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MODERN AND HOLOCENE POLLEN ASSEMBLAGES FROM ARCTIC ICE CAPS

BY

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ABSTRACT

Records of pollen deposition on arctic ice caps are used to infer paleoenvironments of the Holocene and atmospheric circulation patterns in the Arctic. As part of this study, several snow samples were collected over a broad area, over the course of several years, to investigate modern pollen deposition patterns in the Arctic. Pollen assemblages recovered from arctic snow are diverse and consist of tundra and forest types. The results show that pollen percentages and concentrations are related to the density of the regional vegetation and to the distance of the source in more productive regions. In addition, the long-distance transport of tree and shrub pollen permits the identification of regional patterns that might be used to define air mass trajectories in the Arctic. In a more detailed analysis, the seasonal and annual variations in pollen deposition in snow layers were studied on four ice caps, including one in the Russian Arctic. It is shown that the pollen succession in the annual snow layers is related to the flowering periods of arctic and southern plants. The amount of pollen reaching the ice caps varies from year to year. Furthermore the variability in the number of tree and shrub pollen increases with decreasing distance to the treeline. The last section of this study is an interpretation of a Holocene record of pollen distribution in an ice core from the Agassiz Ice Cap, Ellesmere Island. Pollen concentrations, particularly those of tree pollen, were highest in the early Holocene, decreased in the mid-Holocene, and changed relatively rapidly after *ca.* 3500 years ago. In the early Holocene, the pollen profile parallels the $\delta^{18}O$ and ice-melt records from the same ice core, indicating that the warmest summer temperatures occurred very early in the Holocene.The high concentration of tree pollen in the early Holocene, when large parts of Canada were still ice-covered and forest zones more limited and generally farther away, implies that atmospheric circulation was stronger than at present. The data may be of significant value to comprehensive studies of atmospheric dynamics and vegetation changes.

RÉSUMÉ

Les données polliniques recueillies des calottes glaciaires de l'Arctique sont utiles pour tenter de définir les paléoenvironnements de cette région. Aux fins de cette étude, plusieurs échantillons de neige ont été prélevés sur une vaste superficie et sur plusieurs années, pour étudier les modèles modernes de déposition des grains de pollen dans l'Arctique. Les spectres polliniques sont diversifiés et consistent principalement de grains de pollen typiques d'un environnement forestier et de toundra. Les résultats démontrent que les pourcentages et les concentrations polliniques sont reliés à la densité du couvert végétal régional et à la distance des régions sources productives. De plus, la longue distance de transport du pollen d'arbres et d'arbustes permet l'identification de modèles régionaux qui peuvent être appliqués à la définition des trajectoires de masses d'air de l'Arctique. Dans une analyse plus détaillée, les variations saisonnières et annuelles du spectre pollinique dans la neige ont été étudiées pour 4 calottes glaciaires, dont une se situant dans l'Arctique russe. Il apparaît que la succession pollinique dans la neige est reliée aux périodes de floraison des plantes arctiques et du sud. La quantité de pollen qui atteint les calottes glaciaires varie d'année en année. Ainsi, la variabilité du nombre de grains de pollen d'arbres et d'arbustes augmente lorsque l'on s'approche de la limite des arbres. La dernière section s'attache à l'interprétation de la séquence Holocène d' une carotte de glace de la calotte glaciaire Agassiz, île d'Ellesmere. Les concentrations polliniques, particulièrement celles des arbres, ont atteint leur maximum au début de l'Holocène pour décroître par la suite avant de changer relativement rapidement après *ca* .3500 ans. Au début de l'Holocène le profil pollinique s'apparente aux $\delta^{18}O$ et aux données de fonte de glace issues du même forage. Celles-ci nous indiquent que les températures estivales étaient les plus douces au tout début de l'Holocène. Les hautes concentrations du pollen arboréen au début de l'Holocène, alors que de larges superficies du Canada étaient encore recouvertes de glace et que les zones forestières étaient limitées et généralement éloignées, impliquent que la circulation atmosphérique était plus intense qu'aujourd'hui. Ces données peuvent être utiles pour les études portant sur les changements de circulation des masses d'air ainsi que pour les variations de la répartition de la végétation.

PREFACE

The following chapters were written for publication in peer-reviewed scientific journals. The chapters are formatted according to the guidelines of these publications, thus causing some minor inconsistencies in style. Two of the chapters were written with my dissertation supervisor at the University of Ottawa and my research partners at the Geological Survey of Canada. These co-authors helped me gather field data and I benefitted from our discussions of the results. I was involved in all aspects of the field work and counted the majority of the pollen samples. I developed the methodology and I was primarily responsible for writing the papers. Chapter 2 was written for publication with the assistance of K. Gajewski and R.M. Koerner. Chapter 4, written with the assistance of R.M. Koerner, K. Gajewski and D.A. Fisher, has been submitted to *Quaternary Research.* I am the sole author of the paper presented in Chapter 3. This paper has been accepted by *Review of Palynology and Palaeobotany* and is in press.

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CHAPTER 1

INTRODUCTION

1.1 Introduction

There is relatively little information on vegetation and environmental changes in the Canadian high arctic. Three short, late Holocene pollen diagrams have been reported from results of analyses of peat deposits on Axel Heiberg, Devon and Ellesmere islands (Hegg, 1963; Jankovska & Bliss, 1977; Gajewski *et al.*, 1995). Pollen assemblages from these deposits are mostly of local origin and few climatic and regional vegetational changes can be inferred from the data. Longer Holocene pollen records have been obtained from 1lacustrine sediment at only three sites: east-central Ellesmere Island (Hyvärinen, 1985) and two on Somerset Island (Gajewski, 1995). The diagrams suggest that a climatic deterioration has taken place since the mid-Holocene. However, the paleoclimatology of the area is not well known, and few of the proxy records are unequivocal (Bradley, 1990).

The poor coverage of pollen sites in the high arctic can be partly attributed to the region's remoteness and the difficulty of finding suitable sampling sites. Potential sites are often plagued by problems such as: 1) low local pollen production resulting in over-representation of pollen transported over long distances; 2) stratigraphic inversions due to permafrost; 3) contamination with material from older Tertiary deposits, resulting in radiocarbon dates that are too old and the introduction of redeposited Tertiary pollen grains to the Holocene pollen assemblages (Gajewski *et al.*, 1995).

In the Canadian Arctic, ice core records are important sources of paleoenvironmental information (e.g. Koerner & Paterson, 1974; Paterson *et al.*, 1977; Fisher *et al.*, 1983; Koerner & Fisher, 1990; Fisher *et al.*, 1995; Fisher *et al.*, 1998). In the accumulation zone of the larger ice caps, past snowfalls are preserved in layered sequences for approximately the past 100,000 years. Records of temperature changes and of past atmospheric composition have been recovered by drilling through the ice. Pollen grains have also been retrieved from the ice layers of these ice caps. Long records have been produced for the Devon Ice Cap (McAndrews, 1985) and the Agassiz Ice Cap on Ellesmere Island (Bourgeois, 1986). Although these pollen records are not affected by the problems of stratigraphic inversions and contamination by older material listed above, the interpretation of the data has nonetheless been complicated by our limited knowledge of pollen dispersal and deposition in the Arctic, including ice caps. As a result, no complete or reliable Holocene pollen records from arctic ice cores have been produced to date.

1.2 Overview and Organization of Dissertation Research

The aim of this dissertation is to demonstrate that pollen records extracted from arctic ice cores are valuable sources of paleoenvironmental information. These records should complement other paleoecological studies in the Arctic, thereby furthering our understanding of climate change in the north. In addition, because a large percentage of the pollen deposited on ice caps originates from distant sources, pollen grains should be reliable air mass tracers. As such, they would contribute significantly to the reconstruction of past atmospheric conditions inferred from the study of ice cores.

To realize the potential of ice core palynology, several steps are necessary. First we need to understand how pollen is transported to the drill site or to the top of the ice cap. Since there is obviously no local pollen production, all of the pollen grains must be transported to the site from the tundra surrounding the ice sheet or farther away. Observations have indicated that pollen assemblages from ice cores are diverse and consist of tundra as well as forest types. In order to interpret the environmental significance of these assemblages, it is necessary to identify patterns of pollen transport in the Arctic. To accomplish this, a spatial array of snow samples was collected and analyzed to better understand the amounts of pollen produced by different vegetation zones and the distances and pathways of pollen transport. This study, presented in Chapter 2, illustrates that the pollen assemblages in arctic snow are related to the regional vegetation, and that patterns of long-distance transport are also interpretable. In Chapter 3, the results of a high resolution study on pollen deposition in snow layers of three ice caps, a small glacier in the Canadian Arctic and an ice cap in the Russian Arctic are presented for periods ranging from one to thirteen years. In this case, seasonal bands could be identified and the pollen assemblages traced to the approximate source of release. It is shown that the pollen assemblages of summer and spring layers are different and that they are clearly related to the known phenology of arctic and southern plants. The studies discussed in Chapter 2 and Chapter 3 indicate that the pollen transported to ice caps can provide a reliable index of vegetation zones and of atmospheric circulation patterns.

Following these encouraging results, the pollen analysis of an ice core retrieved from the Agassiz Ice Cap, Ellesmere Island was undertaken, and the results are presented in Chapter 4. Significant variations in pollen percentages and concentrations are detected in the 11,000 year record. The pollen profile is compared with other paleoecological records from the Arctic, as well as with the oxygen isotopes ($\delta^{18}O$) and ice melt records from the same ice core, thereby giving some indication of the environmental changes that have taken place in the region. The results can be interpreted to indicate how the vegetation and atmospheric circulation have changed during the Holocene and, thus, provide a significant and detailed source of information about past environments.

There are four appendices to this dissertation. They contain papers, either published or in press, that are each closely related to one of the chapters. They were not included in the main body of the dissertation because they diverge slightly from the unifying theme. Appendix A (*Paleoenvironments of the Canadian high arctic derived from pollen and plant macrofossils: problems and potentials*) discusses pollen analysis in the Canadian high arctic, a subject also presented in the introduction. The paper provides a background for the region and discusses the problems of arctic palynology that must be considered when interpreting pollen profiles. This paper may be a useful introduction for readers unfamiliar with arctic palynology. In Appendix B (*Snow chemistry and pollen deposition in the circum-polar high arctic*), the fluxes of natural (i.e. pollen) and anthropogenic aerosols in the snow of the circum-arctic are investigated. Notwithstanding the considerable interannual and spatial variations, the data show decreasing fluxes from the Eurasian side to the Canadian Arctic Islands. This study, which will be published elsewhere in more detail, is an extension of the survey presented in Chapter 2. Appendix C (*Establishing the chronology of snow and* *pollen deposition events on Agassiz Ice Cap (Ellesmere Island, Northwest Territories) from autostation records*) describes a distinctive snow layer formed each year at the end of the melt season on arctic ice caps. This layer, which is also analysed in Chapter 3, usually has the highest concentration of herbaceous pollen and could possibly be used to infer summer conditions in the surrounding tundra. An automatic weather station installed at the site made it possible to determine the timing of this deposition event and therefore to establish the atmospheric conditions at the time, an approach that will likely be used in future pollen analysis studies on the Agassiz Ice Cap. Appendix D (*Present and past environments inferred from Agassiz Ice Cap ice core records*) reviews the significant amount of data produced from snow and ice studies on Agassiz Ice Cap, the focal point of this dissertation. It provides background information for the Agassiz Ice Cap pollen record discussed in Chapter 4.

1.3 Previous Work on Snow and Ice Core Palynology

Although glaciers and ice caps represent enormous natural pollen traps, their potential for pollen analysis studies has not been fully explored (Faegri and Iversen, 1989). Vareschi first published a series of papers on the subject between 1935 and 1942 that were later summarized by Godwin (1949). He demonstrated that snow deposited throughout the year registers the seasonal changes in the pollen rain, and consequently, pollen spectra can be used to establish the snow stratigraphy. In subsequent studies in southern Alaska and the Austrian Alps, Heusser (1954), Ambach *et al.* (1966) and Bortenschlager (1970) used the pollen content of glacier ice to confirm the annual snow stratigraphy in areas of heavy summer melting. These first studies were based on alpine glaciers where the snow layers were deposited relatively close to the sources of the pollen. In the Canadian Arctic, Lichti-Federovich (1974), Short & Holdsworth (1985), Bourgeois (1990a) and Alt & Bourgeois (1995) studied seasonal pollen variations in the snow layers of the Devon, Penny, and Agassiz ice caps. Bourgeois (1990b) analysed an annual snow sample from the Dye-3 drill site in southern Greenland. In these studies the emphasis was no longer on using pollen for stratigraphic purposes but as a paleoenvironmental indicator. At the Arctic sites, pollen concentrations varied from approximately 10 to 100 grains/L compared to the several thousands grains/L that were obtained from temperate glaciers. As there is no local pollen contribution and the regional tundra pollen input is small, the pollen assemblages retrieved from these ice caps were composed of a very high percentage of long-distance transported pollen, primarily tree pollen from the boreal forest. In these studies as in the previous ones, the pollen assemblages and concentrations were shown to change through the annual snow increment. Alt and Bourgeois (1995) suggested that extreme herbaceous pollen concentrations in the late summer/early autumn snow layer on Agassiz Ice Cap reflect the summer climate conditions of the inland areas of the intermontane region located to the southeast of the ice cap, whereas Bourgeois (1990a) tentatively related unusually high concentrations of tree pollen to atmospheric circulation patterns. These studies showed that high resolution pollen analysis of arctic ice caps is possible and that it has a potential for tracing air mass movement. These arctic records also showed a high variability, but they could not be readily compared as they varied in length, years, number of increments analysed and presentation of the data.

In the first study on the distribution of pollen in arctic snow, pollen concentrations, based on one to fifteen year averages, were obtained from surface snow and shallow ice cores at fifteen sites in the Canadian Arctic, the adjacent Arctic Ocean and in northern Greenland (Bourgeois *et al.* 1985). This study, which established the long-range dispersal of tree pollen from the boreal forest to the Arctic regions, including the Arctic Ocean, revealed a steep, exponential decrease in deposition rates of tree pollen between the tree line and the high arctic, but further decreases were not detected within the high arctic. The few available sites for this study also suggested that pollen concentration could be used to trace pollen transport, if sufficient detailed information were available.

The potential of pollen analysis from deep ice cores was first tested by Fredskild & Wagner (1974) on a few ice samples from the Camp Century and Dye-3 cores in Greenland. Most of the pollen grains in the ice were derived from plants that grow in Canada. Each of the samples consisted of one litre or less and, consequently, not enough pollen grains were obtained for a reliable percentage calculation. However, Fredskild & Wagner (1974) did notice an increase in concentration in the most recent sample from Camp Century.

In her preliminary study of the Devon Ice Cap ice core, Lichti-Federovich (1975) analyzed a much greater volume (between 14 to 73 L per time interval) than had Fredskild and Wagner. However, she did not use ice core sections but relied on the meltwater recovered during thermal drilling. McAndrews (1984) completed the analysis of the 299 m-long Devon ice core. Higher pollen concentrations were found for the late Holocene and last interglacial periods (approx. 7 grains/L) than for the early Holocene and Wisconsinan period (1-2 grains/L). *Alnus* pollen dominated the interglacial periods, whereas the coldest intervals had higher percentages of *Betula* and *Artemisia*. Most of the grains probably originated from the low Arctic; relatively few pollen grains were from south of the summer position of the arctic front. McAndrews emphasized the possibility that pollen from glacier ice could have been redeposited in sediments at the end of the last glaciation, thus creating anomalous pollen assemblages. The study of the Devon Ice Cap core suggested that vegetation records could be obtained from ice core pollen samples, but there were too few analyses to develop reliable stratigraphies.

In a pollen study of the Agassiz Ice Cap, Bourgeois (1986) produced a concentration diagram for the Holocene period based on meltwater samples obtained during the process of drilling a 338 m core with a thermal drill in 1977. Pollen concentrations were extremely low (usually <1 grain/L) and the results were averaged into 500 year increments. Concentrations were lowest in the early Holocene and increased slightly around 7500 years ago due to higher *Alnus* and, to a lesser extent, to an increase in *Betula*. The regional tundra types began increasing at about 6000 years ago, but the highest concentration was reached only 3000 years later. Tree pollen also increased at the time but did not peak until 1000 years ago. Although the late maximum of tundra pollen was tentatively attributed to the persistence of ice age conditions in the northern part of Ellesmere Island until about 6000 years BP, the highest tree pollen concentration in the late Holocene could not be explained.

