylot Glaciation. Erratics of native provenance dominate both the highland and the lowland regions, suggesting that native ice covered most of the island at least once, possibly several times, in the distant past. The event is as old or older than Early Wisconsinan in age, predating deposits with amino acid allelic ratios of 0.6 and possibly 0.8 (table 2.2), and thus may be of pre-Foxe age. The inferred early glacial Bylot event may actually have encompassed many ancient events, but only one can be discerned reliably with the limited evidence available. The evidence for this event includes meltwater channels incised in bedrock in front of several modern glaciers, which indicate native ice of much greater extent than that of the later Aktineq glacial phase of Eclipse Glaciation (table 2.2). Eclipse moraine occupying these meltwater channels also supports a pre-Eclipse age. The event is also considered to pre-date the Salmon River Interglaciation and the Baffin Glaciation because there is no evidence of a native advance within or between the sediments that represent these events.

2) Later, foreign ice flowed northward from Baffin Island to Bylot Island during the Baffin Glaciation, a glacial event which may also be pre-Foxe in age. Quartz arenites occurring at high elevations make up the majority of the foreign erratics of this event.

Proterozoic erratics, probably derived from the Adams Sound Formation near Milne Inlet, were found as high as 1130 m asl on Bylot Island, suggesting that a large, regional ice sheet may have crossed the Byam Martin Mountain divide in several places. Similar quartz arenite erratics were found at 820 and 700 m asl. These three sites are the highest elevations at which foreign erratics have been found in the area, suggesting that the Baffin advance was the most extensive foreign event to affect the area.

3) The Salmon River Interglacial Period followed. Paleosols and fossils in sections at Aktineq Creek, Salmon River, Cache Creek (northeast Bylot Island), Cape Graham Moore (just north of Button Point) and near Oliver Sound indicate a major deglaciation. Klassen (1993) suggested that the Salmon River Interglacial amino acid *Ala:Ile* ratios of 0.55 - 0.60 (table 2.2) may correlate with the 300-500 ka Cape Christian Interglacial of Miller *et al.* (1977).

4) The next inferred event also represents nonglacial conditions, but as there is no evidence of an intervening glaciation, it could be a continuation of the Salmon River Interglacial (Klassen, 1993). The Pond Inlet nonglacial event is evidenced in raised marine sections at Button Point, Canada Point (west coast of Bylot Island) and Janes Creek. *Ala:Ile* ratios of 0.35 - 0.40 (table 2.2) indicate that this warm climatic event is younger than the Salmon River Interglaciation.

5) The Eclipse Glaciation, the second recognized advance of foreign ice onto Bylot Island, was fairly extensive, but not so extensive as the earlier Baffin Glaciation. Erratics and glacial deposits related to this event indicate that ice only advanced up to the 370 m contour in the South Bylot Lowlands and at least to 600 m in the mountainous Lancaster Sound region. Eclipse ice moved through Milne Inlet, bifurcating eastward along Eclipse Sound and northward through Admiralty Inlet. The north arm then flowed eastward down Lancaster Sound. Ice must have been grounded in the channels at this time, because it was able to cross the 500-1000 m deep channels separating Bylot and Baffin Islands (Klassen, 1993, Hodgson and Haselton, 1974).

The provenance of foreign erratics indicates that a regional ice sheet centered in the south [probably the Foxe Dome or similar predecessor] was responsible for the event. On the outer north shore of Bylot Island, lichen-free carbonate-rich moraines of Eclipse age stand out in sharp contrast to lichen-covered Archean bedrock, demarcating the inland limit of foreign ice at this time. Similarly distinctive moraines of Eclipse age are not evident in the thesis area, but various easily recognized carbonate and quartz arenite erratics can be associated with this event. Klassen found Mary River Iron Formation erratics on the plateau east of Milne Inlet and on Mount Herodier, near Pond Inlet. These erratics are strong evidence for regional ice flow from the south, since the bedrock source is limited to a small region to the south (figure 2.6).

Klassen (1993) also noted that during Eclipse Glaciation some of the native glaciers may have been confined within the Byam Martin Mountains by the considerable pressure of the larger foreign ice sheet as it advanced into the area. "Foreign debris occurring inland of modern glaciers, in some cases to elevations more than several hundred meters above them, indicates that foreign ice caused a reversal of the normal directions of outward flow and 'backfilled' valleys" (Klassen, 1993, p. 40). The volume and mass of native ice dammed within the highlands would have been higher than it would have been under normal circumstances, due to continuing accumulation but restricted ablation. The retreat of foreign ice resulted in the redistribution of this excess ice volume, allowing the glaciers to advance unusually far into the lowlands. However, the rate of accumulation was not high enough to keep native ice in these maximum positions, so it rapidly retreated, approximately to the positions in which it is found today. "This native advance is named Aktineq glacial phase and is interpreted as a late glacial (deglacial) phase of Eclipse glaciation" (Klassen, 1993, p. 45). Utuk (now nonexistent) and Ujarasukjulik Glaciers are the only glaciers in the thesis area to have left obvious recessional moraines thought to have formed during the Aktineq glacial phase. Ujarasukjuit and Aktineq Glaciers also advanced at this time, as evidenced by stagnation deposits in front of these glaciers, but

the extent of the advance is less certain. In some parts of northern Bylot Island, Eclipse ice did not reach the termini of native glaciers, as there is no evidence of interaction or overriding by foreign ice in these areas (Klassen, 1993, figures 35 and 37). Klassen therefore suggested that native ice which was not "overpowered" by Eclipse ice did not advance during the Eclipse Glaciation. He further suggested that native glaciers have not been more extensive than they are now since Eclipse time. On the east side of the outer ice-cored neoglacial moraine of Aktineq Glacier, a peat dated at 7860 ± 100 a BP (GSC-2541) was found stratigraphically below a peat dated at 450 ± 100 a BP. Boulders from the neoglacial moraine lie on top of these deposits, implying that Aktineq Glacier is currently as far advanced as it has been in at least the last 7900 years and that the neoglacial moraines are very young (Dilabio and Shilts, 1978). A radiocarbon date from material collected during field work for this thesis supports this interpretation and will be discussed later. A $33\,300 \pm 800$ a BP date on marine shells (possibly not *in situ*) from near the Pond Inlet airstrip (Hodgson and Haselton, 1974) suggests that the Eclipse event is more than 33 ka old. A date on marine shells from Sermilik nonglacial deposits, which postdate the Eclipse event, indicates that this event occurred more than 43 ka ago (Klassen, 1993). It should be kept in mind, however, that radiocarbon dates on shells become less reliable beyond 20 ka, as outlined earlier.

The Eclipse Glaciation has been correlated with the Ayr Lake Stade of the Foxe Glaciation, using deep sea records of detrital carbonate influx (Andrews *et al.*, 1985) and similarities in ice extent and amino-acid stratigraphy (Klassen, 1993). If this correlation is correct, the Eclipse event may have occurred between 110 ka and 70 ka BP (table 2.2), the tentative age range of the Ayr Lake Stade.

6) The Sermilik nonglacial event followed the Eclipse Glaciation. All the amino acid ratios of 0.20 - 0.25 from sections at Aktineq Creek and near the snout of Sermilik Glacier, as well as radiocarbon dates of >43 ka on marine molluscs [which may be

unreliable] of a deltaic deposit at Canada Point provide some indication of timing (Klassen, 1993).

7) The Button Glaciation, the third foreign event identified by Klassen, may represent one or more glacial stades. Marine shells from diamictos and tills in limited exposures in stratigraphic sections at Button Point and Sermilik Glacier have low alle:lle ratios of 0.15 to 0.20 (table 2.2). At Button Point, Button age debris overlies Sermilik Interglacial organic detritus and Eclipse glaciomarine sediments. At Sermilik Glacier, Button age glaciofluvial deposits overlie Eclipse till and indicate flow from Bylot Island to Eclipse Sound. Most Button glacial ice appears to have advanced onto the South Bylot and Salmon River Lowlands only a short distance from the coast, and may have been floating in the channels at this time (Klassen, 1993). Klassen suggests that the Button event may correlate to the [unnamed] glacial event of eastern Baffin Island, suggested to have occurred between 80 and 40 ka by Andrews *et al.* (1984). The correlation is based on the fact that both this event and the Button event are younger and less extensive than former events.

8) Cape Hatt glaciation, the final foreign event, is considered to be Late Wisconsinan or early Holocene in age, based on a 9.5 ka BP date from shells of an ice-contact delta at Cape Hatt. During this event, ice flowed northward from Baffin Island through Milne Inlet at least to Cape Hatt (Klassen, 1993), but did not reach the study area. This glaciation is considered to be of greater extent than the Cockburn event of Miller and Dyke (1974) and Andrews and Ives (1978), and probably represents the maximum Late Wisconsinan expansion of Foxe ice in the area.

9) The final event observed is native neoglaciation. "Most glaciers on Bylot Island appear to be either stationary at, or retreating from, positions attained during the Neoglacial maximum" (Klassen, 1993, p. 47). Many native glaciers front directly on ancient tundra surfaces. Ujarasukjuik Glacier is an example of such a glacier and has overridden organic debris dated at 130 ± 50 a BP (GSC-5354, collected by

C. Zdanowicz), suggesting that it is currently advancing. Ice-cored, lichen-free moraines suggest a very young age for the native neoglacial event, the maxima possibly being reached 400-500 a ago (Dilabio and Shilts, 1978) and within the last 100 a for some glaciers (Klassen, 1993).

Isostatic rebound (the Inugsuin regression of Andrews *et al.* (1984)) is recorded in the region by marked flights of raised beaches in both lowland areas (plate 2.1). Marine limit in the South Bylot Lowland is 40-45m and 50-60m in the Salmon River Lowland. Uplift is greatest in areas closer to the Foxe Dome source area, where ice was thickest, and least in the northeastern parts of Bylot Island (Klassen, 1993).

One of the aims of this thesis is to further clarify the timing of the foreign and native events identified by Klassen, and possibly provide time constraints for the ice sheets portrayed in figures 2.1 to 2.5. The uncertainties in the age assignments of many of these events point to the need for a more reliable and more applicable dating method. To this end, cosmogenic dating is undertaken, both to date glacial events and to further test this relatively new dating method. Radiocarbon dating and lichenometry are used to help clarify the timing of younger events.

Chapter 3

Methods

Erratic Dispersal Study

A study of the dispersal trends of glacial erratics was undertaken for several reasons. The characterization of glacial events is generally accomplished with the aid of measured orientations of till fabric, glacial striations, and depositional features related to ice flow directions, such as drumlins and eskers. However, there are no till exposures, bedrock outcrops, or landforms suitable for yielding directional information on former ice bodies in the thesis area. Striated bedrock surfaces are rare because the crystalline Precambrian bedrock of the highland regions is susceptible to frost shattering and the Cretaceous-Tertiary sediments of the lowlands are too poorly lithified to display striations. Extremely low annual mean temperatures in the region (Environment Canada, Atmospheric Environment Service, 1982) also suggest that ice was cold-based during full glacial conditions, so the basal sliding that would lead to the production of striae was negligible. Cold-based ice does not generate oriented constructional landforms, either. Consequently, the mapping of erratic dispersal is the only practicable means for distinguishing among glacial events that deposited debris from various source areas. Erratic counts were used to differentiate between areas affected by advances of foreign vs. native ice, and the results were then used to select ideal locations for cosmogenic sampling.

Erratics were classified as belonging to either a native or a foreign assemblage on the basis of their lithologies. The native erratic group consists largely of metamorphic and pegmatitic lithologies of Precambrian age (plates 3.1, 3.2, and 3.3). These crystalline native erratics cannot be distinguished from crystalline erratics of foreign provenance, as they all belong to the same unit of the Canadian Shield. Although some foreign crystalline erratics were no doubt misidentified as native because of this, the error is not considered

to be important because gneisses and pegmatites are not particularly diagnostic ice flow indicators in this region. The Cretaceous-Tertiary sediments (plate 3.4) are so poorly lithified that glacial comminution generally produces sand to clay size debris which cannot be quantified in the field. However, occasional calcareous arkosic concretions, rare examples of lithified Cretaceous-Tertiary bedrock, do occur in the coarser fraction, and these are included in the native erratic category.

Even if only one sample of the distinctive foreign lithology erratics is found at particular site, then foreign ice must have been present at some point in the past. Thus, only the distinctive foreign lithologies can be used as diagnostic foreign ice flow indicators. The foreign erratic class consists of distinctive beige to dark grey limestones and dolomites, as well as red and white quartz arenites (plates 3.5 through 3.8). Although some of the Paleozoic limestone erratics contain macrofossils (plate 3.6), the majority of the fossiliferous limestones display trace fossils (plate 3.5a). The carbonate erratics are particularly easy to identify in the field because their alkaline nature inhibits the growth of lichen and other local vegetation (plate 3.9). A 10 molar HCl acid was used to check carbonate content, but it was found that carbonates could be reliably identified using the "lichen-free" criterion. None of the specular hematite of the Mary River Group or volcanic erratics of the Nauyas Formation reported by Klassen (1993) were found in this study.

In order to establish statistical significance, 300 counts per site were performed, with concentrations of foreign erratics expressed as percentages of the total erratics counted. Only pebbles, cobbles and boulders >2 cm in diameter were counted, in order to reduce operator error in field identification.

Cosmogenic Dating

Introduction

Cosmogenic dating, a technique based on the production of radioactive isotopes in minerals by cosmic radiation, was carried out in an attempt to establish absolute age controls. There are three major advantages of the method over other dating methods: 1) the entire Pleistocene Epoch is within datable range, 2) uncontrollable factors such as climate do not affect the ages determined (Nishiizumi *et al.*, 1993), and 3) large boulders or rock surfaces can be directly dated. No other currently available dating method has all of these advantages.

Description of the Cosmogenic Method

Cosmic rays are charged particles (primarily protons) emitted by the sun and other galactic sources. The bulk of these particles are either deflected by the earth's magnetosphere or are trapped in the Van Allen belts. The 0.1% of cosmic rays that do manage to penetrate the atmosphere react with surficial rocks to generate *in situ* cosmogenic isotopes in mineral lattices (Brown *et al.*, 1991). The impact of a cosmic ray with an atom of a particular element causes a nuclear fission reaction called spallation, which produces new radioactive elements. These are termed radioactive nuclides or radionuclides. To initiate a spallation reaction, a cosmic ray must have an energy of more than 100 MeV (mega electron-volts), about 100 times greater than the energy of cosmic rays emitted by the sun, therefore, cosmogenic radionuclides have only galactic cosmic rays as their progenitors. The buildup of such nuclides in rocks over time can be measured with one of the highly sensitive Tandem Accelerator Mass Spectrometers (TAMS) that have been developed recently. Under ideal conditions, the measured *accumulation* of certain nuclides determines the amount of time a boulder has been exposed to cosmic rays. In this aspect, the process is somewhat different than the

radiocarbon method, which measures radioactive *decay*. Glacial ice cover blocks cosmic ray penetration to the earth's surface, and, if glacial debris is sufficiently abraded during transport or deposition, the cosmogenic clock is reset.

^{10}Be , ^{26}Al , ^{36}Cl and ^{41}Ca are the radioactive nuclides that can be used to date Quaternary events (Nishiizumi *et al.*, 1993). The latter two form by neutron capture, rather than by spallation. This occurs when ^{35}Cl and ^{40}Ca capture a neutron type cosmic ray. ^{26}Al and ^{10}Be , on the other hand, form by spallation, and will be used for determinations in this thesis, for the reasons outlined below. ^{26}Al does not form by neutron capture. ^{10}Be can be formed from ^9Be in this way, but the quantity of ^{10}Be produced by neutron capture is extremely low and requires the initial presence of ^9Be . Similarly, the production of ^{26}Al and ^{10}Be by alpha particle reactions on ^{23}Na and ^7Li respectively, is possible, but negligible (alpha particles are products of U and Th decay). The extremely low production rate is due to the small number of alpha particles with enough energy to initiate the reaction and to their very short range (about $10\mu\text{m}$) (J. Klein, pers. comm.).

The aluminum-beryllium method is ideal for dating Quaternary landscapes, as the half-lives of 1.5 Ma for ^{10}Be , and 0.705 Ma for ^{26}Al (Nishiizumi *et al.*, 1989) give the method a datable range of 2 ka to 7 Ma. The entire Pleistocene and part of the Pliocene Epochs fall within this time frame.

Quartz (SiO_2) contains ^{28}Si and ^{16}O , which become ^{26}Al and ^{10}Be , respectively, via the nuclear spallation reaction (protons and neutrons are released from the original element, giving off isotopes of H and He). A common mineral, quartz is highly resistant to chemical and physical weathering, making it ideal for sampling purposes. It is also fairly easy to separate from other minerals in the laboratory (Kohl and Nishiizumi, 1992), and its low Al concentration allows easy detection of ^{26}Al : ^{27}Al (Middleton and Klein, 1987). Quartz is an efficient closed system, as ^{10}Be and ^{26}Al do not diffuse discernibly in this mineral, and the potential for adsorption of atmospheric ^{10}Be is low, owing to the tight crystal structure of quartz and its resistance to weathering (Kohl and Nishiizumi,

1992). Any atmospheric ^{10}Be that does become adsorbed on quartz crystal surfaces is removed by lab procedures.

An important benefit of measuring both ^{10}Be and ^{26}Al is that extra historical information about the feature sampled can be extracted. The differing half-lives of the two nuclides cause the $^{26}\text{Al}/^{10}\text{Be}$ ratio to decrease with burial. This transpires because earlier exposure signals are lost by radioactive decay when exposure ceases (when burial by more than 3.5 m of ice or by more than 1.2 m of rock or sediment occurs). However, the exposure age can still be traced mathematically. The exposure and burial history of a boulder can be determined using figure 3.1, in the following manner. The upper curve of figure 3.1 represents the production rates for both nuclides assuming no erosional loss (loss of rock surface). The lower curve represents constant erosional loss (Nishiizumi *et al.*, 1993), and the space between the two is termed the "steady state erosion island" (Lal, 1991). If the samples plot near the zero erosion line, there has been little loss of the boulder surface and the age of the sample can be read from the upper curve. If the samples plot near the lower line, constant erosion can be presumed, and a more or less vertical line must be traced upward to find the date of exposure. If, however, the samples plot below the curves, burial or ice cover has occurred, but the date of initial exposure can still be deduced mathematically (Nishiizumi *et al.*, 1993). The use of two isotopes rather than one, therefore supplies valuable historical information about the feature sampled, as well as providing the date of its first exposure.

The $^{26}\text{Al}/^{10}\text{Be}$ cosmogenic technique has diverse applications. It can be used to accomplish the following (Nishiizumi *et al.*, 1993):

- 1) the indirect dating of landforms, as done here
- 2) the direct dating of exposed bedrock surfaces
- 3) the interpretation of burial history, if any
- 4) the constraining of complex exposure and burial histories
- 5) the determination of rates of erosion of bedrock surfaces

Figure 3.1. Plot of $^{26}\text{Al} / ^{10}\text{Be}$ vs. ^{10}Be , explained in the text (after Nishiizumi *et al.*, 1993).

The $^{26}\text{Al}/^{10}\text{Be}$ method has been used successfully to date events in the Sor Rondane Mountains, Antarctica; Meteor Crater, Arizona; the Sierra Nevada Range, California; The Himalayas; Death Valley, California; Punta Caballos, Peru; New Zealand (all described in detail in Nishiizumi *et al.*, 1993) and Pajarito plateau, New Mexico (Albrecht *et al.*, 1993). The Meteor Crater study provides a good example of the past success of the method. The crater was dated at 49 ± 3 ka by thermoluminescence, which was corroborated by $^{26}\text{Al}/^{10}\text{Be}$ dates of 49.2 ± 1.7 ka (Nishiizumi *et al.*, 1991a) and ^{36}Cl dates of 49.7 ± 0.85 ka (Phillips *et al.*, 1991). The large number of successful studies already undertaken using this fairly new technique are testament to its accuracy.

There are several important assumptions inherent in the method:

- ① There is no input or loss of nuclides except by *in situ* production.
- ② The nuclide production rate is known and constant over time.
- ③ Erosion (removal of the surface of the rock) must be negligible, since nuclide production decreases exponentially with depth (usually decreasing to zero at 50-60 cm depth).
- ④ The nuclide production rate depends on altitude, latitude, topography, orientation (how boulder is exposed to cosmic rays), composition, and density of the boulder or rock surface.
- ⑤ There are no cosmogenic nuclides at the start of the current exposure period.

Simply, if the location and composition of a bedrock or boulder surface are known, then cosmogenic nuclide concentration depends on 1) nuclide half-life; 2) geometry of rock exposure; and 3) time elapsed since first exposure (Nishiizumi *et al.*, 1993).

However, some explanation of the assumptions listed above is warranted.