Following that study, Koerner *et al.* (1988) analysed several large ice samples from the early Holocene and pre-Holocene sections of a second borehole drilled on the Agassiz Ice Cap in 1979. The purpose of the study was to use pollen analysis to help clarify the stratigraphy in the older, undated section of the core. In order to obtain smaller time increments than normally possible with the 1.5 m core segment while retaining an adequate pollen sum, 30 to 40 kg ice samples were melted laterally, from the bottom of the hole and the water pumped up to the surface. Results of pollen analysis strongly suggest that the ice near bedrock was formed in a relatively warm period, probably towards the end of the last interglacial. Consequently, the ice cap must have melted completely in an earlier part of the same interglacial. From the similar ice properties found at the bottom of other ice cores in Greenland, Koerner (1989) argued that these sites must also have been ice-free during part of the last interglacial.

Higher pollen concentrations were obtained from the early and pre-Holocene ice samples than were recovered from the melt-tank samples of the Agassiz 1977 core. This, combined with new results from surface samples from several sites, made Koerner *et al.* (1988) suspect the validity of the low concentrations previously obtained for both the Devon and Agassiz ice cores. However, the reasons for the discrepancy could not be found. Those samples were obtained from the meltwater that was continuously drawn into the reservoir of the thermal drill during the process of drilling each 1.5 m core increment. For some unexplained reason, some of the pollen grains were apparently lost during the drilling process. Interpretation of the data was based on the remaining pollen grains. Since no other data were available for comparison, the error could not be detected until several other samples were analyzed from snow and ice cores (Koerner *et al.*, 1988). Those results also showed that, in order to obtain a reliable pollen profile, a dedicated ice core would have to be drilled from the Agassiz Ice Cap.

Recently, ice core pollen records have been obtained from outside Canada and

Greenland. A pollen profile was produced from the Vavilov Ice Cap on October Revolution Island (Severnaya Zemlya), in the Russian Arctic (Andreev *et al.*, 1998). The record showed a highly variable number of pollen with an overwhelming long distance component. An increase of West European species, noticed in the upper part of the core, suggests a predominance of summer air masses from the south during the last 500 years.

Pollen records have also been produced from nonpolar ice caps. Preliminary results obtained from an ice core drilled through the tropical Huascaran Ice Cap in Peru (Thompson *et al.*, 1995) showed lower pollen concentrations during the late glacial maximum compared with the late Holocene, thus suggesting a cold environment and a sparsely vegetated area around the ice cap. The results seem to confirm other evidence for cooler tropics during the last glaciation. Lastly, the first complete pollen analysis of a nonpolar ice core was produced for the Dunde Ice Cap, Qinghai-Tibetan Plateau, in China (Liu *et al.*, 1998). The record is viewed as a sensitive indicator of moisture availability and vegetation density in the steppe and desert regions around the ice cap. The Holocene pollen concentrations are correlated with summer precipitation and, hence, with the position of the summer monsoon. Because of the relatively high pollen content of the ice and high resolution in these ice cores, pollen analysis is viewed as an important source of paleoclimatic information, and according to Liu *et al.* (1998), should be an integral part of all nonpolar ice core studies. The value of pollen analysis of ice cores from the nonpolar ice caps is undeniable, but potential sites are limited and they are even more difficult to access than those from the polar regions.

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CHAPTER 2

SPATIAL PATTERNS OF POLLEN DEPOSITION IN ARCTIC SNOW

Abstract

This paper discusses the pollen content of 77 snow samples, collected at 41 sites in the Canadian Arctic, the adjacent Arctic Ocean and Greenland. In spite of considerable variability in time and space, pollen assemblages can be used to identify source regions. Pollen percentages and concentrations are related to the density of the regional vegetation and to distance to the source of more productive regions. Long-distance transport of pollen is important and can be used to indicate the source of air masses. The major vegetation zones produce distinctive pollen assemblages which are retained even in sea ice samples (i.e. from snow on the sea ice surface). Studies of the concentration of pollen in snow have the potential for determining past and present characteristics of atmospheric circulation and for helping develop paleoenvironmental records.

2.1 Introduction

In polar regions, where peat or lacustrine sediment adequate for palynological studies is sometimes difficult to locate, pollen analysis of ice cores offers an alternative or a complement to other paleoenvironmental (proxy) reconstruction. However, pollen assemblages from ice cores often differ from those of other deposits. Pollen assemblages of most depositional sites usually consist of a combination of local, regional and long-distance components, with the local and regional pollen dominating. Pollen assemblages from ice cores lack the local component and are dominated by pollen from distant sources.

Pollen is efficiently transported in the atmosphere, in some cases for 100's or 1000's of kilometres. This component, termed long-distance transported pollen or simply long-distance pollen, is better represented in ice core pollen assemblages as there is no local source and the regional tundra has a low productivity. In this study, we consider *local* pollen as that originating in the immediate catchment of the depositional site and *regional* from a somewhat broader assemblage of catchment basins. Changes in the pollen assemblages through time could be indicative of changes in the composition of the vegetation in the source region, but because of the considerable distances involved, changes could also be related to changing atmospheric conditions that could affect pollen dispersion, transport and deposition on the ice cap. Therefore, it may be possible to determine changes in atmospheric circulation by the analysis of spatial patterns of pollen deposition in the arctic region (e.g. Nichols *et al.*, 1978; Barry *et al.*, 1981). Pollen analysis can provide useful complementary information to other analyses from ice cores since the pollen grains are identifiable to the family, genus or sometimes species level and in some cases, the ecological requirements and, consequently, the source area can be identified.

Pollen analysis of snow and ice cores provides information not obtainable from other sources. Long sequences of environmental change can be recorded, extending in principle to the last interglacial (Koerner *et al.*, 1988). Few terrestrial deposits have records this long. High resolution studies can resolve decadal records for the past 1000 years and annual for the past 100 years. There is a potential for resolving seasonal records, perhaps for the past few decades (Bourgeois, in press). This resolution is limited primarily by the volume of ice needed and time to analyse the data. The potential to investigate regional environmental change as well as large-scale circulation patterns makes ice core palynology a particularly useful tool if the origin of the grains can be determined. This paper contributes to this goal by presenting the results of an analysis of modern pollen in the Arctic, thus providing modern analogues for the interpretation of pollen in ice cores. These data also provide information about the characterization of pollen assemblages in various arctic environments which is of interest to researchers reconstructing paleoenvironmental records from arctic lakes and bogs.

To understand the significance of pollen records retrieved from ice cores, it is essential to understand the present-day spatial distribution of pollen in arctic snow. The snow samples should not come only from the top of ice caps where ice cores are being drilled but also, for good spatial coverage, from as wide an area as logistic support allows. Studies on modern pollen deposition in snow have concentrated mainly on establishing seasonal deposition patterns on ice caps of the Canadian Arctic Islands (Lichti-Federovich, 1974; Short and Holdsworth, 1985; Bourgeois, 1990*a*, in press; Alt and Bourgeois, 1995).

A first survey on the spatial distribution of pollen from surface snow and shallow ice cores was produced from fifteen sites in the Canadian Arctic, the adjacent Arctic Ocean and in northern Greenland (Bourgeois *et al.* 1985). This study, which measured the long-range dispersal of tree pollen from the boreal forest to the arctic regions, including the Arctic Ocean, revealed an exponential decrease in deposition rates of tree pollen between the treeline and the high arctic. The steepest gradient was between the treeline in the Mackenzie Delta area and Melville Island, in the Western Arctic. Significantly, after the initial decrease to low levels, pollen is found to be transported well into the high arctic, where no further decreases could be quantified. A second purpose of the present study is to document the spatial patterns of modern pollen transport to achieve a better understanding of arctic circulation. Because pollen can be identified (and hence its source), the data provide information about past and present air mass trajectories in the Arctic.

In the Canadian Arctic, knowledge of modern pollen dispersal comes from a limited number of sites, mostly confined to the mid and low arctic vegetation zones. Data have been derived from atmospheric samples (Ritchie and Lichti-Federovich, 1967), moss polsters and lichens (Elliot-Fisk *et al.*, 1982), surficial lake sediments (Ritchie *et al.*, 1987; Gajewski, 1995; Gajewski *et al.*, 1995) and marine sediments (Mudie, 1982). Only the snow survey discussed above (Bourgeois *et al.*, 1985) has provided a significant amount of data from the high arctic.

In the current paper, we present results of pollen analysis of snow collected at 41 sites in the Arctic between 1982 and 1995. The study area extends from 59° to 90°N. We have included sites from ice caps, glaciers, land and sea ice. More than half of the sites are located in the Canadian Arctic, the others are on the Arctic Ocean sea ice and Greenland. Site location and related information are presented in Figure 2.1 and Table 2.1.

2.2 Regional Environment

All sites in this study are north of the treeline which in North America is formed by spruce (*Picea*) species. The distance to treeline varies from approximately 50 km at Tuktoyaktuk (site #35) to more than 2000 km on Agassiz Ice Cap (sites #17 & 18) and on the Arctic Ocean. Some sites on the Arctic Ocean, in the vicinity of the North Pole, are closer to the Eurasian treeline than to the North American treeline.

Pine (*Pinus*), a major pollen producer whose pollen is efficiently dispersed in the atmosphere, is an important component of the boreal forest. In western Canada, its northern limit is reached at approximately 65°N whereas in eastern Canada, pine does not grow beyond 55°N, at which point it is limited to small populations. A few of the sites are in the low arctic zone where several shrubs, including alder (*Alnus*) , dwarf birch (*Betula*), and members of the heath family (Ericaceae), dominate the vegetation cover along with herbaceous plants, lichens and mosses. The mid-arctic is characterized by a reduced diversity of taxa as well as smaller and fewer shrub forms. The high arctic zone, the polar desert, is defined by an impoverished and discontinuous vegetation cover. However, the intermontane regions of Ellesmere and Axel Heiberg islands have a much richer vegetation than their northern latitude would suggest (Edlund and Alt, 1989), and are essentially northern extensions of the mid-arctic.

2.3 Methods

At some of the sites, such as the ice caps, sampling was done several times over the course of several years whereas at others, such as those on the Arctic Ocean and most tundra sites, samples were obtained only once at a particular site. Sites were always visited in the spring. On ice caps, shallow pits were excavated through the surface (fall, winter and spring) snow layer to the previous summer layer. At lower elevations on glaciers and sea ice, where in spring only the current year of accumulation is preserved before it is melted in summer, pits were dug to the ice surface. Samples from Arctic Ocean sea ice were collected from multi-year floes, where snow accumulates earlier than on first year ice, which first has to form in the fall. On land-based sites, we avoided well vegetated zones and when sampling the annual snow layer, were careful not to include vegetation.

Snow blocks were cut from the wall of the pit and placed in plastic bags. Care was taken to make the walls vertical to avoid seasonal bias. In a field laboratory, the snow sample was melted inside the bag and the meltwater passed through an 8μm pore size cellulose filter which should retain all of the pollen grains in the sample. The filter was stored in cool conditions until further processing in our Ottawa laboratory. Before melting the sample in the field, a known number of *Lycopodium* spores were added to the snow. This allowed us to assess the percentage of pollen grains lost in the sample due to field and laboratory procedures, and to correct for that loss when calculating the concentration values.

We followed, with minor modifications, the method developed by Bourgeois *et al.* (1985) who found the standard method of pollen preparation, which includes centrifuging and decanting with each processing step, to be inadequate for arctic ice cap snow samples because of the low pollen content. The method consists of (1) soaking the filters in hot hydrofluoric acid (HF), (2) filtering the HF and the pollen residue through an 8μm pore size polycarbonate filter, (3) rinsing with water followed by glacial acetic acid, (4) soaking for fifteen minutes in an acetolysis solution and (5) rinsing with water. At the end of the procedure, the transparent polycarbonate filter was removed from the filtering apparatus, cut in half and mounted on microscope slides for analysis. With this method, the complete pollen content of the sample, including the added *Lycopodium* spores, should be contained on the two slides.

To calculate pollen accumulation rate (PAR: grains/cm^2/yr) we need to ensure that only one year of accumulation was present or to know the time interval of deposition of the snow layer. We did not obtain the snow accumulation rates for some of the sites that would allow us to calculate influx values. In a series of traverses across high arctic ice caps, Koerner (1979) found few changes in the accumulation rates beyond 75°N, with the exception of sites close to Baffin Bay where accumulation rates increase. For example, the accumulation rate on the northern part of the Agassiz Ice Cap is 17.5 g/cm^2/yr compared to the Torngat Mountains, our southernmost site facing the Labrador Sea, which receives 70 g/cm^2/yr. In principle, sites south of 75° N receive more snow accumulation, thereby producing lower concentrations. However, according to Maxwell (1980), continental sites west of Hudson Bay, do not receive more snow than those to the north. Because of some uncertainties for some of the sites, we have left the data in the form of concentrations.

2.4 Results

Interpretation of the data: some considerations

All samples were collected prior to June, before melting has started this far north. At lower latitudes, many tree and shrub species have, by then at least, started or in some cases finished flowering. *Pinus* and *Picea* species, the notable exceptions, reach the peak of their flowering in June (Bassett *et al.*, 1978). The beginning of snow accumulation in the fall varies according to latitude and site location. On high arctic ice caps and on multi-year ice floes on the Arctic Ocean, snow can begin to accumulate as early as Late-July (Alt and Bourgeois, 1995), whereas in the high arctic tundra, it might start in mid-August. At our southernmost sites, the average earliest date for the formation of snow cover is mid-September (Maxwell, 1980). Therefore, in our northern samples, the snow accumulates when plants are still flowering in the tundra, whereas farther south, snow starts to accumulate once most plants have stopped flowering. At the southern sites, most of the local pollen in the snow would come from redeposition. However, some of the exotic herb taxa, such as ragweed (*Ambrosia*), flower later in the season and the pollen can be deposited with the snow.

On the large ice caps included in this study, a complete annual record of pollen deposition is possible because the snow accumulates during the summer and the melting that occurs re-freezes within the snow-pack. In order to be consistent in our survey, we do not include the summer layer in the annual average for these sites. These results are presented elsewhere (Bourgeois, in press). However, for Agassiz Ice Cap (site #17), where we have the longest record of annual pollen deposition, we compared the average annual pollen concentration in samples with and without the summer melt layer and found the concentration to be <1 grain/L lower in samples where the melt layer is excluded from the total. When the summer melt layer is included, *Pinus* is slightly better represented whereas *Sphagnum* moss, whose spores are deposited later in the season, is less well represented. We also compared the pollen concentrations in our study with those of Bourgeois *et al.* (1985), which were obtained on ice caps, from shallow cores representing several complete years of accumulation. We found the concentrations to be comparable in the two studies. The differences on Agassiz Ice Cap and the other high arctic ice cap sites are minor, but they probably increase at the southernmost sites where the snow-free season is longer.

Annual variations

The pollen data were first plotted for each year. The year refers to the year of sample collection although the sample includes the snow that started to accumulate at the end of the previous summer. The data are presented as pollen percentages and concentrations. The percentage diagram shows data for all sites for which a minimum of 50 grains was retrieved from the sample. Where less than 50 grains were counted, the site number and the pollen sum are noted on the diagram but the percentage field is left blank. All data are presented in the concentration diagram regardless of the sum. We assume that the values are representative of the actual concentration in the snow.

The percentage diagram (Fig. 2.2) shows considerable intersite and interannual variability. The tree and shrub taxa (i.e. *Pinus*, *Picea*, *Alnus*, *Betula*, and *Salix*) usually dominate the pollen assemblages. *Pinus* shows the most interannual variability of all tree/shrub taxa. *Picea,* the dominant tree in the boreal forest, is not well represented and high percentages are found only at the southernmost sites. *Alnus*, important in the boreal forest and forest tundra, has the most pronounced latitudinal gradient, whereas *Betula* is present in relatively large amounts in most pollen samples. *Betula* is found in significant quantity in the low arctic tundra. Willow (*Salix*), which grows up to the high arctic tundra, is a major component of the pollen assemblages at all the sites and its percentages increase with latitude. Of the non-arboreal pollen types (herbaceous plants), members of the Poaceae (grass) and Saxifragaceae (saxifrage) families dominate the assemblages. Poaceae plants are widely distributed, whereas Saxifragaceae are important in the arctic regions. Saxifragaceae percentages are larger in the north. *Sphagnum* spores are also a major component of these arctic pollen assemblages but show large interannual variations. The year 1987 stands out because of the exceptionally high percentages at all sites, except at the most southerly sites where *Alnus* dominates.