① *Non-in situ production*

Uranium and thorium within the host rock are possible non-cosmogenic sources of low energy alpha particles which may contribute to production of ^{26}Al and ^{10}Be . However, the amount of radionuclides produced by alpha particle reactions is negligible, as outlined earlier. In addition, alpha particles can only penetrate the outermost surface of a quartz crystal, so any effect they may have is removed when the outer crystal surface is dissolved in the laboratory, as part of the general lab procedure for the cosmogenic dating method (Turner, 1993). The laboratory procedures will be described in detail in a later section.

The short half-life of ^{10}Be means that any ^{10}Be present at the time of crystallization of the Precambrian rocks sampled has since decayed completely.

② *Production Rates*

The production rates have been determined to be 6.1 atom/g SiO_2 per year at sea level (geomagnetic latitude $>50^\circ$) for ^{10}Be and 36.8 atom/g SiO_2 per year for ^{26}Al (Nishiizumi *et al.*, 1989), with an uncertainty of $\pm 10\%$ (Lal, 1991). The rates were determined by studying a glacially polished bedrock surface in the Sierra Nevada Range, dated at 11 ka BP based on ^{14}C ages, corrected for changes in $^{14}\text{C}/^{12}\text{C}$ ratio over time (Nishiizumi *et al.*, 1989). The $^{26}\text{Al}:^{10}\text{Be}$ production ratio was found to be about 6 in Si-rich rocks, which is very close to the theoretically expected value (Nishiizumi *et al.*, 1989) and is in line with production rates measured over one year (J. Klein, pers. comm.). Changes in the geomagnetic field do not affect *in situ* ^{10}Be production (Lal and Jull, 1992). The cosmic ray flux varies with latitude and altitude, so corrections must be made to the production rate for each sample. The flux field has been studied extensively and is well known for the entire planet. For a detailed description of production rates and correction calculations, the reader is referred to Nishiizumi *et al.* (1989).

③ Erosion

Removal of layers of the boulder surface complicates the age interpretation by changing the number of reactions at depth as the outer surfaces erode away (Nishiizumi *et al.*, 1993). The use of quartz, which is highly resistant to physical and chemical degradation, minimizes the amount of erosion. For example, Antarctic sandstone samples up to 300 ka in age were found to have extremely low erosion rates, on the order of 10^{-5} cm/a (Nishiizumi *et al.*, 1991b). However, eolian abrasion and exfoliation could be erosional factors in the thesis area. As mentioned earlier, the use of more than one nuclide for dating constrains the interpretation significantly (Nishiizumi *et al.*, 1991b). If the ^{26}Al and ^{10}Be dates do not match, and the dates plot within the steady state erosion island of figure 3.1, some erosion of the sample has taken place.

④ Variables

The *in situ* production rate of the radionuclides is dependent on several factors. These are taken into account in the following equation, where the concentration of the nuclide (N) in a sample increases with exposure time (t) (after Lal, 1991):

$$N = \frac{Pe^{-\mu x}}{\lambda + \mu\epsilon} (1 - e^{-(\lambda + \mu\epsilon)t}) + N_i \quad (1)$$

where P = *in situ* production rate in quartz, corrected for altitude and latitude of the sample

x = sampled depth below the boulder surface

λ = decay constant

ϵ = boulder surface erosion rate

N_i = initial concentration of radionuclide in the sample (assumed to be zero)

$$\mu = \rho/\Lambda \quad (2)$$

ρ = average density of the sample

Λ = mass attenuation length (about 150 g/cm²)

⑤ *No in-situ cosmogenic nuclides*

The exposure history of a given boulder can never be known with complete certainty, but if the radionuclide ages coincide, then the date probably represents a first exposure. If the sample has no *in situ* cosmogenic nuclides from a previous exposure and if the sample is collected from the uppermost horizontal surface of a boulder ($D=0$ cm), equation (1) can be simplified significantly. Rearranging equation (1) to calculate exposure time gives:

$$t = -1/\lambda' \ln(1-\lambda'N/P) \quad (3)$$

If significant erosional effects are to be included, λ' can be defined as $\lambda + \mu\epsilon$, as above.

Description of sample collection

Quartz-rich boulders were targeted during sampling to assure that enough quartz would be collected for analysis. The majority of rocks sampled were Precambrian gneisses with quartz contents of 20 to 45%.

To simplify calculations, as outlined above, and to minimize errors, samples must be collected with utmost care. A rock that has rolled over in any way since deposition will not give a maximum age. The boulders chosen for sampling were large, to minimize the likelihood of overturning due to cryoturbation or rolling, and a few still had till on their upper surfaces, indicating no overturning at all. On boulders where till was present, it formed a thin veneer or consisted of several small boulders on top of the large boulder sampled. Till or stacked boulders on boulder surfaces is commonly observed in areas where debris is melting out of modern glaciers, an artifact of the melting out of debris entrained at different stratigraphic levels within the ice. As the ice melts, till or boulders located stratigraphically above a large boulder are gently let down on top of the large boulder's surface. Thus, when till or stacked boulders are observed on the upper surface

of a large glacial erratic it is obvious that the boulder has remained undisturbed since deposition. To minimize burial effects, such boulders were sampled on the highest exposed surface, where no till was present.

Weathering of associated carbonate boulders (not sampled) were an indication of both long exposure of the sampled feature and stable substrate. For example, a 30 cm diameter carbonate erratic found half buried in morainic deposits near Sermilik Glacier showed evidence of prolonged weathering where exposed, yet so little weathering had affected the buried portion that striations were still visible (plate 3.10 and 3.11). Other smaller limestone erratics showed similar weathering patterns. A 14 cm long clast of micrite was highly solution pitted on the surface, while the buried portion was so well preserved that tiny ichnofossil borings were still visible. The lack of movement of such a small clast is further evidence of the stability of the surfaces sampled. Such samples were taken to be excellent evidence of the long-term stability of the sampled surfaces, and hence, the boulders resting on the surfaces.

The boulders chosen showed no signs of frost shattering, exfoliation or *grus* development and there was no evidence of differential weathering, in spite of the large amounts of soft minerals such as biotite in the sampled gneisses. There is a possibility that eolian activity has caused some erosion, but one would expect differential weathering if this was the case. Rapid colonization by vegetation after deglaciation may have minimized the effects of eolian erosion by acting as a sand and silt trap, as outlined earlier. Examples of sampled boulders are shown in plates 3.12 through 3.14.

Lab Procedure

To remove atmospheric ^{10}Be and to separate out pure quartz, precise lab procedures must be followed. The rock sample is ground and sieved to a uniform grain size, boiled in HCl , HNO_3 and HF , and is treated with successive leaches of dilute HF mixture in a heated ultrasonic bath (Kohl and Nishiizumi, 1992). *In situ* cosmogenic ^{10}Be and ^{26}Al are

not removed by these procedures. The laboratory process generates a pure core of each quartz crystal, (about 30% of the original) that is free of atmospheric contamination. 10 to 200g of this purified quartz are required for dating by tandem accelerator mass spectrometry (TAMS) (Kohl and Nishiizumi, 1992).

Total ^{27}Al is determined by induced coupled plasma (ICP) mass spectrometry and direct coupled plasma atomic emission spectrometry. Be and Al are extracted by anion and cation exchange and by Be- and Al-hydroxide precipitation. Be and Al oxides are then placed in the Cs sputter ion source of the TAMS at the University of Pennsylvania, in order to determine $^{10}\text{Be}/^9\text{Be}$ and $^{26}\text{Al}/^{27}\text{Al}$ concentrations, from which ^{10}Be and ^{26}Al are calculated in atoms/g. The results are then plotted on figure 3.1.

Lichenometry

Lichenometry is used in this study as a relative dating tool for the neoglacial period. This study is the first to apply the method in the northwest Baffin and Bylot Island region, so a calibrated lichenometric time curve for the area was not available. Surfaces of known age, such as archeological sites, are often used to construct time curves, but the Thule and Dorset culture evidence found in the region to date has always been buried (G. Mary-Rousselière, pers. comm.). In the hamlet of Pond Inlet, gravestones that might have provided age associations for lichen growth have recently been replaced by wood headstones. Sources of lichen of known age thus were not available for the construction of a calibrated time curve. Measuring the rate of lichen growth over a number of years was equally impossible, due to the time constraints of M.Sc. thesis work.

Lichenometry rests on the assumption that lichen grow at a fixed rate which varies with climate, such that the size of the largest individual must be related to the age of the feature on which it is growing (Benedict, 1967, Beschel, 1973, Locke *et al.*, 1979). The time required for lichen to initially colonize a freshly exposed surface has not been studied

in the thesis area, but 50 to 75 years are required for colonization in Svalbard (Werner *et al.*, 1987). Lichen life span is about 9500 years on southern Baffin Island, which is a warmer, wetter region (Miller and Andrews, 1972). Complete lichen cover occurs on moraines of Cockburn age (8000-9000 BP) on adjacent Brodeur Peninsula (Dyke, 1993, pers. comm.).

The *Rhizocarpon* group of species used for lichenometry has been referred to as *Rhizocarpon geographicum sensu lato* or *Rhizocarpon geographicum* agg. by many workers, but Innes (1985) has refined the terminology to fit the botanical nomenclature. The term *Rhizocarpon* section *Rhizocarpon* is now used to refer to the lichen measured by the lichenometric technique, but the new term is essentially equivalent to the terminology of earlier workers. Section level is an intermediate level of classification, between genus and species levels.

Rhizocarpon section *Rhizocarpon* is a slow-growing, common lichen, with a growth rate that can be < 0.4 mm/a (diameter increase) in Arctic environments (Innes, 1985). The largest individual of section *Rhizocarpon* on a given glacial feature is used to determine the relative age of substrate stabilization. The age is expressed in terms of the lichen diameter (i.e., the 20 mm moraine). The largest lichen is assumed to be the oldest and to have the optimum growth rate for the feature on which it is growing. Although the largest lichen may not, in fact, be the oldest, the large number of successful studies to date imply that this is a reasonable assumption.

The base, or thalli, of *Rhizocarpon* section *Rhizocarpon* is black, with bright yellow spots called areoles projecting upward from the thalli surface (plate 3.15); identification to section level is possible in the field. The *Rhizocarpon* section *Rhizocarpon* lichen measured in the thesis area had black thalli with bright yellow, dispersed areoles. The species displaying yellow-green or greenish areoles that can be found in more southerly latitudes were not seen. A similar lichen with lighter yellow areoles and black epihymenia was not measured. Epihymenia are similar to areoles to the naked eye, and are often

paired with one areole. This similar species group may belong to *Rhizocarpon* section *superficiale* (ssp. *boreale* Runem.?) (Poelt, 1988). The bright yellow colour of the lichen sampled rules out the possibility that they belong to *Rhizocarpon* section *Alpicola* or *Rhizocarpon* section *Viridiatrum* (Poelt, 1988). For these reasons, it is believed that only *Rhizocarpon* section *Rhizocarpon* was measured.