Because of the large area covered in this study, pollen concentrations (Fig.2. 3) show large variations. Total concentrations vary between 1 and 5280 grains/L. All of the high concentrations are found in the southern part of the Arctic where they tend to overwhelm the lower concentrations in the diagram. The years 1983, 1984 and 1987 had particularly high concentrations and were followed by four years of relatively low concentrations. In 1992 and 1993, several samples had high concentrations. The interannual variability of *Sphagnum* spores is clearly visible in this diagram. Although large amounts of tree pollen and *Sphagnum* are visible some years, the high concentrations generally occur in low arctic tundra.

Regional variations

To better illustrate the spatial variations in these data, we have also sorted the sites by type of environment or vegetation zone. Within each zone, the sites are listed with the most northerly site always at the top. Average values are used for those sites that were sampled more than once. In the percentage diagram (Fig. 2.4), we exclude sites with a pollen sum of <50 grains, but include those sites in the concentration diagram (Fig. 2.5). In both diagrams, variability has been reduced by averaging the annual data for some of the sites. However, there is still considerable variability within vegetation zones which seems typical of pollen assemblages in tundra environments (e.g. Gajewski, 1995). Some of these variations may be partly explained by the location of the site (i.e. an ice cap vs. a tundra site) but others (i.e. high *Alnus* and *Sphagnum* at some Arctic Ocean sites) are more difficult to explain, since the pollen at these sites is presumably transported from a similar source to the south.

Pollen assemblages from the low arctic are characterized by high percentages of the shrubs *Alnus* and *Betula* (Fig. 2.4). *Pinus* is also important in that zone whereas *Picea* reaches >20% only in the Torngat Mountains (site #41) of northern Labrador. The dominance of low arctic shrubs and boreal forest trees in the low arctic suppresses the other pollen types. Only Ericaceae, which is widespread in the northern boreal forest and low arctic, and Poaceae, which is common throughout the area, remain relatively abundant in that zone. Pollen assemblages from the two mid-arctic sites do not differ significantly from those of the low arctic. Similarly, sites from the southern half of the high arctic have high percentages of tree and shrub pollen. However *Salix*, and other herbaceous taxa, such as

Cyperaceae, *Oxyria*, Poaceae and Saxifragaceae, and *Sphagnum* become increasingly important in the northern part of the high arctic zone. Pollen assemblages from the southernmost sites on the Arctic Ocean are generally similar to those from the northern part of the high arctic, except for *Sphagnum* which completely dominates the assemblages at some Arctic Ocean sites. Interestingly, percentages of *Betula, Pinus* and Poaceae increase towards the north and *Pinus* reaches a maximum in the vicinity of the North Pole. A few other pollen types, such as *Ambrosia*, *Artemisia*, Chenopodiaceae and Cyperaceae, also show a small increase at the higher latitudes.

Concentrations decrease considerably towards the north. In the low arctic, the highest concentrations of *Pinus, Picea* and *Sphagnum* are found at sites that are closest to the treeline whereas *Alnus* and *Betula*, which grow near some of the sample sites, remain high throughout that zone. Southern tree types, *Ambrosia*, *Artemisia* and Ericaceae are also highest in the low arctic. Pollen concentrations remain relatively high in the mid-arctic. In the high arctic, concentrations of tree and shrub pollen are high at the southernmost sites but decrease rapidly northwards (after site #27, Bathurst Island). *Salix* and Saxifragaceae increase in the high arctic. At the northernmost sites of the high arctic zone and on the Arctic Ocean, most pollen types are at their lowest concentrations. *Pinus* is the only tree pollen that shows an increase towards the Pole. Chenopodiaceae, present in small amounts but not included in the concentration diagram, also increases in the vicinity of the North Pole.

Principal components analysis

A principal components analysis (PCA) was performed on the pollen percentages

(Fig. 2.6). All samples were included in the analysis irrespective of the year of collection. The first three components explain 13%, 10% and 8% of the variance of the 24 pollen types in the 77 pollen samples, respectively. The first factor is negatively loaded on pollen types typical of the high arctic (Saxifragaceae, *Salix*, *Oxyria*, *Papaver*) and positively loaded on trees and shrubs of the boreal forest and forest tundra (*Pinus*, *Picea*, *Alnus*). The loadings are the correlations of the pollen types with their representation in the component (Johnston, 1978). The scores are interpreted as the projections of the original points onto the new component axes (Johnston, 1978). As expected, high positive scores are found to the south, especially in Quebec-Labrador and Baffin Island. However, some samples in the Arctic Ocean also have positive scores, suggesting the increased importance of tree pollen in assemblages from this region. Negative scores are found primarily in the northern Queen Elizabeth Islands. Pollen assemblages in the northwest also have negative scores. The second component contrasts pollen types of the forest tundra and low arctic (*Alnus*, *Betula*) against pollen types characteristic of southern areas (*Ambrosia*, southern trees (ST), Chenopodiaceae, and Tubuliflorae). Negative scores are typically found to the south. Interpretation of the third component is not as clear as that of the first two. *Sphagnum* and Caryophyllaceae are positively loaded on the third component. The major habitat of *Sphagnum* is in peat bogs in boreal regions and there are few species found in the Canadian Arctic (Crum, 1986). Negative scores are found on the continental Northwest Territories and the southern Arctic Islands. Many of the samples from the Arctic Ocean also have negative scores.

2.5 Discussion

The principal components analysis summarises the variability of these pollen assemblages. The loadings illustrate that the pollen from the major communities of the arctic (high, mid, low arctic) are most important in the region of deposition. Although there is considerable intersite variability, large regions can be identified. In spite of the large distances that the pollen is transported, the major pollen assemblages can be identified in these samples, even on the Arctic Ocean. This suggests that these can be used to trace air mass movements if sufficient data are available. The origin of the samples can be identified by interpretation of the assemblages.

Herb pollen

The major herbaceous pollen types presented in this diagram are found throughout the Arctic, and this is reflected by a more gradual northward decrease in concentrations than was obtained for the tree/shrub pollen. Higher concentrations are obtained in the south, but these taxa are still well represented until at least 80°N. The lower concentrations are to be expected as plant density decreases to the north. In this study, *Salix* and Saxifragaceae are commonly representative of the high arctic environment and the PCA shows that taxa such as Papaveraceae, *Oxyria*, and *Dryas* are also correlated with these types. These results suggest that arctic environments can be distinguished by the presence of certain pollen types in spite of the large amount of pollen from southern areas.

Tree pollen

Our study clearly shows the importance of long-distance transport of pollen into the arctic region. Trees and low arctic shrubs are well represented far beyond their range. Their percentages remain high into the Islands, to a latitude of approximately 75° N, at which point species associated with the high arctic tundra as well as *Sphagnum* spores replace them in relative importance. *Betula* and *Pinus* percentages increase on the Arctic Ocean, the latter increasing towards the North Pole. However, their relatively high values may be partly due to the reduced representation of high arctic tundra types. The pollen concentration diagram (Fig. 2.5) confirms that *Pinus* values are higher between the North Pole and 86°30' N, whereas *Betula* concentrations remain constant on the Arctic Ocean. This suggests that the source of this *Pinus* pollen is in Eurasia, whereas the constant values of *Betula* on the Arctic Ocean suggest that the sources of the pollen are from both Eurasia and North America. If this were not the case, there should be a decrease away from the source. Because these Arctic Ocean sites were sampled in different years, it suggests that the intrusion of pollen bearing air masses from Eurasia towards the Canadian high arctic could be a common occurrence.

Sphagnum spores

The distribution of *Sphagnum* in the percentage diagram is not clearly related to latitude. Part of the variability is due to the percentage computations. For example, *Sphagnum* percentages are lower in the low arctic samples, where there is more of the plant growing than in the Arctic Islands where there is less of it growing (Crum, 1986). This is probably due to the fact that the percentages in the low arctic are swamped by the high pollen-producing taxa such as *Pinus, Picea* and *Alnus*. However, *Sphagnum*, as indicated by the concentrations, is apparently well transported in the atmosphere, as it is an important component of the pollen assemblages of Arctic Ocean samples. This could be misleading because the high percentages and concentrations of sites #7, 10, 13 and 15, on the Arctic Ocean, were sampled the same year. Although shown for the year 1987, this event occurred in the fall of 1986 (as determined by the seasonal results from Agassiz Ice Cap; Bourgeois, in press). Given the high values of *Sphagnum* percentages and large peaks in concentrations, the productivity and transport of this species requires more study. It is not clear why so many spores of this type are being transported to the high arctic and this needs to be resolved with more study.

Pollen concentration and deposition

The highest tree and shrub concentrations are found at sites at/or near where these taxa grow as shown by high values in low arctic samples. Notwithstanding the large intersite variations, the concentrations of these taxa are highest between the southernmost site in this study and approximately 75°N. Southern tree types and *Ambrosia* pollen, although present in smaller amounts, follow the same pattern regardless of the more distant origin of the pollen grains. We have no explanation for this possible boundary but perhaps further work will better determine the location of this transition zone. This is of particular of interest because Devon Ice Cap is situated at this latitude, and therefore past variations in the location of this transition may be recorded in the Devon Ice Cap ice cores. In an earlier study on pollen dispersal in arctic snow, Bourgeois *et al.* (1985) found a steep exponential decrease in tree pollen deposition rates between treeline and approximately 74°N and a relatively flat distribution north of 74° regardless of distances from the treeline. The steepest decrease was between treeline in the western Arctic and Melville S. Ice Cap (our site #24). By examining the mean atmospheric conditions in both the summer and winter at the surface and at upper levels, Bourgeois *et al.* (1985) concluded that in the Arctic Islands, the year-round dominance of north-westerly flow at all levels of the troposphere could not generally support a direct transport of tree pollen into the high arctic from sources directly to the south. The results of this study tend to confirm the existence of a boundary situated between 74° and 76°N. Our data from Melville S. Ice Cap still show low concentrations of tree and shrub pollen but some regional differences are revealed by the addition of other sample sites. Samples from Mould Bay (site #23) at 76°N and from northern Banks Island (site #30), both on the western edge of the Arctic Islands, have considerably higher *Pinus* and *Picea* concentrations than sites of comparable latitudes in the central and eastern Arctic and are thus reflecting the relative proximity of treeline in the Mackenzie Delta basin. Conversely, sites from the central Arctic, as far north as 75° N, have much higher concentrations of *Betula* and *Alnus* than corresponding sites in the western Arctic and this may be a reflection of transport from sources in Keewatin. The few sites from the eastern low and mid-arctic do not show particularly high concentrations, with the exception of the Torngat site (#41), which is close to the treeline. However, the higher snow accumulation, at sites facing Baffin Bay, has a diluting effect on the concentrations. Furthermore, we have not compared the effect of elevation, but at Dye-3 (site #38), our highest site at 2474 m, the high elevation combined with high accumulation must certainly account for the lower than expected pollen concentrations.

2.6 Conclusions

The considerable variability in time and space of the pollen concentrations in particular, suggest that many samples would be needed to reconstruct paleoenvironments and/or past atmospheric circulation patterns. In spite of considerable interannual variability, pollen assemblages of each different vegetation zone are clearly distinguished, especially in the concentration data. The pollen grains derived from the different vegetation zones are retained as assemblages regardless of the long distance of travel. The ability to identify pollen grains to family, genus or spp., in combination with the geographic specificity of vegetation zones offers the possibility of tracing air mass movements from their source regions. Many boreal and arctic plants are widely distributed and this reduces the resolution of the air mass movement that can be interpreted. However, the use of pollen assemblages (percentages and concentrations) and the analysis of spatial distribution does permit atmospheric circulation patterns to be analysed. In this case, we could identify the presence of air masses from Eurasia by the increase in pine pollen in the northernmost samples, and incursions of temperate air to the North American Arctic by the presence of taxa typical of boreal regions. Increased regional productivity can be determined by increased in the representation of tundra types. This has application to reconstructing past climate as well, since past assemblages are preserved in ice cores. Pollen analysis of these cores can provide useful information at many scales that can complement other analyses in ice cores.

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Table 2.1 Location of sites and environmental setting.

	SITE / YEAR	LAT. (°N)	LONG. (°W)	ASL (m)	ENVIRONMENT	VEGETATION ZONE
1	North Pole *	90°00	-	0	sea ice	-
2	Transglobe / 1982 **	88°30	73°30	0	sea ice	-
3	A.O. Site 1 / 1990	86°29	66°19	0	sea ice	-
4	A.O. Site 2 / 1989	86°26	89°22	0	sea ice	-
5	A. O. Site 4 / 1993	86°00	75°00	0	sea ice	-
6	Cesar / 1983 **	85°52	108°48	0	sea ice	-
7	A.O. Site 1 / 1987	85°02	130°16	0	sea ice	-
8	A.O. Site1 / 1986	84°48	73°09	0	sea ice	-
9	A.O. Site 1/1989	84°39	68°22	0	sea ice	-
10	A.O. Site 2 / 1987	84°20	119°34	0	sea ice	-
11	White Dwarf / 1982 **	83°56	4°05	0	sea ice	-
12	A.O. Site 1 / 1988	83°39	73°08	0	sea ice	-
13	A.O. Site 3 / 1987	83°34	109°37	0	sea ice	-
14	Ward Hunt / 1984	83°05	73°30	-	land	high Arctic
15	A.O. Site 4 / 1987	82°39	103°21	0	sea ice	-
16	Nord / 1982 **	81°45	18°00	-	land	high Arctic
17	Agassiz *	80°49	72°54	1670	ice cap	high Arctic
18	Agassiz 2 *	80°49	72°58	1730	ice cap	high Arctic
19	Ice Island / 1990	80°40	98°00	0	sea ice	-
20	Eureka / 1989	79°59	84°45	10	land	high Arctic
21	Meighen *	79°57	99°10	240	ice cap	high Arctic
22	Sawtooth / 1993	79°39	83°05	750	glacier	high Arctic
23	Mould Bay / 1989	76°15	119°10	15	land	high Arctic
24	Melville S. *	75°40	115°00	610	ice cap	high Arctic
25	Truelove / 1985	75°40	84°32	20	land	high Arctic
26	Devon *	75°25	82°40	1710	ice cap	high Arctic
27	Bathurst / 1984	75°22	100°36	15	land	high Arctic
28	Devon 2 / 1988	74°40	91°18	0	land	high Arctic
29	Lowther / 1984	74°28	97°37	20	land	high Arctic
30	Banks / 1985	74°10	118°40	25	land	mid-Arctic
31	Beaufort S. / 1985	73°30	135°00	0	sea ice	-
32	Prince Wales 1 / 1993	73°07	96°41	243	land	high Arctic
33	Prince Wales 2 / 1984	72°41	98°47	25	land	high Arctic
34	Barnes / 1992	70°31	72°54	-	ice cap	mid Arctic
35	Tuktoyaktuk *	69°15	133°00	-	land	low Arctic
36	Hayes River / 1987	67°32	94°05	-	land	low Arctic
37	Penny *	67°17	65°55	1800	ice cap	mid-Arctic
38	Dye-3 / 1983 ***	65°11	43°49	2475	ice sheet	low Arctic
39	Baker Lake / 1987	64°18	96°05	-	land	low Arctic
40	Yathkyed L. / 1987	62°42	98°18	150	land	low Arctic
41	Torngat / 1983 **	59°00	63°45	925	glacier	low Arctic

* Site sampled more than once
** From Bourgeois et al., 1985
*** From Bourgeois, 1990b