A number of possibilities for errors in measurement or interpretation are discussed by Benedict (1967) and Beschel (1973). Their studies identified the following criteria that must be met when sampling in the field:

1) Rocks with smooth surfaces should be avoided because they are more difficult for lichen to colonize. Most of the glacially deposited rocks in the thesis area have the same somewhat smooth but not polished texture, making such variation minimal.

2) Lichen are acidophilic. Lichen were measured only on acidic Precambrian gneisses, and never on alkaline rocks such as carbonate erratics.

3) Lichen are resistant to abrasive damage by eolian activity, so windswept sites can be used.

4) Because shading may be a problem on north-facing slopes, these were generally avoided.

5) When overturned and deprived of light, lichen quickly die. Therefore, a lichen "age" represents the time elapsed since the stabilization of a surface (i.e., ice-cored moraines are unstable surfaces and measurements must be made with care). Stacked boulders and boulders with till caps were again good evidence of substrate stability in some areas.

6) Because moisture availability is an important limiting factor in growth rate, wet areas were generally avoided.

7) Boulder surfaces must become snow-free in summer in order for the algae component of lichen to photosynthesize, allowing the lichen to grow. Nevertheless, lichen are quite resilient and can remain dormant if covered by snow for many years. They are

able to recover and continue their growth process if exposed in time (Koerner, 1980). All surfaces measured were snow-free at some point in July and early August, although their past history cannot be known with certainty.

8) Haines-Young (1983) discovered that the base of proximal moraine slopes is a "green zone" where lichen seem to grow better. In this study, only boulders resting on moraine crests or on flat, boulder-strewn surfaces such as valley floors, were measured.

Anomalously large lichen should be avoided, as they can represent unusual growth circumstances; for instance, a lichen may have grown on a supraglacial boulder exposed for a much longer period of time than the morainic boulders with which it is associated. Distinguishing such lichen is not possible in the field, so a statistical method for identifying anomalies was chosen instead. It was decided that 50 measurements would be statistically significant, but because it was the largest lichen that was to be approximated, 50 of only the largest size class of lichen on a given feature were measured. The measurements were taken over a large area at each site, with the mean of the largest five thalli measured used to approximate the size of the largest lichen in the population. If a lichen was found to be more than 20% larger than the next closest lichen diameter, then it was discarded as anomalous, following the technique of Calkin and Ellis (1980). The statistical method chosen for this study is in accordance with Innes (1984), who stated that for a large sample area, the mean of the five largest thalli gives a close estimate of the single largest thallus. Innes (1985) further suggests that the five largest "acceptable" (not anomalous) thalli of an entire substrate be used, and a qualitative decision made as to whether or not to exclude thalli which may be anomalous. The >20% anomaly cutoff described here makes the exclusion of anomalies somewhat less qualitative than a subjective elimination.

An area approximately 30 m in radius (2800 m²) was visually estimated and searched until 30 to 50 measurements of the largest thalli were recorded. This method was adopted in the field when it was noted that searching a large area would be necessary in order to locate 50 of the largest lichen. 50 measurements were made when conditions were

optimal, but inclement weather and paucity of large boulders with measurable lichen sometimes limited sample size. The accuracy of the data is therefore variable.

The largest lichen were defined with respect to the other lichen at each sample site. From the center of the sampling area, large lichen were easily identified and their approximate size mentally noted as a target. Measurement of 50 lichen in this upper size class then ensued over the entire sample area. Although searching an entire feature would certainly be preferable, time constraints were a problem in the thesis area, as much of the work was carried out by helicopter.

Lichen diameter was measured across the largest inscribable circle that could be imagined to fit within the lichen, which is effectively the smallest diameter of a nearly circular lichen (after Locke *et al.*, 1979). Lichen were measured to the nearest millimeter with a flexible plastic measuring tape, which facilitated measurement on irregular rock surfaces. Very elongate and coalescent lichens were ignored. Lichenometry sites were dictated by the presence of exposed boulders, as sedge and moss often covered the features.

No lichen covered rocks were found melting out of glaciers, and the supraglacial load of a medial moraine near the snout of Aktineq Glacier was completely lichen-free (plate 3.16). It is possible, however, that lichen do occur in the supraglacial load elsewhere.

Radiocarbon Dating

Organic samples for radiocarbon dating, taken from delta, river terrace and raised beach sections, consisted of dark brown plant debris occurring as interbeds or disseminated within the sediments. Some marine shell samples were retrieved from raised marine deltas, but were not dated because the samples were not found in growth position

and so may not indicate the true age of the feature. One sample consisted of dead lichen scraped from several rock surfaces with a knife.

The shell and organic debris samples were found in naturally exposed sections. Surface material was removed to a depth of 10 cm with a sampling shovel, and the samples were transferred to plastic bags for shipment.

The organic detritus samples were wet-sieved to separate the organic component from the enclosing sediments in order to avoid problems of contamination by old carbon from Cretaceous-Tertiary coal lenses. Individual plants or rootlets were carefully picked out of the sieved component for AMS dating. Two samples were oven-dried overnight in a 105° oven and converted to CO₂ gas at the University of Ottawa. The gas was then dated at IsoTrace Laboratories, an AMS facility at the University of Toronto. The rest of the organic samples were air dried and sent whole to IsoTrace.

Chapter 4

Results

Erratic Dispersal

A survey of erratic dispersal was carried out over much of the thesis area, as suitable sample sites were plentiful in the extensively boulder-strewn landscape. Erratics were categorized by provenance as foreign (from distant sources and continental ice sheets) or native (from local sources and glaciers), as outlined earlier.

Boulder count locations and sample numbers are shown in figures 4.1 and 4.2. 300 counts were done at each site, and the results are given in table 4.1. Results are shown as the total percentage of foreign lithologies, rounded to the nearest 0.1%. Carbonate lithologies make up the majority of foreign erratics in 90% of the 60 sample sites.

Three distinct erratic dispersal zones can be identified from the data presented in table 4.1: Zone A), a completely native zone of 0.0% foreign debris; Zone B), a zone of minor foreign influx containing 0.1 to 7.0% foreign debris; and Zone C), a major foreign influx zone comprised of 17.0 to 30.0% foreign clasts. The distinct dispersal zones appear to indicate that three separate glacial events are represented in the study area. Using the three-fold zonation, the results are plotted schematically in figures 4.3 and 4.4.

Sample sites 92-BC-13, 92-BC-14, and 92-BC-15, which are located below marine limit, represent a wave-reworked environment, so they are excluded from figure 4.3.

Cosmogenic Dates

Cosmogenic sample numbers and sites are listed in figures 4.5 and 4.6.

Eight of these samples were selected for cosmogenic dating, based on the potential of each for differentiating the three dispersal zones. Of these, five yielded the required

Figure 4.1. Locations of boulder count sites for erratic dispersal study, South Bylot Lowlands

Figure 4.2. Locations of boulder count sites for erratic dispersal study, Salmon River Lowlands.

ERRATIC DISPERSAL COUNTS

SAMPLE SITE	% FOREIGN	SAMPLE SITE	% FOREIGN	SAMPLE SITE	% FOREIGN
92-BC-01	0.7	93-BC-01	0.0	93-BC-24	0.0
92-BC-02	0.0	93-BC-02	6.3	93-BC-25	0.0
92-BC-03	0.0	93-BC-03	0.3	93-BC-26	0.0
92-BC-04	0.0	93-BC-04	0.0	93-BC-27	0.0
92-BC-05	0.7	93-BC-05	5.3	93-BC-28	0.0
92-BC-06	3.7	93-BC-06	4.3	93-BC-29	1.3
92-BC-07	0.7	93-BC-07	6.7	93-BC-30	0.0
92-BC-08	0.0	93-BC-08	0.0	93-BC-31	21.7
92-BC-09	0.7	93-BC-09	1.3	93-BC-32	0.0
92-BC-10	0.0	93-BC-11	6.3	93-BC-33	0.0
92-BC-11	0.3	93-BC-12	1.0	93-BC-34	17.3
92-BC-12	0.0	93-BC-13	18.3	93-BC-35	19.3
92-BC-13	3.7	93-BC-14	0.0	93-BC-36	2.7
92-BC-14	3.0	93-BC-15	0.0	93-BC-37	0.0
92-BC-15	13.6	93-BC-16	0.0	93-BC-38	6.3
92-BC-16	0.0	93-BC-17	0.0	93-BC-39	1.7
92-BC-17	0.0	93-BC-18	0.0		
92-BC-18	0.0	93-BC-19	0.3		
92-BC-19	0.0	93-BC-20	0.0		
92-BC-20	0.0	93-BC-21	25.0		
92-BC-21	0.0	93-BC-22	28.7		
92-BC-22	0.0	93-BC-23	0.3		

Table 4.1. Results of erratic dispersal counts for entire study area, shown as percent foreign lithologies.

Figure 4.3. Erratic dispersal, shown schematically, for the South Bylot Lowlands. Small, mid-size and large circles represent Zones A, B and C, respectively.

Figure 4.4. Erratic dispersal, shown schematically, for the Salmon River Lowlands. Small and mid-size circles represent Zones A and B, respectively.

Figure 4.5. Locations of sample sites for cosmogenic dating, South Bylot Lowlands.

Figure 4.6. Locations of sample sites for cosmogenic dating, Salmon River Lowlands

amount of quartz for analysis by TAMS. Laboratory measurement data for the five samples are listed in Appendix 1. The data cannot be plotted on figure 3.1, as the ^{26}Al results are not available at this time¹. However, the five ^{10}Be results are shown in table 4.2 and are plotted in figures 4.7 and 4.8. The errors shown in table 4.2 include analytical measurement errors, as well as possible errors due to unknown erosion rates and uncertainties in the production rate of ^{10}Be . A significant amount of erosion would increase the age of the samples, while possible spikes in cosmic ray flux over the past 60 ka would decrease the age. The errors given are the best estimates available at present, but the cosmic ray flux error may change as the cosmogenic method becomes better defined by dating of surfaces of known age.