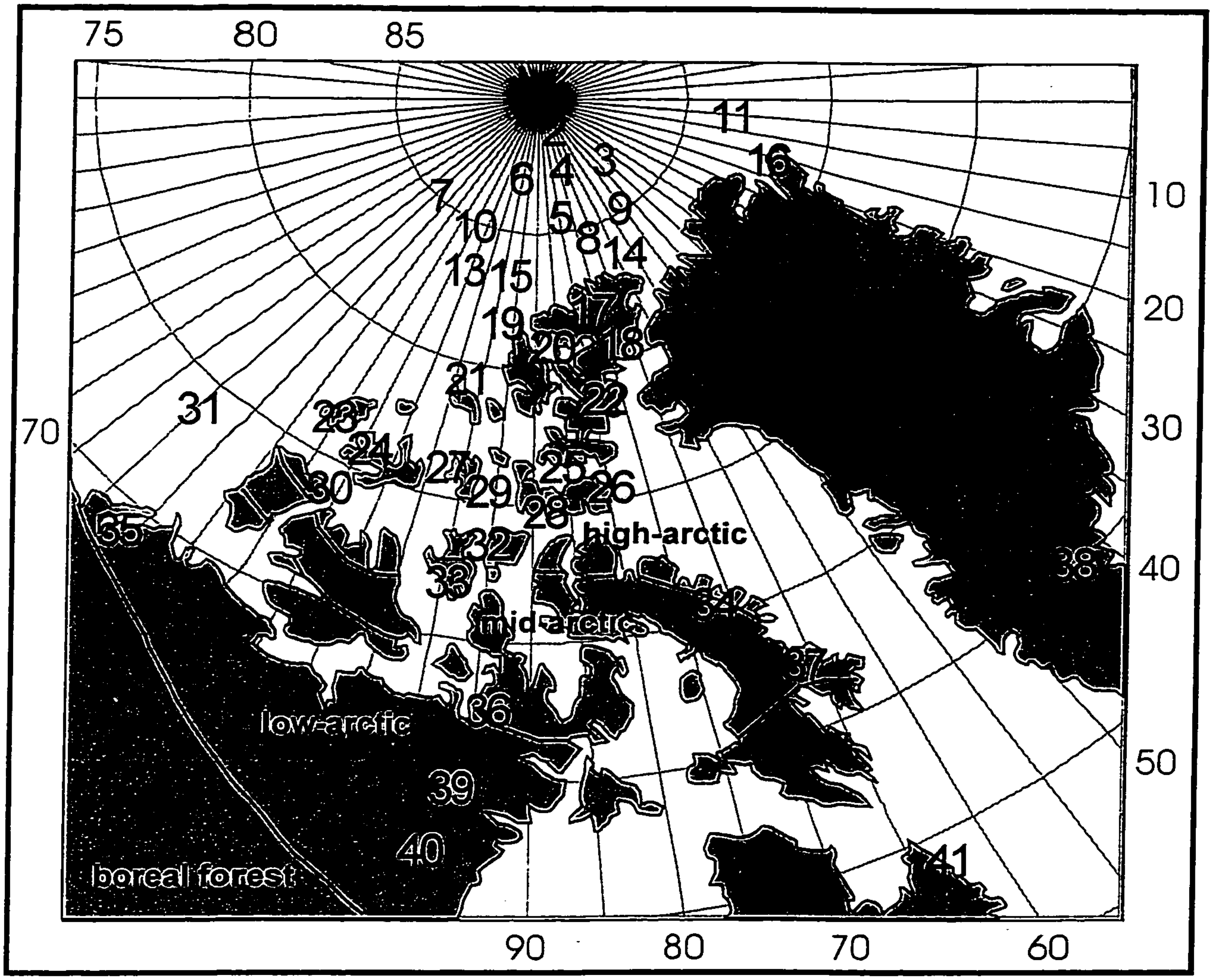

Figure 2.1. Location of sample sites. The boundaries for the high, mid, and low arctic are based on Polunin (1951). For other information related to sample sites, see Table 2.1.

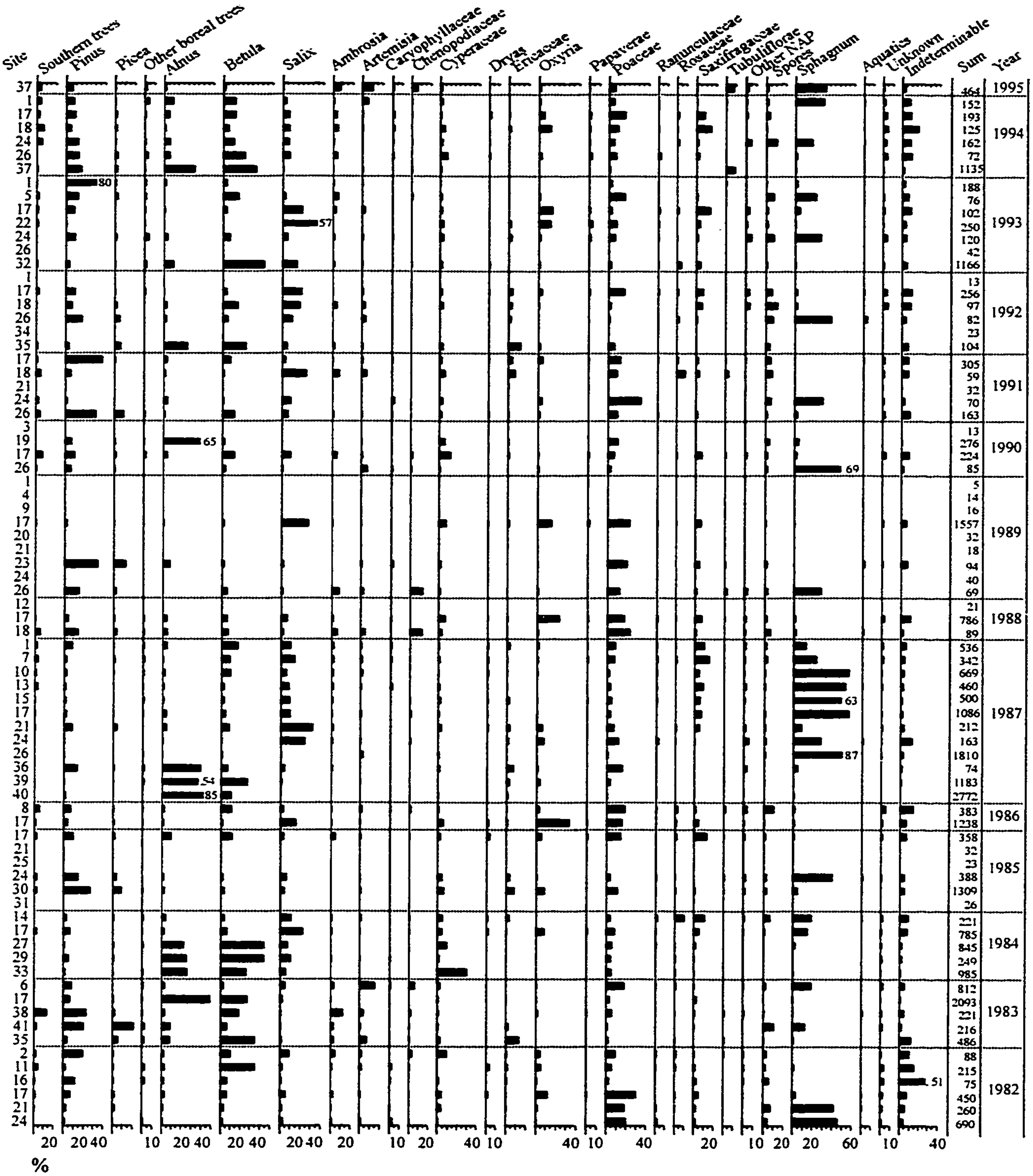

Figure 2.2. Pollen percentage diagram by year. For each year, sites sampled are listed with the most northerly site always at the top. Sample sites with pollen sum of <50 grains are listed, however, pollen percentages are left blank.

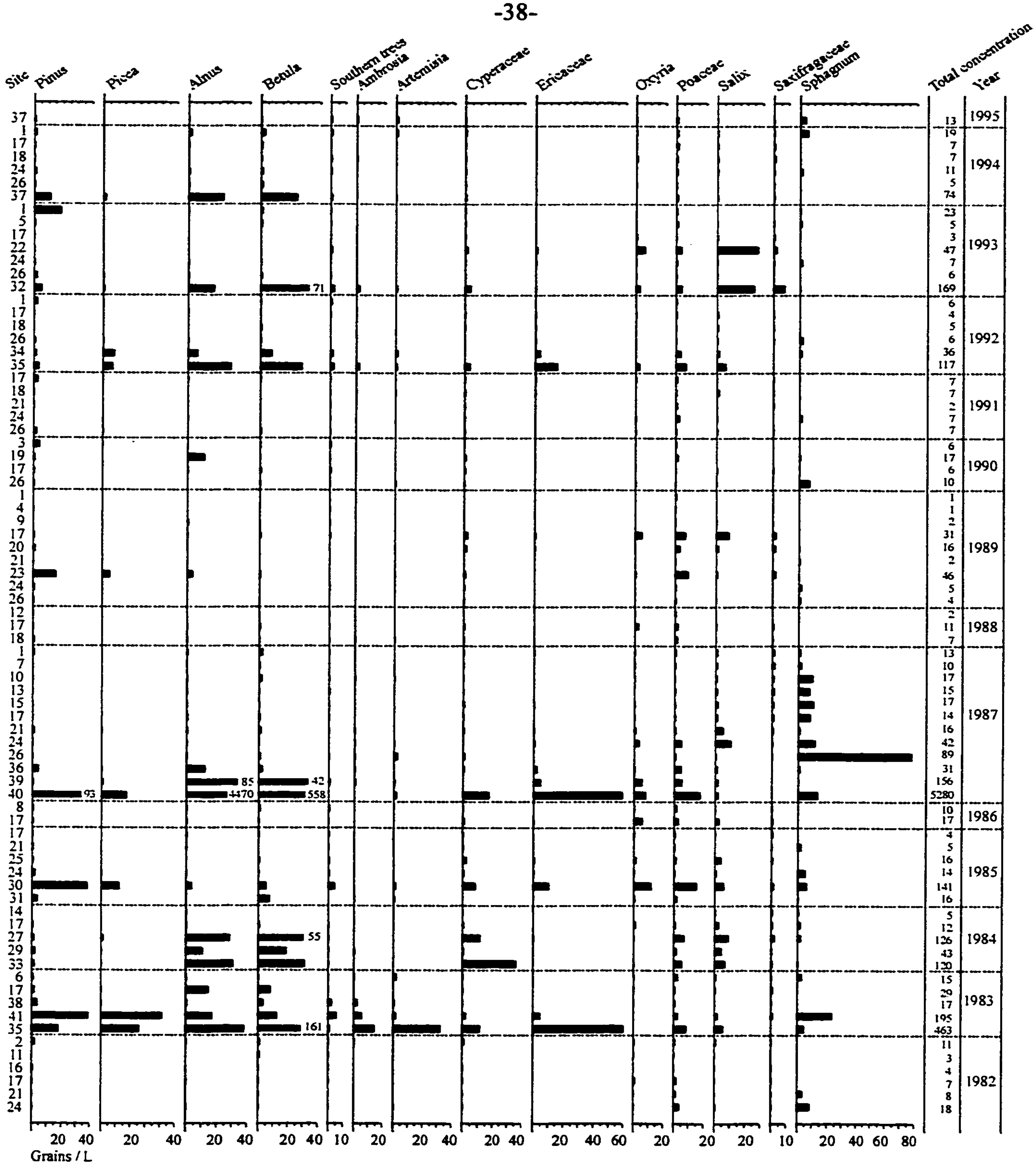

Figure 2.3. Pollen concentration diagram by year. Sites are sorted as in Figure 2.2.

Figure 2.4. Pollen percentage diagram sorted by environment. Samples with a sum <50 grains do not appear in the diagram. An average value was calculated for sites sampled several times.

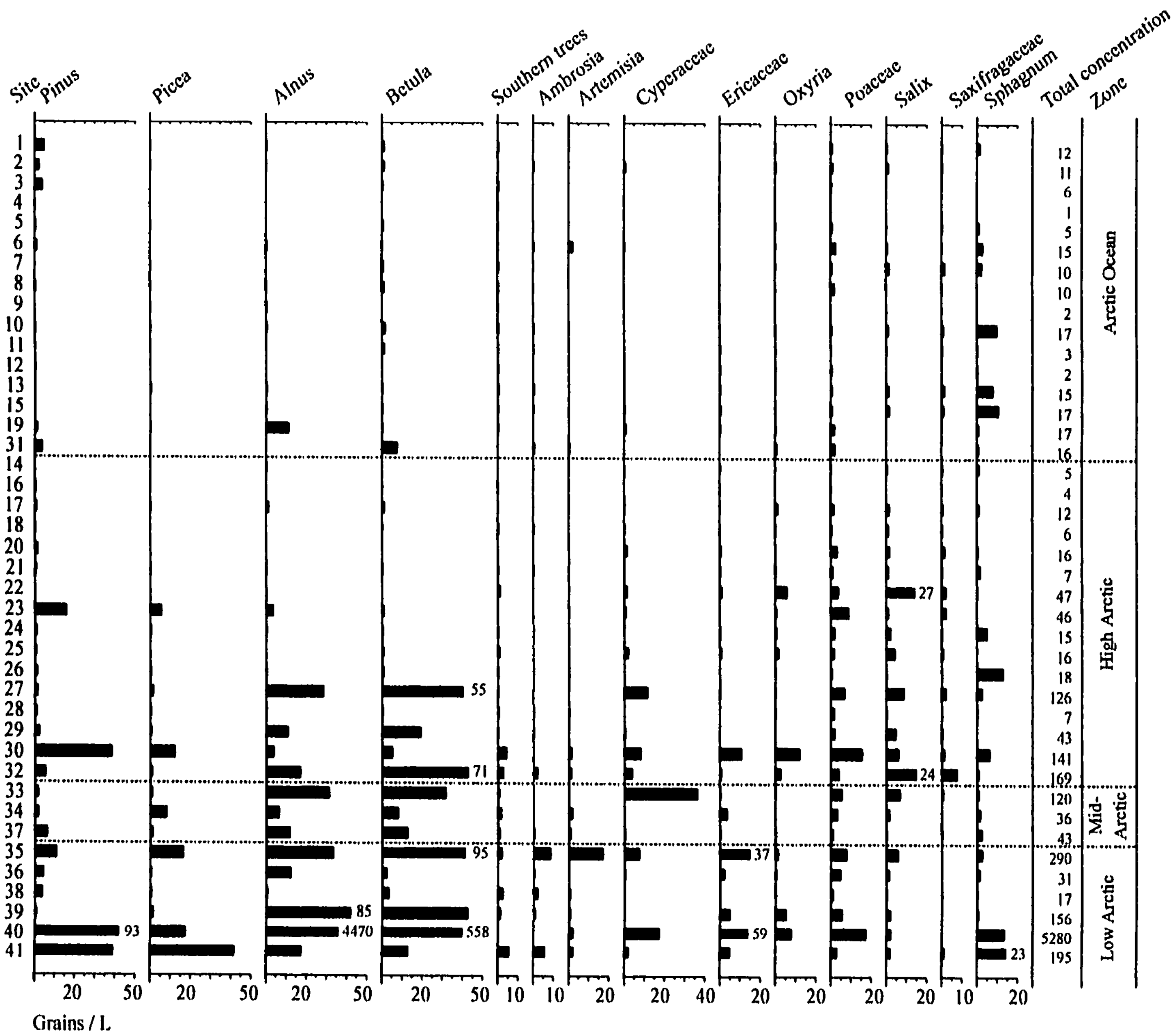

Figure 2.5. Summary pollen concentration diagram sorted as in Figure 2.4. All samples, regardless of pollen sum appear in the diagram. An average value was calculated for sites sampled several times.

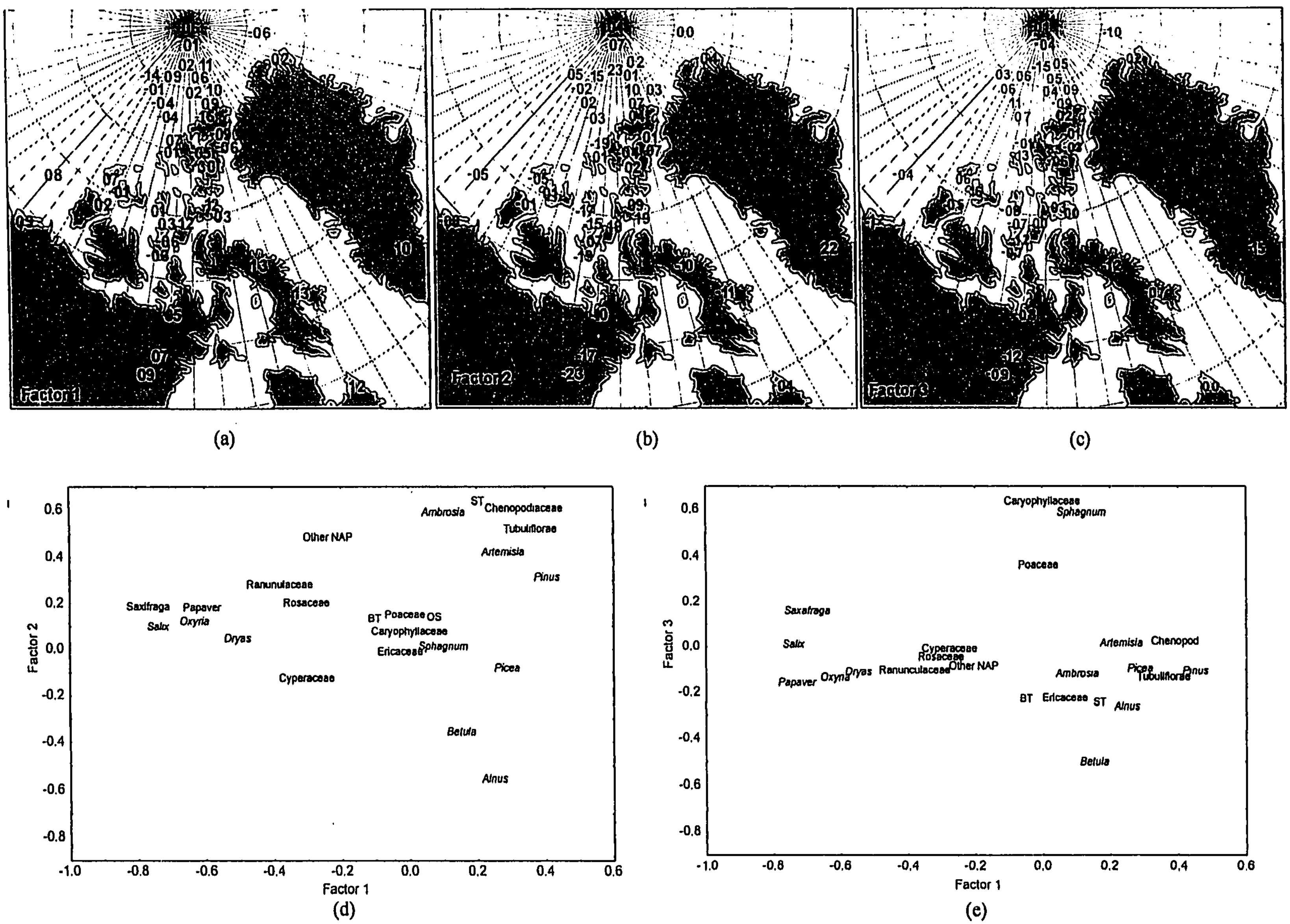

Figure 2.6. Principal components analysis of pollen percentages illustrated in Fig. 2.2. (a–c) Maps of the principal component scores. The scores have been multiplied by 10 for plotting. In locations where more than one sample were collected, only the mean score is plotted. (d,e) Plots of factor loading. Loading of factors 1&2 (d) and 1&3 (e). ST=southern trees (including deciduous and coniferous trees found today to the south of the boreal forest), BT= boreal trees not plotted separetly; OS=other spores; Other NAP=other non-boreal pollen; Saxifraga=Saxifragaceae; Chenopod=Chenopodiaceae.

CHAPTER 3

SEASONAL AND INTERANNUAL POLLEN VARIABILITY IN SNOW LAYERS OF ARCTIC ICE CAPS

Abstract

The pollen content of snow deposited at the summit of three ice caps and a small glacier in the Canadian Arctic and at one ice cap in the Russian Arctic was determined for periods ranging from one to thirteen years. On the ice caps, boreal forest trees and low arctic shrubs account, on average, for 26% to 49% of the total pollen. On the small glacier, the assemblages are characterized by about 90% tundra pollen types. *Sphagnum*, which is rare in the region, is sometimes one of the major components of the assemblages at these sites. The concentration of the various pollen types varies in the annual snow pack. The concentration of tundra pollen usually increases in the summer melt layer, but on some ice caps the highest concentrations are found in snow deposited after the melt season. Concentrations of tree/shrub pollen increase substantially in the late winter/spring snow layer and/or in the summer melt layer. The number of tree/shrub pollen reaching the ice caps varies considerably from year to year and this variability increases with decreasing distance to treeline.

3.1 Introduction

Numerous studies have reported on the variations of chemical species, isotopes, and insoluble microparticles in polar ice cores, but remarkably few studies have dealt with the biological components such as pollen, diatoms or bacteria (Fredskild and Wagner, 1974; McAndrews, 1984; Bourgeois, 1986; Kellogg and Kellogg, 1996; Andreev et al., 1997). These components have the potential of providing information on past environments in the vicinity of these ice caps and also, because they often originate from distant sources, contribute valuable information on past atmospheric circulation patterns (Fredskild, 1973, 1984; Nichols et al., 1978; Barry et al, 1981).

In order to interpret the biological components of ice cores, their transport and deposition processes in arctic environments must be understood. It is the objective of this paper to determine the composition of modern pollen assemblages derived from snow layers of arctic ice caps and to assess the interannual and seasonal variability at these sites. Comparing the sequences at these sites will also aid in documenting pollen dispersal in the Arctic.

Data are available for three ice caps and one glacier in northern Canada, and for one ice cap in northern Russia (Table 3.1; Fig. 3.1). In Canada, the sites are situated in the mountainous, eastern section of the Canadian Arctic Archipelago. Two of the sites, Agassiz Ice Cap and Quviagivaa Glacier (unofficial name), are on Ellesmere Island while the other two sites, Devon Ice Cap and Penny Ice Cap, are on Devon and Baffin islands respectively. The single site in Russia, Academy of Sciences Ice Cap, is located on Komsomolets Island, in the Severnaya Zemlya Archipelago. All four ice caps have been drilled to bedrock (e.g.

Koerner, 1977; Paterson et al., 1977; Fisher et al., 1983, 1995, 1998; Koerner and Fisher, 1990; Kotlyakov et al., 1991), and some of the ice cores were used for pollen analysis studies (Lichti-Federovich, 1975; McAndrews, 1984; Bourgeois, 1986; Koerner et al., 1988). Sections described in this study were collected from the walls of snow pits excavated in the vicinity of these borehole sites.

Studies have reported on modern pollen deposition in the snow of some of these ice caps (Lichti-Federovich, 1974; Short and Holdsworth, 1985; Bourgeois et al., 1985; Bourgeois, 1990; Alt and Bourgeois, 1995). The Devon Ice Cap record (Lichti-Federovich, 1974) covers one year of accumulation and is divided into six 10 cm increments. The Penny Ice Cap study (Short and Holdsworth, 1985) covers an eight year period and is based on a 6 m long firn core cut in approximately 0.5 m segments, without exact knowledge of the time-depth relationship. On the Agassiz Ice Cap, average pollen concentrations were obtained from a 5 m core (Bourgeois et al., 1985). They were also obtained from the wall of a pit where each year of accumulation was divided into four increments based on the stratigraphy of the snow. An eight year record was presented (Bourgeois, 1990) and the summer record was later extended by another two years (Alt and Bourgeois, 1995). These records cannot be readily compared as they vary in length, number of years, number of increments sampled, and presentation of the data. However, they do indicate that ice core palynology has the potential to provide valuable information about past environments.

The pollen concentration is extremely low on these arctic ice caps, ranging from approximately 10 to 100 grains per litre, whereas on temperate glaciers the concentration can be more than one thousand grains per litre (Vareschi, 1935; Godwin, 1949; Heusser, 1954; Ambach et al., 1966; Bortenschlager, 1970, Haeberli et al., 1983). As there is no local pollen contribution and the regional tundra pollen input is small, the pollen assemblages from the arctic ice caps are composed of a high percentage of long-distance transported pollen, primarily tree pollen from the boreal forest. The highest percentages were found on the Penny Ice Cap (Short and Holdsworth, 1985), which is the closest to treeline of the three ice caps.

3.2 The Region

Ellesmere and Devon islands are situated well within the high Arctic zone (Polunin, 1951) which is characterized by short, cold summers and severely cold winters. At low elevation, the melt period is generally less than 85 days, and the highest mean July temperatures (approx. 8°C) are recorded in the intermontane region of Ellesmere Island (Edlund and Alt, 1989; Bradley, 1990). The vegetation cover is discontinuous and lichens and mosses make up a significant percentage of the cover. Within the high Arctic, the vegetation types vary considerably and do not follow a north to south gradient. Indeed, the intermontane region of Ellesmere Island, protected from the northwesterly winds that prevail in the northern part of the archipelago in summer, has a much richer vegetation than does most of Devon Island to the south. This intermontane region supports over 100 species of vascular plants. It has been described by Edlund and Alt (1989) as an enriched prostrate shrub zone dominated by the shrubs *Salix arctica* and *Dryas integrifolia* with a diverse flora that includes Ericaceae, Asteraceae, Saxifragaceae, *Oxyria digyna*, *Epilobium latifolium*, a variety of sedges and grasses, and the occasional *Salix polaris* and *Salix reticulata*. On

Devon Island, a few narrow coastal zones around the ice cap have a similar vegetation type but the island generally has a poorer vegetation (<60 vascular plants) where dwarf shrubs are sometimes present but never dominant (Edlund and Alt, 1989). Baffin Island possesses the three sub-zones of arctic vegetation (high, mid, and low) of Polunin (1951). There is a particularly large climatic and vegetational gradient across the Cumberland Peninsula where the Penny Ice Cap is situated (Andrews et al., 1979). Dwarf birch (*Betula nana* and *B. glandulosa*) occurs to the south and west of the ice cap where the species reach their northern limit in the Eastern Canadian Arctic (Andrews et al., 1980a). *Artemisia borealis* also reaches its northern limit for Eastern Canada in the region (Porsild, 1957). Other dwarf shrubs (Ericaceae and *Salix*), sedges, and grasses are the dominant vascular plants in the region.

The Academy of Sciences Ice Cap, in the Russian Arctic, covers 69 % of the surface of Komsomolets Island. The unglaciated parts of the island support an extremely impoverished vegetation where vascular plants make up only a small percentage of the cover. Only seventeen species of vascular plants have been identified on the island, those are members of the Poaceae, Caryophyllaceae, Papaveraceae, Brassicaceae, and Saxifragaceae families (Aleksandrova, 1988). The central part of October Revolution Island, south of Komsomolets Island, has areas with greater plant cover and diversity. A total of seventy five vascular plant species have been identified on that island including three species of *Salix*, some Rosaceae, and also rare occurrences of *Artemisia borealis* (Aleksandrova, 1988).

On arctic ice caps, arboreal pollen types often represent 40% or more of the pollen deposited in the snow (Bourgeois et al., 1985). While the source area for these pollen grains is obviously not limited to the treeline, this nevertheless gives the minimum dispersal distances for some of those pollen grains. At 80°N, the Russian site is approximately 900 to 1,000 km from the treeline to the south of the Taymyr Peninsula. The Agassiz Ice Cap on Ellesmere Island, also at 80°N, and the Devon Ice Cap, at 75°N, are respectively about 2,100 km and 1,550 km from the treeline in north-west Canada. On Baffin Island, the closest treeline is in north-eastern Canada and it is approximately 1,000 km from the Penny Ice Cap. In Arctic Canada, the northern treeline is formed by spruce (*Picea*) whereas in Russia it is formed mainly by spruce in the west, by larch (*Larix*) and spruce in central Siberia and by larch in the east. Pine (*Pinus*), an important component of the pollen rain in the Arctic, grows from approximately 100 km to 800 km to the south of the treeline in Canada. In western Russia, the northern limit of pine is close to treeline and in the far east, pine forms a shrub conifer treeline. Alder (*Alnus*) and dwarf birch (*Betula*) grow north of the treeline, the former reaches the southern part of Baffin Island.

3.3 Study sites

The choice of ice caps and the sampling programme was usually determined by the needs of other glaciological investigations, such as ice core drilling, snow chemistry studies or glacier mass balance surveys. As a result, the length of the pollen records vary and some have discontinuities. However, with one exception, all sample sites are in the accumulation zone of the ice caps and, although a certain amount of summer melt takes place, the melt-water does not leave the snow pack. Therefore, the records should be complete for the period covered.

Agassiz Ice Cap (Ellesmere Island)

On the Agassiz Ice Cap, samples were collected from two sites, both situated in the northern part of the ice cap. They are referred to as Agassiz 1 and Agassiz 2, the former being the principal site of this investigation. Agassiz 2 (1,730 m a.s.l.) is at the top of a dome while Agassiz 1 (1,670 m a.s.l.) is about 1 km downslope. The sites are relatively close to each other, but the total annual accumulation differs as a result of snow drifting (scouring) which removes snow from the top of the dome during the winter months (Fisher et al., 1983, 1995).

Agassiz 1 has the longest record presented in this study, covering the period from the late summer of 1981 to the spring of 1994. Each annual snow layer was divided into four increments. In 1986, samples were collected from the surface to the bottom of a 3-m pit which exposed the years 1981 to 1986. After 1986, samples were collected each spring, from the surface to the bottom of the previous summer layer. The data for this site have been in part presented elsewhere (Bourgeois, 1990; Alt & Bourgeois, 1995; Gajewski et al., 1995). The sampling interval for the Agassiz 2 site is coarser than at Agassiz 1 site. Two snow layers, the summer melt layer and the remaining accumulation, were collected per year. The record covers the summer of 1990 to the spring of 1994. However, the snow layers from the summer of 1992 to the spring of 1993 were not sampled at this site.

Quviagivaa Glacier (Ellesmere Island)

Quviagivaa Glacier is a small valley glacier in the Sawtooth Range on the Fosheim Peninsula of Ellesmere Island. It was sampled in late spring of 1993 during a hydrometeorological investigation of the glacier. The glacier is always near the equilibrium line which suggests that some years it might lose all of the previous winter accumulation (Wolfe, 1995). The pollen record covers only one year but is of interest because the site is at a relatively low elevation (750 m a.s.l.) and is surrounded by a relatively rich tundra vegetation.

Devon Ice Cap (Devon Island)

Snowpit pollen samples have been collected in the high elevations (1,710 m a.s.l.) of the Devon Ice Cap since 1990. The samples were usually collected during the mass balance surveys of the ice cap and were limited to two layers per year. The record covers the summer of 1989 to the summer of 1994 with one increment missing (summer of 1992). A one year record, reported by Lichti-Federovich (1974), came from the same area.

Penny Ice Cap (Baffin Island)

Samples for the present study were collected in the vicinity of a deep borehole drilled in 1995 at an elevation of 1,900 m a.s.l. It is 2 km away from the site previously sampled by Short and Holdsworth (1985) in the summit region of the Penny Ice Cap. Samples were collected on two occasions, in the spring of 1992 and in 1995. In 1995, samples covering approximately 1.5 year of accumulation were collected from the wall of a pit used for snow chemistry studies (Grumet et al., 1998). Each year was divided into four increments, and the data cover, with some gaps, the summer of 1991 to the spring of 1995.

Academy of Sciences Ice Cap (Komsomolets Island)

The Academy of Science Ice Cap is the largest ice cap in the Severnaya Zemlya Archipelago of the Russian Arctic. In the spring of 1993 and 1994, samples were collected in the summit area (810 m a.s.l.) for pollen analysis and snow chemistry (Bourgeois et al., 1997). The pollen record consists of four increments per year and covers the summer of 1992 to the spring of 1994.

3.4 Methods

The glaciological term "balance year" is used to label the annual layers. A balance year starts with the first snow accumulation of the season, usually in late July or August (depending on the year and the location of the ice cap), and ends the following summer once melting has stopped. Consequently, a balance year cuts across the middle of the pollen-producing season.

Sample collection

Each spring field season, snow pits were excavated to cover at least the current year of accumulation and the previous summer melt layer underneath. When possible, 20 to 30 kg of snow were collected for each layer sampled in order to obtain an adequate pollen sum. Snow blocks were cut from a wall of the pit with a shovel and placed in thick plastic bags.

At Agassiz 1, Penny, Academy of Sciences and Quviagivaa sample sites, four layers were sampled per balance year (Fig. 3.2). They are labeled, from bottom to top, the *a*, *b*, *c*, and *d* layers, as they were in a previous study (Bourgeois, 1990). The thickness of these increments was determined from the snow stratigraphy observed from the pit wall. The lowermost layer of a balance year, *a*, is formed of highly metamorphosed, loose snow grains. This distinctive layer is found immediately above the summer melt surface. On the Agassiz Ice Cap, the thickness of the *a* layer can sometimes be as little as 10 cm, but on the Penny Ice Cap, which is further south and receives more accumulation, it can be as much as 40 cm thick. The usually thicker, fine-grained winter snow that overlays the *a* layer was divided into two equal parts (from bottom to top, *b* & *c*), unless a clear boundary could be identified. Although no particular time of deposition can be assigned to these layers, we can assume that the *b* layer contains snow deposited in autumn/early winter and the *c* layer contains late winter/spring snow. The melt layer which completed the balance year and was always sampled the following spring, is labeled the *d* layer. It is usually formed of dense, icy, granular snow. When the meltwater penetrates the layers underneath and refreezes, it may form ice lenses. The upper limit of the *d* layer is normally easily identified but it is often more difficult to assess the lower limit because it consists of meltwater that percolates into the annual snow pack and refreezes. The sampling of this increment was usually limited to a 10-20 cm thickness in order to avoid sampling snow already collected the previous spring as part of the *c* layer. In the discussion that follows, the *a-d* layers are sometimes associated with seasons. In a very broad sense, the *a* layer represents the late summer/early autumn; the *b* layer, the autumn/early winter; the *c* layer, the late winter/early spring; and the *d* layer, the late spring/summer.