The three erratic dispersal zones yield closely related exposure ages, most notably on the South Bylot Lowlands: the 59.8 ka date is found in erratic dispersal Zone A; the 58.5 ka date falls within Zone B; and the 60.0 ka figure represents Zone C. Similarly, the 62.9 ka date represents Zone B on the Salmon River Lowland, while the 52.1 ka date falls within Zone A. The data appear to indicate that the three dispersal zones do not represent three separate glacial events. Rather, the clustering of the cosmogenic dates around 60 ka BP suggests that the native and foreign glacial events represented in the study area were nearly coeval. Although they may not be exactly synchronous, the exposure ages all fall within the range of error of the method. 60 ka BP is taken to be a minimum estimate for the retreat of foreign and native ice, as erosion may have occurred on the boulder surfaces sampled. Although erosion appears to be minimal in the study area, the ^{26}Al ages are required to further clarify the exposure/erosion/burial history, as outlined earlier.

¹The ^{26}Al results are now available, and are summarized in Appendix 2.

Sample Number	Sample Description	University of Pennsylvania Laboratory Number	Age (ka BP)
AC-93-01	2m X 3m quartzofeldspathic gneiss 118 m asl	AC 93 001	62.9 $\pm$ 10
AC-93-03	1m X 1m biotite gneiss 60 m asl	AC 93 003	58.5 $\pm$ 10
AC-93-06	4m tall biotite gneiss, no weathering, till and grass on part of surface 340 m asl	AC 93 006	59.8 $\pm$ 10
AC-93-07	1m X 1.5 m biotite gneiss, hummocky vegetation in lee 130 m asl	AC 93 007	60.0 $\pm$ 10
AC-93-09	2m tall quartzofeldspathic gneiss, weathering minor 80 m asl	AC 93 009	52.1 $\pm$ 9

Table 4.2. Details of cosmogenic sample sites and ^{10}Be dates.

Figure 4.7. Cosmogenic dates (ka) for the South Bylot Lowlands. Analytical errors for these dates are shown in table 4.2.

Figure 4.8. Cosmogenic dates (ka) for the Selmon River Lowlands. Analytical errors for these dates are shown in table 4.2.

Lichenometry

The majority of the surfaces in the thesis area are too ancient for lichenometric techniques to be applied. Tundra is commonly well developed, and lichen cover is generally complete on boulder surfaces, indicating that most parts of the study area are of greater than Holocene age. Lichenometry could only be applied in sectors close to modern glaciers, as shown by the sample locations and numbers plotted in figures 4.9 and 4.10.

Mean thalli sizes, which represent a statistical estimate of the largest lichen at each site, and individual lichen measurements are given in table 4.3. The means are calculated from the five largest lichen at each site, which are circled on the table. Anomalously large lichen, not included in the calculation of the means, are shown in squares. The measurements at sample sites 92-BL-04 and 93-BL-08 appear anomalously large when compared to nearby sample sites. These two locations were wetter than other areas, and are included only for comparison to normal sample sites. The effect of moisture on lichen growth rate is obviously quite significant.

Near Aktineq Glacier, a detailed survey of morainic benches was carried out in the 1992 field season. These subdued features, which occur outside the modern moraines of the glacier at different elevations on the side of Etajaguaq Mountain, are visible in air photos and appear to record several former positions of Aktineq Glacier (plate 4.1). Four lichen sites were recorded on the lowermost (ice-proximal) moraine and three on the uppermost (distal) moraine. The moraines will be referred to hereafter as the lower and upper moraines. Another moraine between the lower and upper moraines was not sampled, as it was often difficult to distinguish in the field. The mean of the largest five lichen was calculated from all available data for each moraine, rather than from the means for each lichen site. The accuracy of the statistical estimate of the largest lichen is consequently much better for these features, because the whole moraine was searched, and the largest five lichen from the entire feature were used to calculate the mean. The

Figure 4.9. Locations of lichenometry sample sites, South Bylot Lowlands.

Figure 4.10. Locations of lichenometry sample sites, Salmon River Lowlands.

anomalous site 92-BL-04 is not included in the determination of the mean for the lower moraine. The total means of the lower and upper moraines of Aktineq Glacier are listed in table 4.3, with brackets showing which sites are included in the calculation of these means.

Estimates of total percent lichen cover on boulders (which includes all lichen species) were made at some sites. Sites too old for lichenometry generally show 90-100% lichen cover, 49-59 mm sites show 40-60% cover, and recent features vary from 0 to 5%. None of the lichen found on recent features belong to the target lichen group, except at "Stagnation" Glacier.

The lichen ages (means of the largest five lichen) are plotted in figures 4.11 and 4.12. 9 of the 19 samples fall within the 49-52 mm range, and 5 fall within the 55-59 mm range, suggesting two closely spaced correlative events (figure 4.13).

The 55-59 mm age group is the oldest neoglacial event recorded, and correlations among features of this age can be made on the South Bylot Lowlands. The upper moraine of Aktineq Glacier, the older moraine near "Stagnation" Glacier, and a few boulder-strewn hilltops near Ujarasukjuik Glacier all fit into this age category.

Ice must have receded shortly after this period, because a second group of features are observed in the 49-52 mm size group. These include the rest of the boulder-strewn hilltops near Ujarasukjuik Glacier, the older part of "Dead Lichen" Valley, the lower moraine of Aktineq Glacier, the raised beach deposit near the mouth of Aktineq Creek, and the oldest of the terminal moraines in the highland area that is just southeast of the Salmon River Lowlands.

Lichen-free zones are defined here as those that are free of *Rhizocarpon* section *Rhizocarpon* at any stage of growth, unlike the "lichen-free" zones of Locke and Locke (1977), which are defined as low in total lichen quantity relative to other substrates. A "lichen kill zone" was found, however, consisting of both dead and live lichen, in "Dead Lichen" Valley on Bylot Island. The dead lichen of sample site 92-BL-17 were black and grey, yet were still identifiable as *Rhizocarpon* section *Rhizocarpon*. They had not been

Figure 4.11. Lichen ages (in mm) for each sample site, South Bylot Lowlands.

Figure 4.12. Lichen ages (in mm) for each sample site, Salmon River Lowlands.

Lichenometry Data *Largest Lichen*

Figure 4.13. Frequency histogram of approximated largest lichen size. Most lichen fall within the 49-52 mm range, but there is a secondary grouping of 55-59 mm.

scraped off the sample area boulders in any way by ice, so they must have been subjected to extensive, continuous snow cover. The live lichen growing on the same boulders as the dead ones measured 31 mm, so this locality has a much lower lichen age than the surrounding area, which fits into the two categories described above (sites 92-BL-13 to 92-BL-18, figure 4.11). A lichen surface in the same valley showed more complete coverage and no dead lichen, which was the usual circumstance encountered during sampling. The lichen age of this site (92-BL-18) is 51 mm; it stands out in sharp contrast to the lichen kill zone (plate 4.2). The younger relative age for the lichen kill zone suggests that a snow field stayed in the area long enough to kill the local lichen, with recolonization of the surface occurring after disappearance of the snowpack. Had an ice body been responsible for the younger surface, the dead lichen would have been scraped off the boulders.

Only one glacier was found to have lichen on its most recent neoglacial moraine (hereafter referred to as modern moraines). The large, ice-cored moraine of "Stagnation" Glacier, with a lichen age of 10 mm, appears light grey against the older 59 mm moraine just beyond it (plate 4.3). This glacier must have begun its recent retreat at some time prior to the onset of retreat for nearby glaciers, which also have modern recessional moraines.

The modern moraines of most native glaciers are considered to be lichen-free, in that they exhibit no lichen of the target species group at all (plates 4.4 through 4.6). The lack of distinguishable *Rhizocarpon* section *Rhizocarpon* suggests that recession of native ice from such moraines has been a recent and rapid event, possibly having started within the last century, as the initial colonization period obviously has not yet passed. The lack of lichen on these modern moraines indicates that lichen pre-dating the age of a moraine (anomalous lichen) are extremely rare.

Though none of the mean lichen sizes are larger than 59 mm, *Rhizocarpon* of much greater size have been measured elsewhere. At Grothreen Glacier, Norway, for example,

lichen of up to 360 mm were measured on neoglacial moraines (Matthews and Shakesby, 1984). However, an upper size limit of about 80 mm is common to many parts of Baffin Island, and is thought to be due to a combination of factors, such as lichen competition and the rate of weathering of boulders on which the lichen grow (Andrews and Barnett, 1979). However, since the lichen cover on the upper size class of 59 mm in this study is about 60%, competition is not likely to be a limiting factor, nor is the negligible amount of weathering that is observed in the area. If these factors do not apply here, the 59 mm limit must be related to a climatic event. The lack of a gradual transition from the 59 mm surfaces to the ancient surfaces with complete lichen cover suggests that a large time period elapsed between the deposition of the older drift and that of the neoglacial moraines, an interpretation consistent with the cosmogenic results.

Recessional moraines near the southwest portion of the Barnes Ice Cap, which is located about 300 km southeast of the study area, have been dated using a lichenometry growth curve which was calibrated with radiocarbon dates (Andrews and Barnett, 1979). The age of a moraine with a maximum *Rhizocarpon* diameter of 53 mm is estimated to be approximately 1500 a BP; a nearby 30 mm moraine is thought to be about 750 a BP (Andrews and Barnett, 1979). Although the Barnes Ice Cap lichenometry dates cannot be extrapolated directly to the thesis area, they do give an idea of the order of magnitude of time ranges that the lichen in the study area may represent.

Radiocarbon Dates

Samples of organic detritus were collected from several sites for radiocarbon dating. Such sites are, however, quite rare in this Arctic field area, as shown by the limited number of sample sites and numbers in figures 4.14 and 4.15. Only the most ideal samples were chosen for dating at IsoTrace Laboratories in Toronto, and the results are given in table 4.4. The age data are plotted in figures 4.16 and 4.17.

Figure 4.14. Location of radiocarbon sample sites, South Bylot Lowlands.

Figure 4.15. Location of radiocarbon sample sites, Salmon River Lowlands.

Sample Number	Sample Description	Isotrace Laboratory Number	Age (a BP)
93-RD-03	<i>Salix?</i> roots	TO-4181	13 800 $\pm$ 300
93-RD-05	<i>R. section Rhizocarpon</i> lichen	TO-4180	modern
93-RD-07	moss branches with leaves	TO-4179	7260 $\pm$ 80
92-RD-02	<i>Cassiope tetragona</i>	TO-3781	80 $\pm$ 50
92-RD-03	woody plant	TO-3782	1320 $\pm$ 50

Table 4.4. Details of radiocarbon samples and dates.

Figure 4.16. Radiocarbon dates, South Bylot Lowlands. Analytical errors for these dates are shown in table 4.4.