At the Agassiz 2 and Devon sites only two snow layers were sampled per balance year. One layer consists of the snow accumulated from the late summer/early autumn to the following spring (labeled *a-c*), and the other consists of the melt layer *d*. The length of the records and the number of increments per year are summarized in Figure 3.2.

Processing

Before melting the samples in a field laboratory, a known number of *Lycopodium* spores was added to each sample to assess the number of pollen grains lost during field and laboratory procedures. The samples were melted inside their bags and the water transferred to a pressure vessel. Using a low pressure, the meltwater was filtered through an 8 μm pore-size cellulose filter. The filter was kept in cold storage until further processing. In the Ottawa laboratory, the filters were soaked in hot hydrofluoric acid, rinsed with water followed by glacial acetic acid and then left for fifteen minutes in an acetolysis solution before again rinsing with water. This method does not involve centrifuging the sample. The pollen residue remains on the filter, and at the end of the procedure the filter is cut in half and mounted on two microscope slides for analysis.

Pollen analysis

The entire pollen content of the sample was analyzed under the microscope. A correction factor based on the percentage of *Lycopodium* spores recovered in the sample was applied to account for the loss of pollen and the pollen concentration diagram (pollen grains per litre of melted sample) is based on the corrected values. When sampling the snow layers, the aim was to obtain a minimum of 100 pollen grains per sample. However, at some sites the concentration was much lower than expected and smaller pollen sums were obtained.

Several pollen types, present only sporadically or in low numbers, were grouped. The main pollen types included under "Southern trees" are those of *Acer*, *Fraxinus*, *Quercus*, and *Ulmus*. "Other boreal trees" include the rare *Populus, Abies* and *Juniperus* pollen grains. The latter, although a shrub, was included with the trees in order to simplify the data. Some of the recurrent types listed under "Other NAP" include Brassicaceae, Liguliflorae, Fabaceae, and *Pedicularis*. Listed under "Spores" are the fern taxa, while the "Aquatics" are represented by the very occasional grain of Nymphae, *Potamogeton*, *Sparganium*, and *Typha*.

3.5 Results

Agassiz Ice Cap

The pollen assemblages of the Agassiz 1 site are dominated by *Pinus*, *Alnus*, *Betula*, *Salix,* Poaceae, Saxifragaceae, and *Oxyria* (Fig. 3.3). With the exception of *Salix*, which is found in the regional tundra, the tree and shrub taxa are associated with the boreal forest and low arctic tundra. *Pinus*, *Alnus*, and *Sphagnum* are the most variable taxa and can represent up to 60-80% of the total pollen.

The diagram (Fig. 3.3) may be divided into three sections: a lower section (81-82*a* to 86-87*b*) dominated by *Pinus*, *Alnus*, and *Sphagnum*; a middle section (86-87*c* to 90-91*b*) where these taxa decrease by comparison to tundra species, and an upper section characterized by higher percentages of *Pinus* pollen. This is best seen in the summary record of pollen concentration (Fig. 3.5).

Some conclusions can be made about the seasonal variations. Higher percentages of *Pinus*, *Picea*, *Alnus*, and *Betula* pollen are most often associated with the *d* layer (late spring/summer) or, to a lesser extent, the *c* layer (late winter/early spring). Southern tree types are found in small numbers throughout the annual snow pack, but they tend to peak in the *c* layer. Most of the herbaceous pollen types could have been derived from plants growing in the Canadian high arctic tundra. However, a few types such as *Ambrosia, Artemisia,* and Chenopodiaceae are exotic to the region, although they are regularly part of the pollen assemblages. *Ambrosia*, the most common of the exotic herb species, and *Artemisia,* tend to peak in the *b* (autumn/early winter) or *c* layers. *Salix* and the other dominant tundra types such as the Cyperaceae, *Oxyria*, and Poaceae, and Rosaceae almost always peak in the *a* layer (late summer/early autumn) or, less often, in the *d* layer. Saxifragaceae, mainly represented by *Saxifraga oppositifolia*, is common throughout the record but, contrary to other tundra types, it reaches a maximum in the *c* or *b* layers. *Sphagnum* does not grow in the region, but the spores are found regularly in the record. Like the fern spores, it peaks mostly in the *b* layer. Surprisingly, pollen of aquatic and semi-aquatic plants are found sporadically in the record but in too small numbers to detect any seasonal trend.

The average total pollen concentration is 13.5 grains/litre, but there are considerable variations (between 2-90 grains/litre) in the record (Fig. 3.4). The most obvious difference (Fig. 3.5) is the low tree pollen concentration throughout the years 1984 to 1989 and the larger seasonal variations recorded prior to 1983 and again after 1989. In the upper part of the record, maximum tree pollen concentrations vary between 6.5-15 grains/ litre by comparison to <5 grains/ litre in the middle section and 18-83 grains/ litre, respectively, for 1982 and 1983. The herbaceous pollen follows a different trend with the highest concentrations between 1985 and 1990 during the time when tree pollen concentrations are generally low.

Boreal tree and shrub taxa tend to peak in the spring or summer layers. However, a peak in one taxon may or may not be accompanied by a peak in the others. Typically, the boreal trees, *Pinus* and *Picea,* tend to increase in the same year, while the shrubs *Betula* and *Alnus* sometimes peak in other years. Southern trees and *Ambrosia*, the most common of the exotic herb species, show similar irregular variations throughout the record.

The concentration of herbaceous pollen usually peaks in the *a* layer (late summer/early autumn) or the *d* layer (late spring/summer). The highest values were obtained in the 1988-89 *a* layer followed by the 1989-90 *d* layer and 1985-86*a* layer. Several taxa increased in those layers, the major types being *Salix*, Poaceae, and *Oxyria*. *Sphagnum* is found sporadically in the record with higher concentrations usually in the *b* layer (autumn/early winter). During the 1986-87 balance year, a large number of *Sphagnum* spores (17 grains/litre) were deposited at the surface of the ice cap; no other similar event has been registered.

Pinus, *Salix*, *Betula*, Poaceae, and *Oxyria* are the dominant pollen types in the Agassiz 2 record (Fig. 3.6). *Pinus* is the most variable pollen type, with percentages varying between 3 and 47%. The highest percentage was found in the 90-91*d* layer, as was also found in the Agassiz 1 record. Both tree and tundra pollen types have higher concentrations in the *d* layer (Fig. 3.7). However, in this record the *a* layer, which often has the highest concentration of tundra pollen, is not separated from the winter layers.

Quviagivaa Glacier

The Quviagivaa pollen percentage diagram (Fig. 3.8) is dominated by tundra species which may reflect the proximity of the site to the relatively rich tundra of the Fosheim Peninsula. *Salix* represents 40-72% of the total pollen. *Oxyria*, Poaceae, *Papaver*, Cyperaceae, and Saxifragaceae are other dominant tundra types.

The pollen concentration diagram (Fig. 3.9) shows a maximum in the *d* layer (late spring/summer) and a minimum in the *b* layer (autumn/early winter). The most abundant pollen type, *Salix*, reaches a maximum concentration of 214 grains/litre, nearly an order of magnitude higher than at Agassiz 1 site. Contrary to the Agassiz records, tree pollen are almost absent. When present, they are more likely to be southern tree types than boreal or low Arctic types.

Devon Ice Cap

The pollen percentage record for the Devon Ice Cap is dominated by *Pinus*, *Betula*, *Salix*, Poaceae and *Sphagnum* (Fig. 3.10). Arboreal pollen, especially *Pinus*, and *Sphagnum* spores are more abundant on Devon than on Ellesmere Island. Herbs that are exotic to Devon Island (*Ambrosia*, *Artemisia*, and Chenopodiaceae) also increase at some levels. In the uppermost sample, the exceptional increase of *Alnus* and, to a lesser extent of *Betula,* dwarfs all other pollen types. Throughout the rest of the profile, *Alnus*, like *Picea*, is not particularly well represented.

The most prominent feature in the concentration diagram (Fig. 3.11) is a very high concentration of *Alnus* (255 grains/litre) and *Betula* (106 grains/litre) pollen recorded in the summer melt layer of 1993-94 balance year. In the remaining part of the record the concentration of these two taxa never reach more than 2 grains/litre which is comparable to values obtained on Agassiz Ice Cap. Tundra species are poorly represented, both in numbers and diversity, which may reflect the poor vegetation cover found on most of Devon Island. *Sphagnum* spores are present in the lower half of the profile, and they generally have higher concentrations than those recorded on Agassiz Ice Cap.

Penny Ice Cap

The pollen record for Penny Ice Cap is short, and there is a gap from the summer of 1992 to the late summer of 1993. In three of the lower samples (Fig. 3.12) the pollen sum is low and the pollen percentages may be unreliable. In the upper half of the diagram, *Alnus* and *Betula* dominate the assemblages. Tubuliflorae, *Pinus*, and *Sphagnum* are also major components in that part of the record. *Salix*, surprisingly, is poorly represented in the diagram.

Pollen concentrations (Fig. 3.13) were highly variable, between 4 and 424 pollen grains/litre. The most striking feature in the record is the large increase of tree/shrub pollen in the *c* (late winter/early spring) and *d* (late spring/summer) layers of the 1993-94 balance year. *Betula* and *Alnus* dominate the pollen assemblages but *Pinus* and *Picea* also increase in those layers. *Artemisia*, Chenopodiaceae, Ericaceae and Tubuliflorae are slightly better represented on Penny Ice Cap than on the farther north sites. S*phagnum*, which is found sporadically in the record, reaches a concentration of 30 grains per litre in the 90-91*d* layer.

Academy of Sciences Ice Cap

In the pollen percentage diagram, *Pinus and Sphagnum* dominate the pollen assemblages (Fig. 3.14). In the 91-92*d* layer (late spring/summer), *Pinus* comprised 97% of the total pollen and in the layer immediately above it, 75%. *Sphagnum* dominates the upper half of the diagram, reaching a percentage of 70% in the 92-93*d* layer. With the exception of *Artemisia* (which grows on a neighbouring island) and Poaceae, the herbaceous species are not particularly well represented on Academy of Sciences Ice Cap.

Pollen concentrations (Fig. 3.15) decrease from a maximum of 598 grains/litre in the 1991-92*d* layer to a minimum of 5 grains/litre in the 1993-94*c* layer (late winter/early spring). The snow accumulation recorded in the spring of 1994 was twice the amount observed the previous spring, but this can only account for a small part of the high variability. The very high concentrations in the lower part of the diagram were driven primarily by *Pinus.* The concentration of *Sphagnum*, *Poaceae*, *Salix*, and the southern trees remained more or less constant.

3.6 Discussion

The snow layers of the four ice caps discussed in this study contain large percentages of tree and shrub pollen that are associated with the boreal forest and low arctic tundra. When pollen percentages of these tree/shrub types are averaged for each site, the lowest percentages are found on the Agassiz Ice Cap, the farthest north Canadian ice cap in this study. Average percentages of 26% and 31% were recorded for the two Agassiz Ice Cap sites compared to 42-49% for the other ice caps. At all sites, southern trees (e.g. *Acer*, *Quercus*, *Ulmus*) made up to 2-4% of the total pollen. *Sphagnum*, which is rare in the study area, is nevertheless a major component of the pollen/spore assemblages at all sites, particularly on Academy of Sciences Ice Cap where it accounts for 27% of the total pollen and spores.

At the Ellesmere Island sites, *Salix* is the dominant tundra taxon. On the small Quviagivaa Glacier, *Salix* makes up 54% of the total sum. At the above sites, *Oxyria*, Poaceae, Saxifragaceae, and Cyperaceae are the other major components of the regional pollen assemblages, as they are in pollen assemblages obtained from other high arctic locations (e.g. Ritchie et al., 1987; Gajewski, 1995). The relative representation of herbaceous species decreases on Devon and Penny ice caps, regardless of their more southerly location and, with respect to the latter site, much richer vegetation cover. Short and Holdsworth (1985) remarked on the relatively low amount of *Salix* and herbaceous tundra species in their pollen record from the Penny Ice Cap, which contrasted with pollen assemblages derived from moss polsters and fossil peat samples from the same area (e.g. Boulton et al., 1976; Andrews et al., 1979, 1980b). The percentage of herbaceous pollen types is also low on Academy of Sciences Ice Cap. On Devon, Penny and Academy of Sciences ice caps, the most common herbaceous type is Poaceae, but *Artemisia* and Tubuliflorae sometimes dominate the assemblages.

Seasonal variations are more pronounced in the pollen concentration than in the percentage records. The sequence of pollen assemblages in the annual snowpack reflects, in a broad sense, the flowering sequences of plants growing from probably 50 to over 2000 km from the sampling sites. The concentration of tree/shrub pollen types starts to increase in the spring snow layer and continues to do so in the summer melt layer. By the time summer melt has ceased on these ice caps, which can sometimes be as early as mid-July, the concentration of tree/shrub pollen in the snow has decreased substantially. The concentration of most herbaceous pollen types increases in the summer melt layer and, depending on site location, often reaches a maximum in the layer immediately above it. This pattern is most clearly seen at the Agassiz 1 site. On Penny Ice Cap, melting can persist until September, therefore fewer tundra pollen are incorporated in snow deposited after the melt season. *Sphagnum* spores usually peak in snow deposited after the tundra pollen maximum has been reached. On Penny Ice Cap and Academy of Sciences Ice Cap, pollen of the late-flowering plants *Artemisia*, *Ambrosia*, and Chenopodiaceae also peak in the late summer-autumn layers. Winter snow layers should have few pollen grains. However there is evidence for redeposition, particularly after a pollen event such as the one which occurred in the 1988 summer on Agassiz Ice Cap (Fig. 3.4). Slightly higher tundra pollen concentrations were registered afterwards thereby masking the following seasonal cycle.

Interannual variability

At sites where more than one year of data were obtained, the pollen concentration records showed significant interannual variability. Tundra pollen showed the most variability on Agassiz Ice Cap. The tree/shrub pollen concentrations on Agassiz Ice Cap show variability on two scales. The years between 1983-89 generally had lower tree/shrub pollen than years before and after. Superimposed on this trend of several years are interannual and seasonal trends. One unusually large peak of *Alnus* and *Betula* pollen was registered in the spring of 1983. These two taxa were also found to be responsible for the large interannual variations observed in the shorter Devon and Penny ice cap records. An extreme *Alnus* and *Betula* event was recorded at both of these sites in the spring/summer of 1994. However, this event was not registered as far north as Agassiz Ice Cap that particular year. Similarly, *Sphagnum* spores were found to vary considerably from year to year. In the summer/autumn of 1991, *Sphagnum* concentrations increased on the Penny Ice Cap and to a lesser extent on the Devon Ice Cap, but did not increase on the Agassiz Ice Cap. In 1986 a relatively large amount of *Sphagnum* spores did reach the Agassiz Ice Cap, and that particular year *Sphagnum* spores were present in the snow cover on the Arctic Ocean sea ice (Bourgeois, 1990).

At 80°N, the Academy of Sciences Ice Cap in the Russian Arctic is approximately at the same latitude as the Agassiz Ice Cap, but pollen assemblages and concentrations are more similar to those observed on the Penny Ice Cap situated at 67°N. On Academy of Sciences Ice Cap, *Pinus* pollen shows the greatest variability. The large spike of *Pinus* pollen, observed in the late winter/spring snow in 1993, was also found at other locations in the Russian Arctic that year as well in the snow covering the Arctic Ocean sea ice (Bourgeois et al., 1997). Because the data cover only two years it is difficult to assess representativeness of the record. However, Andreev et al. (1998) noted the high variability in the snow and ice of the Vavilov Ice Cap on the neighbouring October Revolution Island.