Figure 4.17. Radiocarbon dates, Salmon River Lowlands. Analytical error for this date is shown in table 4.4.

Sample 93-RD-03 (*Salix?* roots, $13\,800 \pm 300$ a BP) was collected from a natural exposure at 37 m asl on the distal slope of an ice-contact glaciomarine delta. Part of the town of Pond Inlet is located on the surface of this feature. The sample was taken from the lowest of three laterally continuous, conformable, organic debris bands overlying a surface displaced by a normal fault (plate 4.7). The sample did not appear to be contaminated by overlying roots. There is an unconformity in the sequence, where the normal fault is truncated by the organic laminations. It appears that the faulted surface was exposed and eroded, allowing the *in situ* growth of vegetation, represented by the organic beds. The sequence is overlain by 10-12 m of plane parallel and crossbedded sands, in which there are no other organic units exposed. The sample site may be affected by slumping, although geomorphological evidence for slumping was lacking at the site.

93-RD-07 is a moss sample collected by W.W. Shilts from beneath 50 cm of (eolian?) sand, in a section draped by lichen-free boulders of the outermost modern moraine of Aktineq Glacier. The age of 7260 ± 80 a BP agrees well with samples of similar age from nearby sections (Dilabio and Shilts, 1979, Klassen, 1993). This suggests both that the glacier had completed its retreat from Eclipse Sound a minimum of 7200 years ago, and that the modern moraines represent the farthest point of advance of Aktineq Glacier in at least the last 7200 years.

Sample 92-RD-03 (1320 ± 50 a BP) is from an organic-rich gravelly sand unit underlying gravelly sand, which was exposed near the outwash plain of Aktineq Glacier. The sample may have been contaminated by modern roots, but this did not seem to be the case at the sample site, as the overlying grass roots were fine, while the organic debris within the sediment was fairly coarse (1-3 mm thick branch and root fragments). The date documents a period of Holocene rebound, since it is part of a raised beach unit below marine limit, at an elevation of 35 m.

Sample 92-RD-02 was collected from an organic mat 1 m below the upper surface of a sandy abandoned river terrace, which correlates with the raised beach of "Bigroot"

Lake, just downstream of the sample site. The sample was an extremely well preserved branch of *Cassiope tetragona*, a common flowering heath plant found in sheltered parts of the lowlands (identified by A.W. Dugal, National Herbarium of Canada). The radiocarbon date of 80 ± 50 a BP implies that the ice-dammed lake decreased in volume very recently (plate 4.8). However, air photos from 1961 and 1982 show that the lake, at its maximum level in 1961, had drained to its current level by 1982. Since the radiocarbon date is bracketed between 1820 and 1920 A.D., the plant must have grown in place during a previous drainage event. The 1 m of overlying sediment may have been deposited during the later lake highstand shown in the 1961 air photo. The lake subsequently drained again, to a lower level than before, sometime between 1961 and 1982.

A similar catastrophic drainage event appears to have occurred in a large southwest-trending lake within the highland portion of the study area on Baffin Island, near lichenometry site 93-BL-08 (the lake with two glaciers flowing at right angles to the long axis of the lake in figure 4.17). A prominent lichen trim line marks the former level of the lake, which is dammed by a glacier at its northeast end, just outside the boundary of the figure (plate 4.9). The lake had already drained when the 1958 air photos were taken, but the glacier flowing from the northwest was calving into the lake at that time. The same glacier is currently almost completely bisecting the lake (plate 4.9).

Sample 93-RD-05 is from "Dead Lichen" Valley. 20 dead lichen were scraped from several rocks in order to pinpoint the timing of the extensive snowdrift event that caused their demise. The modern date implies either that: 1) this event occurred shortly after 1950; or 2) some of the dead lichen were not actually dead, but rather were dormant. Since lichen of 31 mm were found growing alongside the "dead" ones, the event could not have occurred shortly after 1950, because the initial colonization period is unlikely to have passed yet. Consequently, some of the lichen in the sample must have been dormant.

Chapter 5

Interpretation

Erratic Dispersal

A native dispersal zone (Zone A) occurs within a series of subdued moraines that appear to be related to an advance of Ujarasukjuk Glacier. The moraines, which were mapped by air photo interpretation, are shown in front of the glacier in figure 4.3. The dispersal data confirm that these features are recessional moraines of Ujarasukjuk Glacier. A similar native zone is also defined in front of Ujarasukjuik Glacier, where no moraines are observed, indicating that this glacier was at one time as far advanced as its neighbour. Native debris also characterizes a modern medial moraine located on the surface of Aktineq Glacier (92-BC-02), as well as its modern terminal moraine zone (92-BC-03). The 30 m high ice-cored moraine of "Stagnation" Glacier and sites in the vicinity of "Dead Lichen" Valley, which are close to modern ice margins, are also part of Zone A. Dispersal site 93-BC-01 is from a bench at the foot of Unguaq Mountain. The 100% native debris and presence of talus from the nearby mountain suggest that this feature is a talus cone, which appears to have been truncated by longshore current erosion along Eclipse Sound. The native zone also characterizes much of the Salmon River Lowland, suggesting that native ice was previously more extensive than the recessional moraines of the former Utuk Glacier show.

Zone B extends about 20 km inland on the South Bylot Lowlands, and at least 10 km inland on the Salmon River Lowlands. It defines a zone of moderate foreign influx in both regions. The foreign ice which deposited this debris must have been related to a large ice sheet, as it was able to cross the 1000 m deep channels of Eclipse Sound and Navy Board Inlet, and to advance a considerable distance inland. The highest elevation at which foreign erratics were found in this study was 630 m, near the top of Etajaguaq

2 of /de 2

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS-
1963-A
AND REFERENCE MATERIAL 1010A
NBS 1010A-1 (1963-A) (1963-A)

Mountain, where a limestone erratic was observed. Klassen (1993) found foreign quartz arenite erratics as high as 1130 m asl near Sermilik Glacier, which were considered to be related to the Baffin Glaciation. The limestone erratic observed at 630 m is more likely to be related to the Eclipse event.

Zone C is found only on Bylot Island, in areas mostly near the coast, and thus was suspected to represent a separate glacial event.

Most of the foreign zones are dominated by limestones, rather than arenites, in spite of the fact that ice flow from carbonate bedrock sources must have passed over the Proterozoic quartz arenite sources (figure 2.6). This may be due in part to the softer nature of this rock type, which was probably more easily eroded and incorporated into the generally cold-based ice sheet than the more resistant lithologies.

Cosmogenic Dates

Although the dispersal study appeared to show three separate glacial events, cosmogenic dating suggests that this is not the case. The clustering of dates around 60 ka, regardless of dispersal zone, shows that advances of both native and foreign ice occurred within a narrow time frame. The ^{10}Be cosmogenic evidence confirms Klassen's hypothesis that native ice was "dammed" by foreign ice of the Eclipse Glaciation (Klassen, 1993). Thus, at all cosmogenically dated sites, the last glacial phase must have been the Eclipse event. Ice of the later Button event must not have advanced very far inland, or it would have been detected by cosmogenic dating. Cosmogenic samples AC-93-01 and AC-93-03 are located 1 and 2 km from marine limit, respectively, indicating that Button ice did not penetrate even 1 or 2 km inland on either the South Bylot or the Salmon River Lowlands. Events of pre-Eclipse age were not identified among the few samples dated so far.

The erratic dispersal and cosmogenic age data allow a glacial scenario to be suggested for events of Eclipse and younger age. Eclipse ice must have flowed into the

area more than 60 ka ago, crossing the deep channels separating Baffin and Bylot Islands and confining native ice to the mountainous regions. Upon the retreat of foreign ice, the excess native ice volume was redistributed, causing a short-lived advance called the Aktineq glacial phase (Klassen, 1993). Once the excess ice was dissipated by ablation, native ice fronts rapidly retreated from the Aktineq maximum, to or inland of, the positions in which they are found today. The advance of Button ice in later times affected only coastal areas, so ice must have been floating in the channels at this time, as had been suggested by Klassen (1993).

Radiocarbon Dates

The $13\,800 \pm 300$ a BP date from the Pond Inlet delta is problematic. The classic delta morphology and the presence of crossbedded sands indicates that the feature is a glaciomarine delta or subaqueous fan. However, the fact that the crossbeds at the sample site dip roughly northeast, despite a deltaic shape which implies water flow from the southeast, implies that the feature is more complex than simple glaciomarine genesis can explain. In addition, the radiocarbon date of $13\,800 \pm 300$ a BP for sample 93-RD-03 contradicts two other pieces of geological evidence: 1) there was no ice flowing seaward from the Salmon River Lowland in Late Wisconsinan time (from the cosmogenic results) and 2) the 33.8 ka shell date from the upper surface of the feature implies that the delta is much older (Hodgson and Haselton, 1974). Although modern root contamination may explain this anomalously young date, the modern grass roots present at the sample site appear to be much finer than the (*Salix*?) roots found in the organic beds. The sample area does fall within a zone identified by Klassen (1993) as having been affected by Button ice, which was floating in Eclipse Sound. One possible explanation for the young age is that the sample site represents a marginal ice-contact deposit of Button age, accreted onto the distal facies of an Eclipse aged delta. The Button event is thus tentatively proposed to

be at least 13.8 ka in age, but further radiocarbon dates are required to either confirm or disprove this interpretation.

The date of 7260 ± 80 a BP near Aktineq Glacier reinforces similar dates from other samples collected from deeper parts of nearby sections, suggesting that the glacier had retreated to a position within the modern glacial moraines by at least 7260 a BP.

The 1320 ± 50 a BP date from a raised beach near the mouth of Aktineq Creek probably records a period of transition from marine to terrestrial conditions during Holocene uplift.

The drainage of "Bigfoot" Lake to its present level occurred between 1961 and 1982. The 80 ± 50 a BP date may record a previous lake level drop, indicating that "Bigfoot Lake" has undergone more than one drainage event, but these drainage and refilling events are highly site specific and are related to thermal ice-dammed lake phenomena, rather than to climate.

The modern date of the dead lichen sampled in 93-RD-05 implies that the lichen were dormant, rather than dead, leaving the timing of the snow expansion event in "Dead Lichen" Valley unresolved.

Lichenometry

A period of native Holocene retreat appears to be represented by the lichenometry results, since the majority of lichen sample sites are proximal to modern glaciers and lichen diameters tend to decrease toward the ice bodies. Most of the data on Bylot Island fall within the 49-59 mm range, so it appears that these surfaces all became stable at approximately the same time within the Holocene. The grouping into a 49-52 mm and a 55-59 mm group suggests two closely spaced events.