The presence of distinctive, periodic peaks of *Alnus* pollen associated with higher

Pinus and *Picea* pollen, were detected in sediment sequences from Baffin Island (Nichols et al., 1978; Andrews et al., 1979) at sites close to the Penny Ice Cap. These peaks were interpreted to be periods of higher frequencies of southerly air flows, most probably originating in Labrador-Ungava. However, the periodicity and the paleowind implications were later questioned by Barry et al. (1981). In view of the results obtained thus far from the snow layers of arctic ice caps, it would seem pertinent, as a follow-up to this study, to investigate the meteorological conditions during high and low pollen deposition years.

This study of pollen assemblages in snow layers of arctic ice caps represents a first step towards interpreting pollen records retrieved from ice cores. The data, which show the extent and variability of long distance atmospheric transport of pollen in the Arctic, may be relevant for the interpretation of pollen sequences retrieved from other types of sediment. The records presented here are short and some are highly variable. However, in the longer Agassiz Ice Cap record, some trends are emerging, and these could have some climatic significance.

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Table 3.1. Location of sites.

Site	Latitude	Longitude	Elevation (m. a.s.l.)
Agassiz Ice Cap 1	80°4' N	73°1' W	1670
Agassiz Ice Cap 2	80°4' N	73°1' W	1730
Quviagivaa Glacier	79°3' N	83°2' W	750
Devon Ice Cap	75°3' N	82°4' W	1710
Penny Ice Cap	67°2' N	65°5' W	1900
Academy of Sciences Ice Cap	80°4' N	95°0' E	810

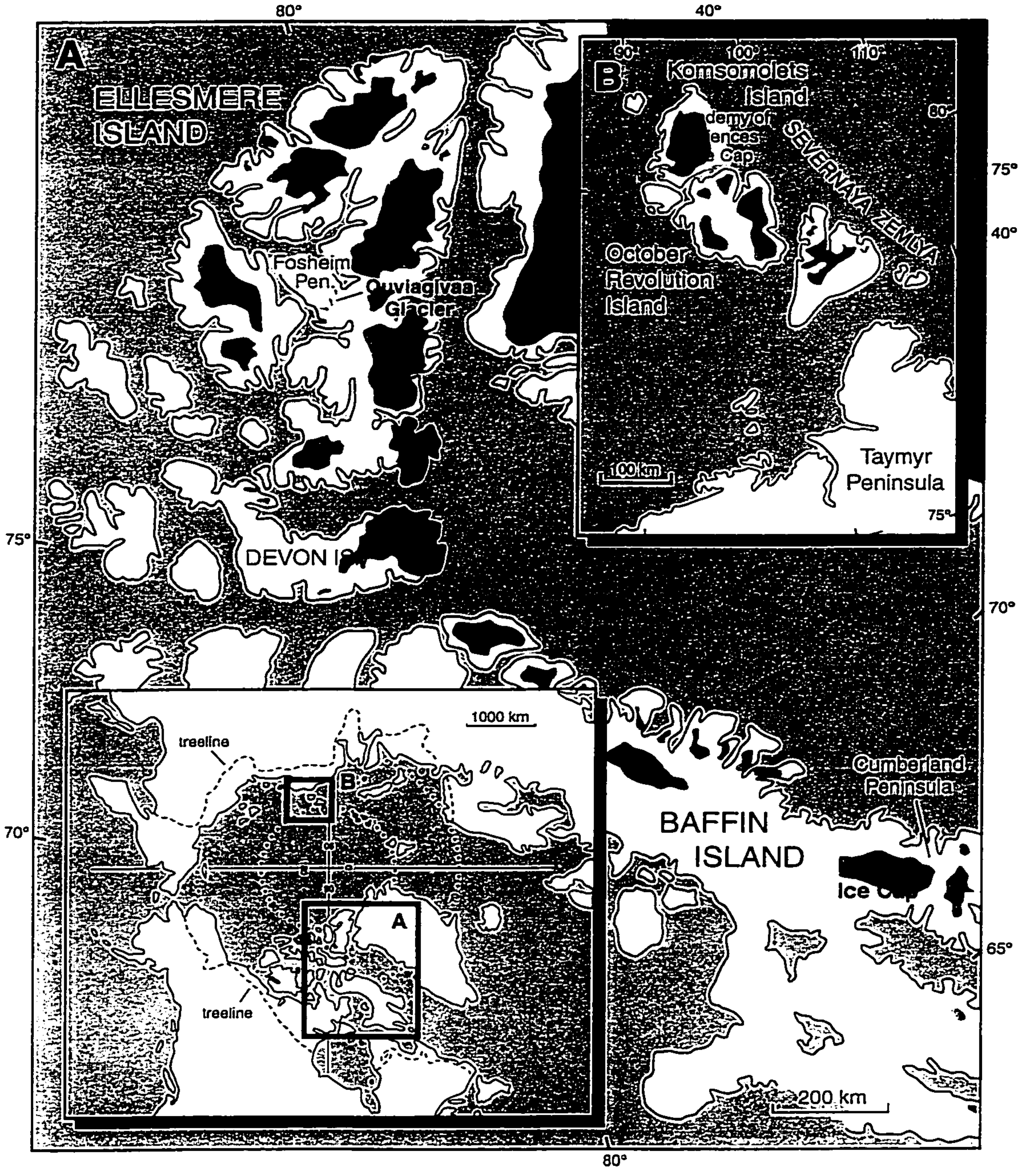

Figure 3.1. Location of sample sites.

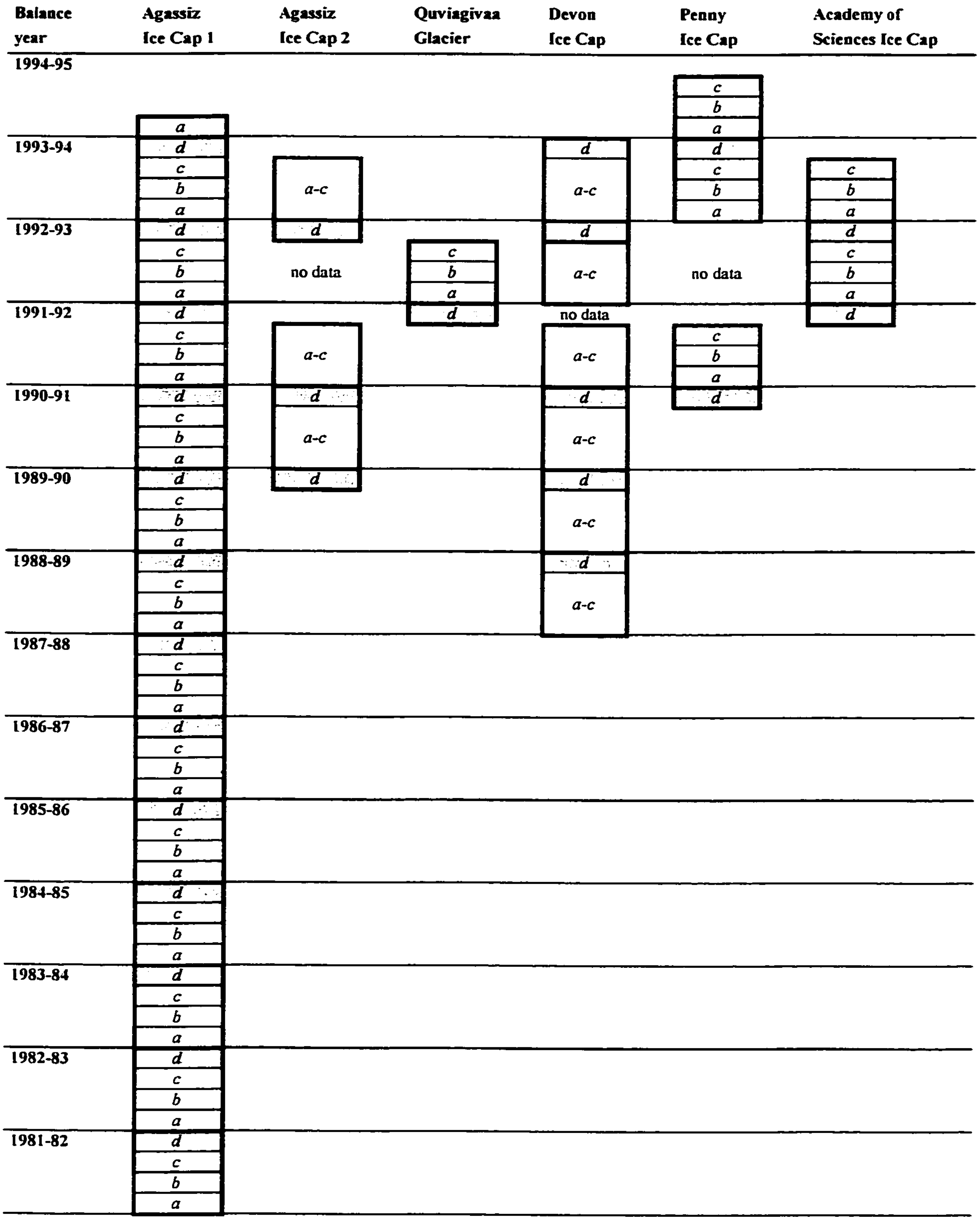

Figure 3.2. Annual increments and number of layers sampled at each site. In this study, the balance year starts with the "a " layer (which covers the first accumulation after the melt season) and ends with the "d" layer (which covers the melt events of the following summer).

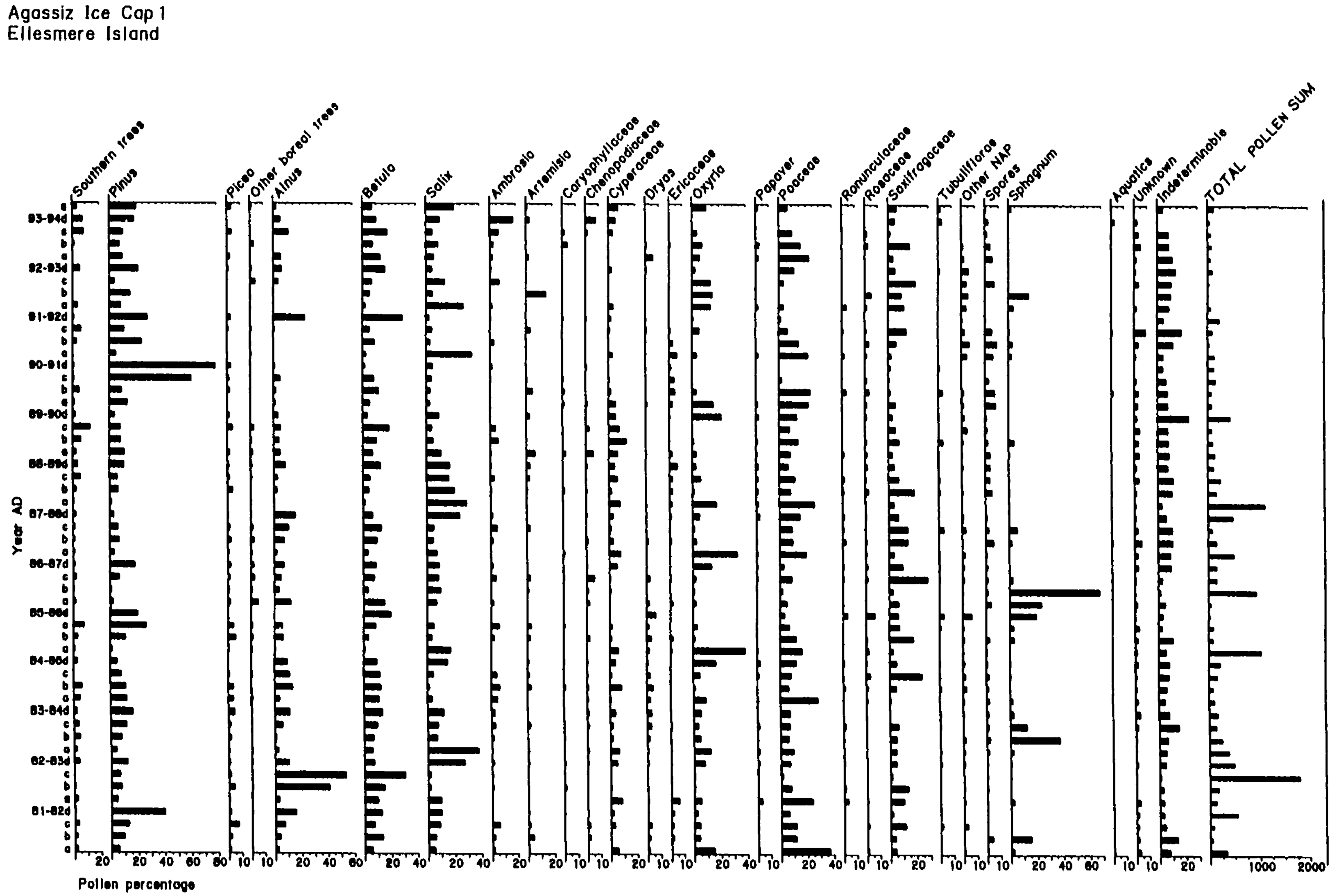

Figure 3.3. Pollen percentage diagram for Agassiz Ice Cap 1.

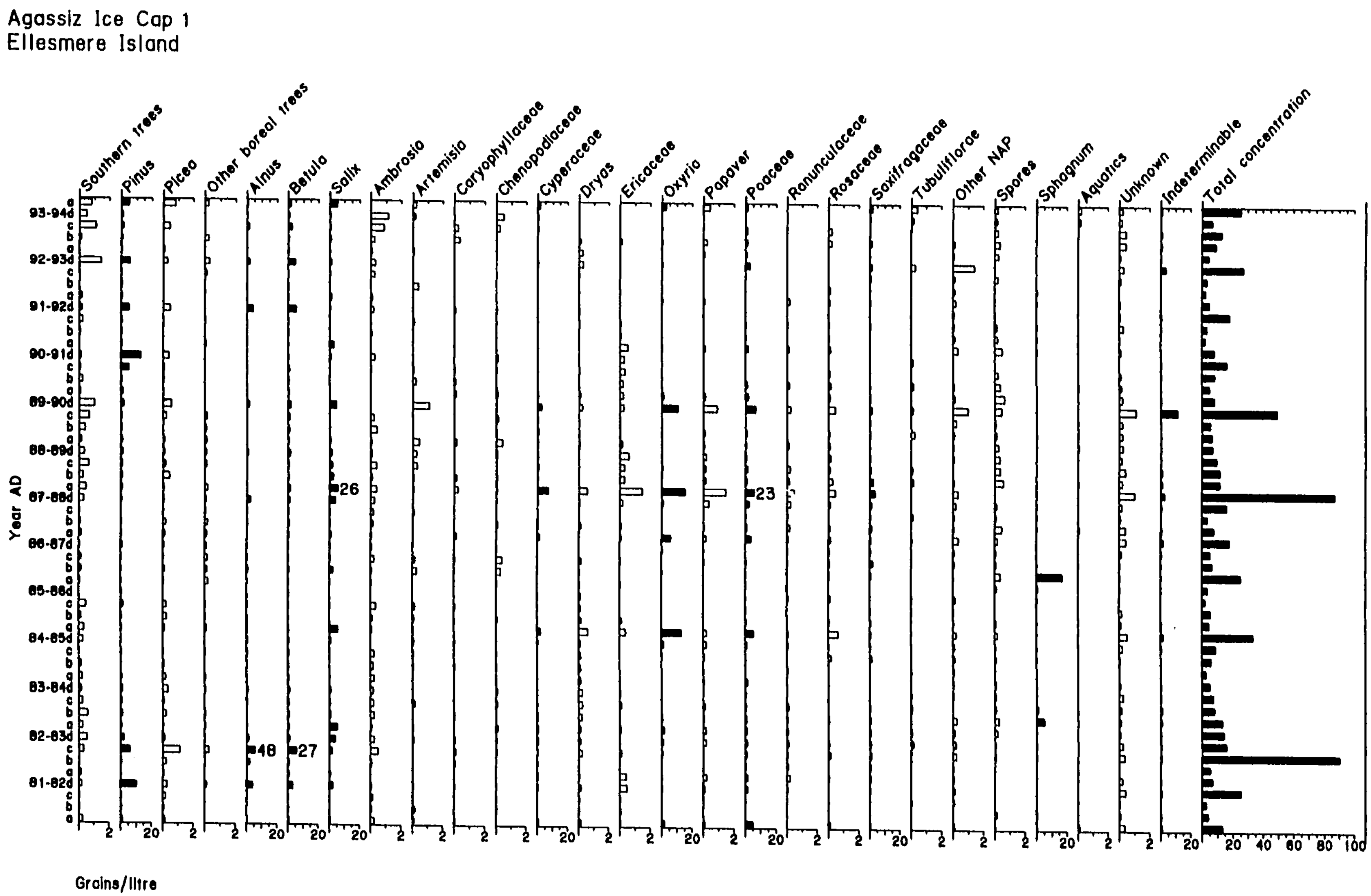

Figure 3.4. Pollen concentration diagram for Agassiz Ice Cap 1. Note the change in scale.