The 39 and 49 mm lichen ages southeast of the Salmon River Lowland suggest that native ice retreated to the southeast in neoglacial times. The ages are from an extensive

series of terminal moraines that also suggest recent ice retreat to the southeast. The native ice body that may have produced these terminal moraines can be seen in the southeast corner of figure 4.12.

Since the younger features (49-52 mm group) occur closer to existing ice near Aktineq and Ujarasukjuit Glaciers, it appears that a Holocene recession has been recorded in these areas. The ice that may have been present at the latter site cannot be due to an overall expansion of Ujarasukjuit Glacier, because the cosmogenic date of 59.8 ka in front of the glacier indicates that it has not advanced beyond its present position for some time. Ice may have been diverted from the glacier down the narrow valley at some time in the past, however, in much the same way as ice is currently diverted from Ujarasukjuit Glacier toward the northwest. In "Dead Lichen" Valley, an ice-capped peak, which probably had more extensive ice cover in the past, can still be seen (plate 4.2).

The lichen age of 52 mm at site 92-BL-11 is from the raised beach area near the mouth of Aktineq Creek, indicating that the 52 mm age represents a feature which formed within the period of Holocene uplift.

Lichenometry evidence points to a period of Holocene retreat from Eclipse Sound for Aktineq Glacier. The lower 52 mm moraine of Aktineq Glacier falls into the 49-52 mm group, while the upper 57 mm moraine (and the 56 mm measurement from the top of Etajaguaq Mountain) fall into the 55-59 mm group. The 31 mm feature (92-BL-01) that occurs close to the modern moraines (more proximal to the glacier than the lower moraine) seems to reinforce the idea that a recession has been recorded. However, the radiocarbon dates of 7260 ± 80 a BP (this study) and 7860 ± 100 a BP (Dilabio and Shilts, 1978), indicate that the glacier had completed its retreat by at least 7860 years ago, so possible Holocene retreat of Aktineq Glacier would be restricted to a period prior to about 7800 a BP. If this timing is correct, the 60 ka cosmogenic exposure ages for other parts of the study area do not apply to Aktineq Glacier. The lichen ages from Etajaguaq Mountain are thus somewhat of an enigma, suggesting that Aktineq

Glacier is the only large glacier in the region to have undergone a major Early Holocene or Late Wisconsinan recession. The vegetation of the area surrounding the glacier is slightly different than that of the older cosmogenically dated areas in that it is much greener than the tundra in front of Ujarasukjulik, Ujarasukjuit and former Utuk Glaciers. This may mean that Aktineq Glacier did undergo an Early Holocene retreat, and that the climax vegetation community has not yet been reached in the surrounding area. However, the vegetation differences could simply be due to differing microclimates. In fact, differing microclimates could be the reason that there are measurable lichen on Etajaguaq Mountain, but it does not explain decreasing lichen ages toward Aktineq Glacier.

Another possible interpretation is that the lower and upper moraines have undergone Holocene downslope movement, but this also does not explain the decrease in lichen sizes observed toward the modern glacier margin. Detailed future investigations in the vicinity of Aktineq Glacier may shed some light on this problem.

The 10 mm age from the moraine crest of "Stagnation" Glacier implies that the glacier has been retreating slightly longer than other glaciers, such as Aktineq and Ujarasukjulik, which have no *Rhizocarpon* at all on their modern moraines. Considering the relative instability of the ice-cored moraine sampled, the 10 mm age is probably a minimum estimate.

Chapter 6

Implications

The cosmogenic age data are the first pre-Late Wisconsinan absolute dates obtained for glacial features in Arctic Canada. Since the exposure ages indicate that Eclipse ice retreated about 60 ka ago, the glacial maximum of the Eclipse event must have occurred prior to this time. A minimum estimate of the Eclipse glacial maximum would thus be early Middle Wisconsinan or late Early Wisconsinan depending on how long ice remained at its maximum (the boundary between the Early and Middle Wisconsinan Substages is placed at 65 ka). It is likely that the maximum occurred in Early Wisconsinan time, so ice of this period must have been related to the Foxe Dome of the Laurentide Ice Sheet, and can thus be termed Early Foxe in age.

No events prior to the Eclipse event were identified in this study. Since the Eclipse event probably achieved full glacial conditions in the Early Wisconsinan, these pre-Eclipse events may have occurred in Sangamonian or earlier time.

The Sermilik Nonglacial Period began with the retreat of foreign and native ice approximately 60 ka ago, and has continued to the present in most of the study area. Many surficial deposits have remained remarkably stable for the entire 60 000 year period.

The Button glaciation appears to have been confined to coastal regions and a tentative minimum age estimate for its maximum is 13.8 ka BP. Thus, the study area may have been affected by Late Wisconsinan (Foxe) Ice, but it was of minor extent, floating in the channels.

Native neoglacial advance and recession also appears to have occurred in some parts of the study area. Recent rapid retreat is evidenced by the lichen-free moraines of several native glaciers.

An important implication of this study is that the ice sheet configuration of Vincent and Prest (1987), shown in figure 2.1, can be assigned an Early Wisconsinan age, at least

in its northeastern sector. Cosmogenic dating in other areas not affected by Late Wisconsinan ice would be a desirable avenue for future research.

The data also support the 70 ka BP estimate for full glacial conditions of the Laurentide Ice Sheet derived from ice core data (Koerner and Fisher, 1985).

Lack of extensive foreign ice in Middle Wisconsinan time favours hypothesis II of the ice sheet configuration proposed by Dredge and Thorleifson (1987) in figure 2.3, which appears to be an accurate representation of foreign and native ice in the northern Baffin Island region at this time. If the entire configuration is accurate, there are further implications for ice activity at the southern margins of the Laurentide Ice Sheet, for example, open water conditions in Hudson Bay. Again, further cosmogenic dating in other areas may prove useful.

Although there may have been Late Wisconsinan ice in the region, its limited extent and tentative age assignment do not warrant its addition to the configuration of Dyke and Prest (1987), shown in figure 2.5.

If the cosmogenic exposure ages are interpreted correctly, native ice has either remained reasonably stable or has advanced slightly over the last 60 ka, a time frame in which Laurentide (Foxe, Eclipse) Ice retreated, readvanced (Button, Cape Hatt, Cockburn) and then completely disappeared. So limited a response to such major climatic fluctuations suggests that these high latitude ice bodies are quite stable, compared to mid-latitude ice sheets.

Stable, high-latitude ice bodies would be in line with the glaciological theory of Boulton *et al.* (1985), which predicts sluggish response to global climate change in the higher latitudes. These authors suggested two possible reasons for the predicted stability: 1) the relatively stable high latitude climate compared to that of the mid-latitudes; and 2) the generally more resistant bedrock in these areas (Boulton *et al.*, 1985). Globally, high latitude glaciers are usually underlain by Precambrian basement and cold-based ice

has not been able to deposit the deformable beds thought to be instrumental in causing instability in mid-latitude ice sheets (Boulton *et al.*, 1985).

The study area differs from other high-latitude locations in one significant respect. Although the source area of native ice in the thesis area is crystalline basement, the friable Cretaceous-Tertiary sandstones of the lowland regions are, in fact, highly deformable. They have been little affected by either native or foreign glaciations, so it seems that the cold-based nature of high Arctic ice sheets is more important than the presence or lack of deformable beds. A reasonable conclusion is that the relative stability of high Arctic climate is solely responsible for the minimal glacial response shown by the Bylot and North Baffin glaciers. The Laurentide Ice Sheet as a whole was affected by mid-latitude climate and possibly by the presence of deformable beds; hence it fluctuated greatly during the Wisconsin Stage. Increased moisture availability to the High Arctic upon the complete disappearance of Laurentide Ice may or may not have caused an advance of native ice, as evidence for ice advance is much harder to find than evidence for its retreat. However, the advance, if there was one, was not very extensive in comparison to native advances in the past (Bylot, Aktineq). Future paleoclimate studies are required to determine whether or not climate in the region has actually been stable for such a long period of time.

On a smaller scale, some glaciers, such as "Stagnation" glacier, apparently have been retreating during the Late Holocene, while others have begun to retreat from their modern neoglacial maxima only very recently (Ujarasukjulik and Aktineq Glaciers). Utuk Glacier has completely disappeared since its maximum in Eclipse time. Given the minor climate variation between glaciers, this response must be glaciological. Net variations in accumulation and ablation between these glaciers may be explained by variations in glacial debris load which may affect albedo, or some other as yet unknown glaciological controls. This is certainly another potential avenue for future research.

The study area also has great potential for future paleoclimate work:

1) The absence of extensive Late Foxe (Wisconsinan) Ice means that the area might have been a Late Wisconsinan plant and animal refuge, assuming that the Button glaciation did not cause polar desert conditions at its periphery.

2) Arctic soils may be better developed in the study area than in areas glaciated during the Late Wisconsinan, however, extensive soil development was not observed during the course of field work for this study.

3) The permafrost record, as well as the micropaleontological and palynological records preserved in lakes that have been exposed for 60 ka, are potentially valuable sources of paleoclimate evidence.

Finally, cosmogenic dating appears to be an excellent tool to apply to other Arctic locations, especially those beyond the periphery of the Late Wisconsinan Ice Sheet, or in other areas where classical dating methods have been unsuccessful in defining the timing of older events.

Chapter 7

Summary and Conclusions

The thesis area, which lies beyond the outer margins of the Late Wisconsinan Laurentide Ice Sheet, has provided a rare window into pre-Late Wisconsinan Arctic history. A study of glacial erratic dispersal identified three distinct zones of glacial influx, one native and two foreign, implying that three separate glacial events were represented in the region. However, dating by ^{10}Be cosmogenic methods showed that, chronologically, only one event was actually represented, as the five dates clustered around 60 ka BP, regardless of the dispersal zone represented. Thus, foreign and native ice appear to have retreated from the study area within a narrow time frame, indicating that Klassen's (1993) earlier interpretation of the relationship between native and foreign ice during the Eclipse Glaciation is correct. The cosmogenic evidence strongly supports Klassen's hypothesis, which stated that Eclipse (foreign) ice advanced into the region, causing native ice to be dammed within the highlands, where a volume excess built up. Upon retreat of Eclipse ice 60 ka ago, native glaciers advanced beyond their normal steady state configurations, but quickly retreated to more or less their modern positions. The Eclipse maximum is likely to have occurred in Early Wisconsinan time, so ice flow would have been related to the Foxe Dome of the Laurentide Ice Sheet, and the event can be assigned to the Early Foxe Glaciation of Baffin Island. Events which pre-date the Eclipse event may be pre-Foxe in age.

The Sermilik nonglacial period began with the retreat of Eclipse ice at the onset of the Middle Wisconsinan Substage and has continued to the present over much of the study area.

The Button Glaciation was not identified by cosmogenic dating, in spite of the proximity of some samples to marine limit, so this event must have been extremely limited in extent, possibly represented in this area by ice floating in the channels. A tentative

minimum age of 13.8 ka BP is suggested for this event, but the age inference is supported by only one radiocarbon date.

Native ice retreat appears to be represented in areas proximal to modern native ice bodies, as shown by lichenometry. Two periods of retreat are indicated by lichen size groupings of 49-52 mm and 55-59 mm, but an absolute age cannot be assigned to either event. Another very recent retreat is shown by several glaciers that have lichen-free modern moraines.

These results have several important implications:

- The "maximum Post-Sangamonian *sensu stricto*" configuration of the Laurentide Ice Sheet of Vincent and Prest (1987), shown in figure 2.1, is Early Wisconsinan in age, at least at its northeastern boundary.
- Hypothesis II of the Middle Wisconsinan Laurentide Ice configuration of Dredge and Thorleifson (1987), shown in figure 2.3, appears to show a reasonable configuration for the north Baffin area.
- The Late Wisconsinan configuration of Dyke and Prest (1987), shown in figure 2.5, is also accurate for the study area. If Late Wisconsinan ice was present, it was restricted to the channels separating Bylot and North Baffin Islands, and as such, does not merit incorporation into the configuration of the main ice sheet.
- Large parts of the study area have remained ice-free since the beginning of Middle Wisconsinan time. The area may thus contain advanced soil profiles, climax vegetation communities and lengthy paleoclimate signals, so there is much potential for future work.

- The lack of significant native ice advance from Middle Wisconsinan time to the present suggests that high Arctic climate has been fairly stable during this time period. Future paleoclimate work is required to verify this hypothesis, however.

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Plate 1.1 Typical prostrate tundra vegetation, Salmon River Lowland.

Plate 1.2 Tundra vegetation, South Bylot Lowland.

Plate 1.3

Tundra mats formed in the lee of glacial erratics, Etajaguaq Mountain.

Plate 1.4

Lag deposit on Etajaguaq Mountain, visible in foreground. The erratic-strewn mountain peak is visible in the distance.

Plate 1.5

Close up view of lag deposit, Etajaguaq Mountain.

Plate 2.1 Raised beaches, coastal South Bylot Lowland.

Plate 2.2 Incised meltwater channels, east of Aktineq Glacier. Paleoflow direction was from upper right to lower left. Aktineq Glacier, flowing from right to left, can be seen in the background.

Plate 3.1

Native erratics: chloritized augen gneiss

Plate 3.2

Native erratics: a) garnet - biotite gneiss
b) quartzofeldspathic gneiss

Plate 3.3

Native erratics: pegmatites

Plate 3.4

Native erratics: a) unlithified arkosic arenite

b) poorly lithified siltstone and arkosic wacke

Sample b) has been impregnated with resin to preserve its features

Plate 3.5

Foreign erratics: a) micrite with burrows (fossiliferous micrite)
b) micrite

Plate 3.6

Foreign erratics: intraclastic biomicrudite with marine pelecypod
macrofossils

Plate 3.7 Foreign erratics: fine to medium-grained red quartz arenites

Plate 3.8 Foreign erratics: fine-grained clean quartz arenites

Plate 3.9

A conspicuous limestone erratic standing out against the surrounding tundra due to the lack of vegetation on its surface.

3.9

Plate 3.10 Highly weathered exposed surface of a limestone erratic.

Plate 3.11 Buried surface of the same erratic. The buried portion is so well preserved that glacial striations are still visible.

3.10

3.11

Plate 3.12 Sampling the upper surface of AC-93-01, Salmon River Lowland,
for cosmogenic dating.

Plate 3.13 Sampling of AC-93-05, South Bylot Lowland.

Plate 3.14

Sampling of AC-93-09, Salmon River Lowland.

Plate 3.15

Rhizocarpon section *Rhizocarpon* lichen

Plate 3.16

Lichen-free supraglacial debris load of a medial moraine,
Aktineq Glacier.

Plate 4.1

Morainic benches on Etajaguaq Mountain, visible in the background. Cretaceous-Tertiary bedrock can be seen at the right of the photo and native boulders of the modern moraine of Aktineq Glacier are visible in the foreground. A figure is visible in the middle distance for scale.

Plate 4.2

"Dead Lichen" Valley. The figure is in the lichen kill zone, which has a largest live lichen size of 31 mm. The 51 mm surface in the foreground appears dark in comparison, and has no dead lichen. Gneissic bedrock is visible in the background.

Plate 4.3

The 10 mm modern moraine of "Stagnation" Glacier appears pale in comparison to the 59 mm moraine just west of it (to its left in the photo).

Plate 4.4

Lichen-free modern moraine, Aktineq Glacier, composed completely of native debris.

Plate 4.5

The lichen-free modern moraine of Ujarasukjulik Glacier appears grey against the adjacent ancient tundra.

Plate 4.6

Boulders of the modern moraine of Ujarasukjulik Glacier are completely lichen-free, consisting of 100% native lithologies.

4.5

4.6

Plate 4.7

Ice contact deposit near the town of Pond Inlet. A normal fault can be seen in the crossbedded and plane parallel laminations below the dark organic stratum, marked by the shovel. 93-RD-03 was taken from this organic bed, which appears to have formed after the fault occurred, implying *in situ* growth on a paleosurface. The shovel is about 30 cm long.

Plate 4.8

Ice-dammed "Bigfoot" Lake, looking northwest from Aktineq Glacier. The raised beach surface is visible at the periphery of the lake, which drained to its current level sometime between 1961 and 1982.

Plate 4.9

Partially drained lake on metamorphic bedrock, east of the Salmon River Lowlands. A lichen trim line marks the former lake level and rock glaciers can be seen on the right side of the photo.

APPENDIX I

¹⁰Be Measurement Data, University of Pennsylvania

Sample Number	Sample Weight (g)	⁹ Be Weight (mg)	Normalized ¹⁰ Be/ ⁹ Be	Normalized ¹⁰ Be/g	Elevation (m)	Correction for Elevation
AC-93-01	10.107	1.236	5.130×10^{-14}	4.193×10^5	118	3.728×10^5
AC-93-03	5.681	1.147	2.736×10^{-14}	3.692×10^5	60	3.469×10^5
AC-93-06	9.954	1.302	5.586×10^{-14}	4.884×10^5	340	3.544×10^5
AC-93-07	10.618	1.174	5.472×10^{-14}	4.044×10^5	130	3.555×10^5
AC-93-09	10.088	1.233	4.104×10^{-14}	3.353×10^5	80	3.091×10^5

APPENDIX 2

Aluminum Data

The samples have very low $^{26}\text{Al}/^{10}\text{Be}$ ratios, due to anomalously low quantities of ^{26}Al in each. The measurement data are shown in table A2.1 and the burial and erosion histories in table A2.2. The maximum possible erosion rate indicated by the quantity of ^{10}Be present in the samples is also shown in table A2.2. Figure A2.1 presents the data set on the cosmogenic dating curve.

The data indicate burial for about 2.2 Ma for all samples, followed by 150-220 ka of exposure. It is difficult to imagine continuous ice cover in the area for such an extensive period of time, if burial was due to ice. If it was due to sediment, the cover would have to have been at least 4 m thick (sample AC-93-06, table 4.2), a thickness which is also difficult to imagine to have been formed by cold-based ice and subsequently completely eroded away.

Because the samples were small and the quartz separation poor (three of the original eight samples did not yield enough quartz for the method), it is suspected that stable aluminum contamination is responsible for the anomalously low ^{26}Al results. ^{26}Al is measured as a ratio of $^{26}\text{Al}/^{27}\text{Al}$ after adding a ^{27}Al carrier. The amount of ^{26}Al is calculated using the known quantity of carrier. Thus, a significant amount of ^{27}Al in the sample can swamp the ^{26}Al signal such that the ratio becomes much too low, resulting in a ^{26}Al measurement that is anomalously low. This appears to have happened for these five samples. In fact, sample AC-93-03 contained such a minute calculated quantity of ^{26}Al that the $^{26}\text{Al}/^{10}\text{Be}$ ratio could not be obtained.

It is unfortunate that the samples were too small to obtain good quartz separations and useful ^{26}Al data. However, the ^{10}Be data are still considered valid, as ^{10}Be is measured as a ratio of $^{10}\text{Be}/^9\text{Be}$. ^9Be is extremely rare in granites and

gneisses, so contamination is unlikely. The conclusions of the thesis based on ^{10}Be data thus remain valid.

A suggestion for future work is to take large samples, especially when quartz contents are low. A volume of rock chips should ideally be about 7500 cm^3 , a volume approximately equivalent to the volume of two shoe boxes. If quartz contents are high (50-60%), less whole rock is necessary, but overestimation of quartz content is always a danger. The author would recommend taking as much sample as possible under all circumstances.

APPENDIX 2

Table A2.1 ^{26}Al Measurement Data, University of Pennsylvania

Sample Number	Normalized ^{26}Al	Normalized $^{27}\text{Al/g}$	Elevation (m)	Correction for Elevation	$^{26}\text{Al}/^{10}\text{Be}$
AC-93-01	1.80×10^{-13}	5.97×10^5	118	5.311×10^5	1.42
AC-93-03	7.20×10^{-15}	2.90×10^4	60	—	—
AC-93-06	7.70×10^{-14}	9.46×10^5	340	6.861×10^5	1.94
AC-93-07	2.13×10^{-13}	8.30×10^5	130	7.295×10^5	2.05
AC-93-09	1.49×10^{-13}	5.77×10^5	80	5.323×10^5	1.72

APPENDIX 2

Table A2.2 *Burial, exposure and maximum erosion of samples*

Sample Number	Burial (ka)	Exposure (ka)	Maximum Erosion Rate
AC-93-01	2647.7	221.8	8.52
AC-93-03	—	—	9.17
AC-93-06	2096.0	161.2	8.97
AC-93-07	1989.4	153.6	8.95
AC-93-09	2320.5	155.8	10.32

Figure A2.1.1. Samples AC-93-01, -06, -07 and -09 plotted on the cosmogenic dating curves. Sample AC-93-03 is not shown, as its ^{26}Al quantity was too low to obtain a $^{26}\text{Al}/^{10}\text{Be}$ ratio. Graph by J. Klein.

END

0 9 - 1 1 - 9 4

FIN