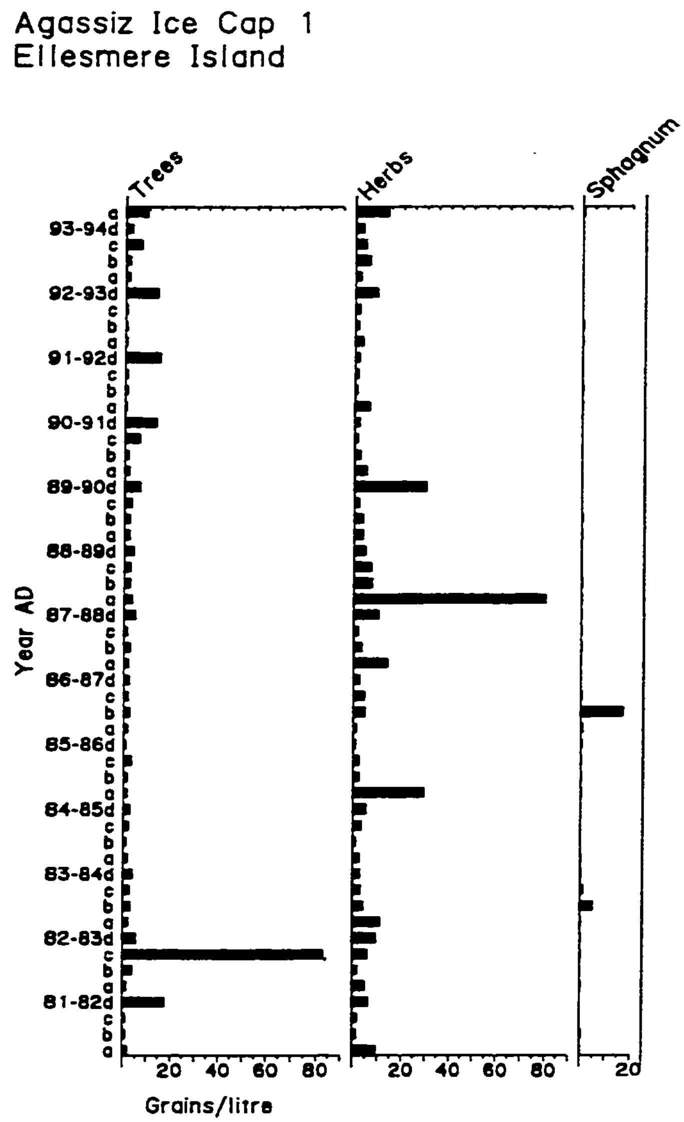

Figure 3.5. Summary concentration diagram for Agassiz Ice Cap 1. The herb column includes all taxa, regardless of origin. In this study, *Salix* is considered a tundra type and is included with the herbs.

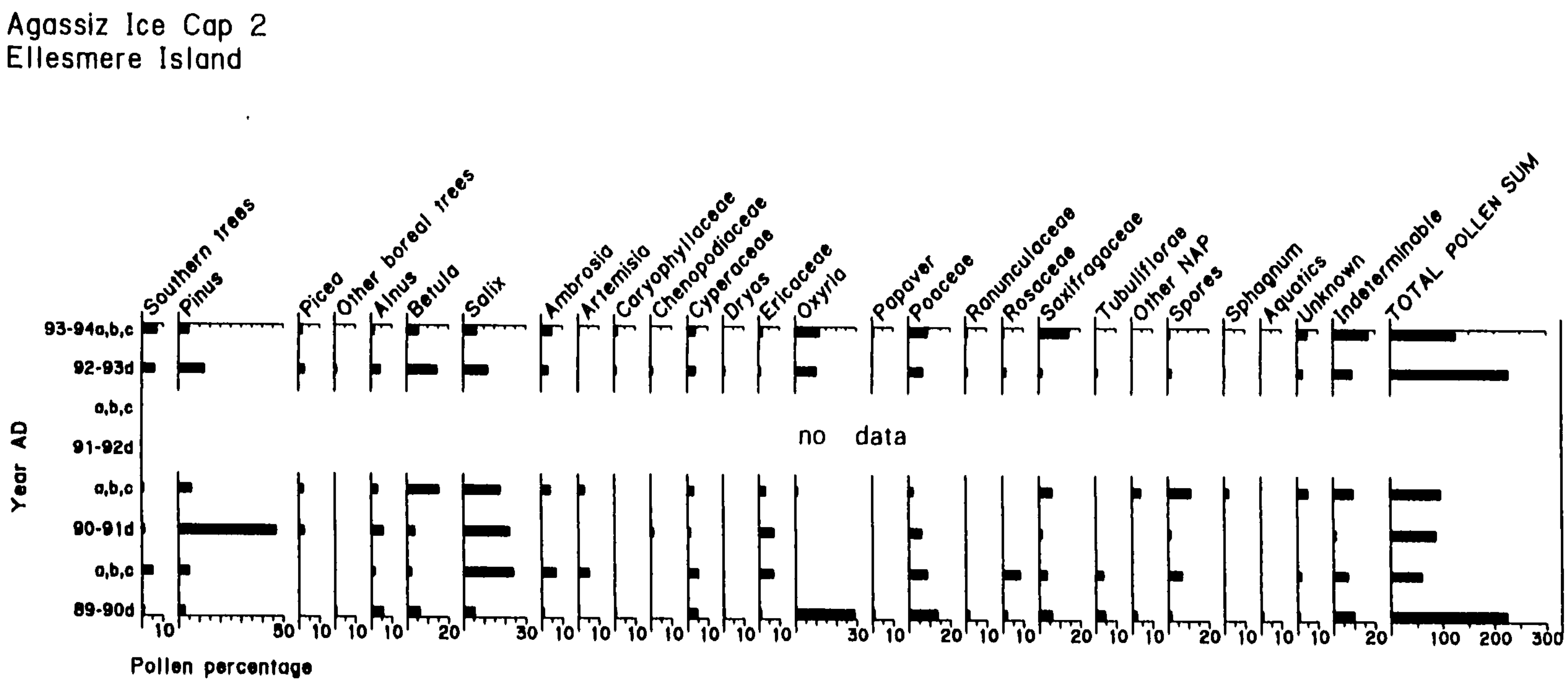

Figure 3.6. Pollen percentage diagram for Agassiz Ice Cap 2.

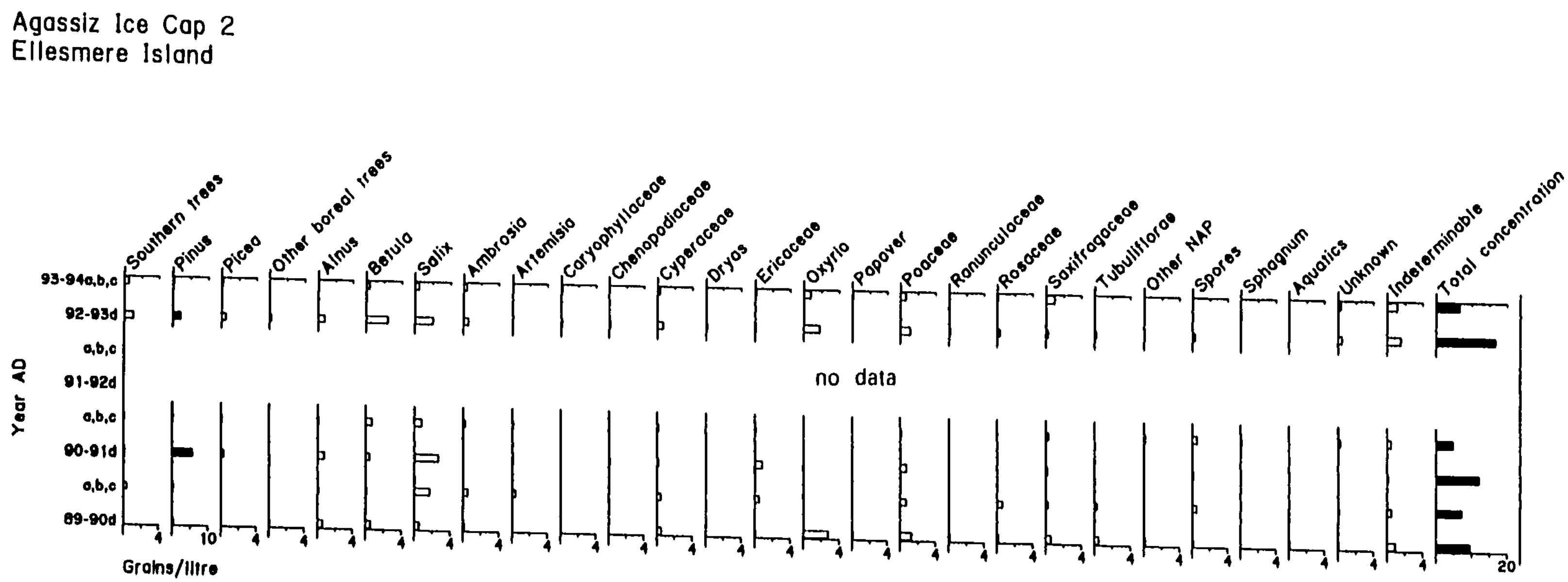

Figure 3.7. Pollen concentration diagram for Agassiz Ice Cap 2. Note the change in scale.

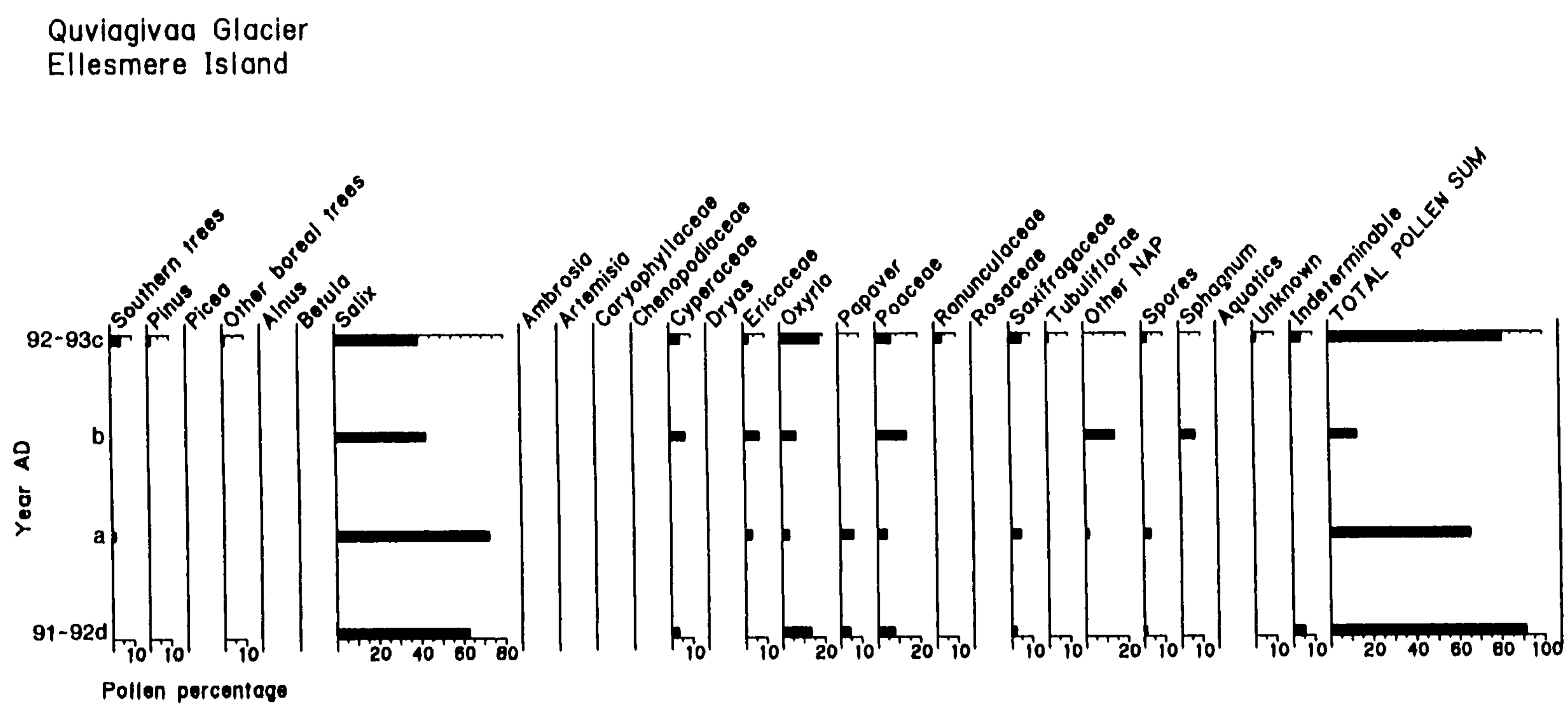

Figure 3.8. Pollen percentage diagram for Quviagivaa Glacier.

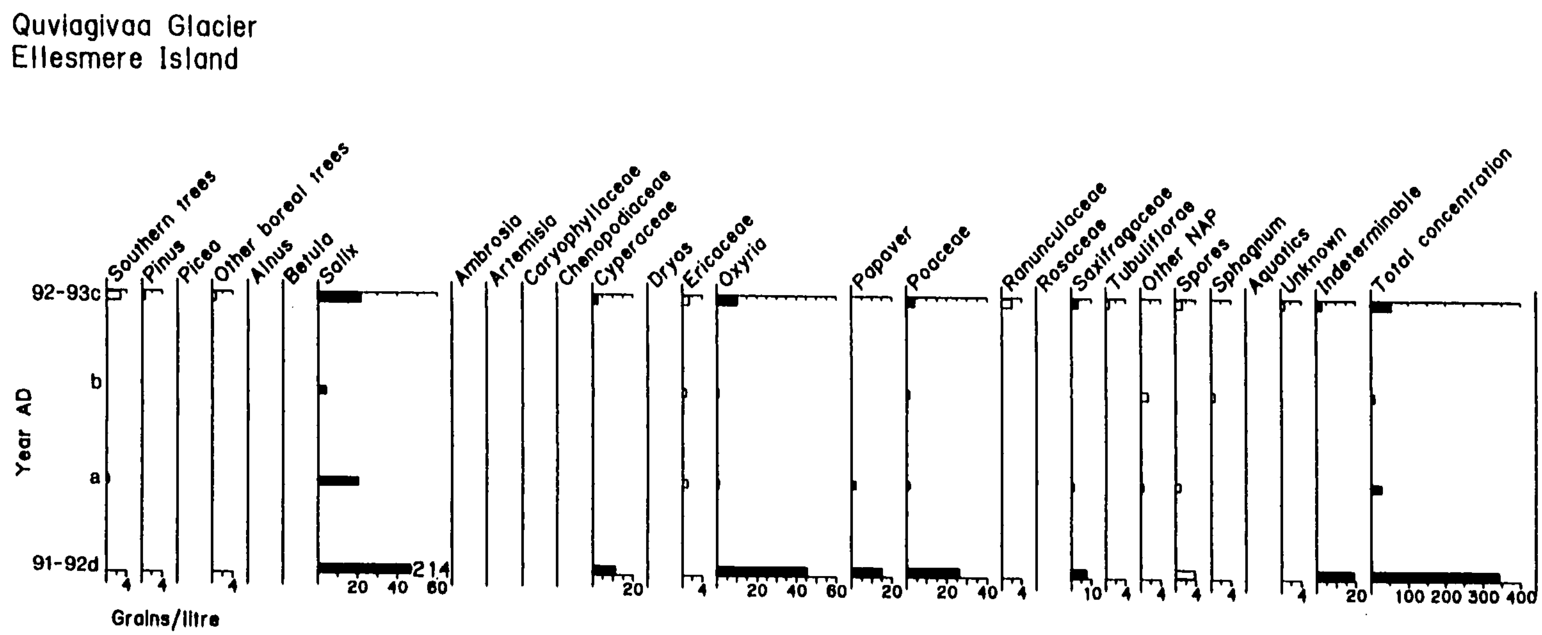

Figure 3.9. Pollen concentration diagram for Quviagivaa Glacier. Note change in scale.

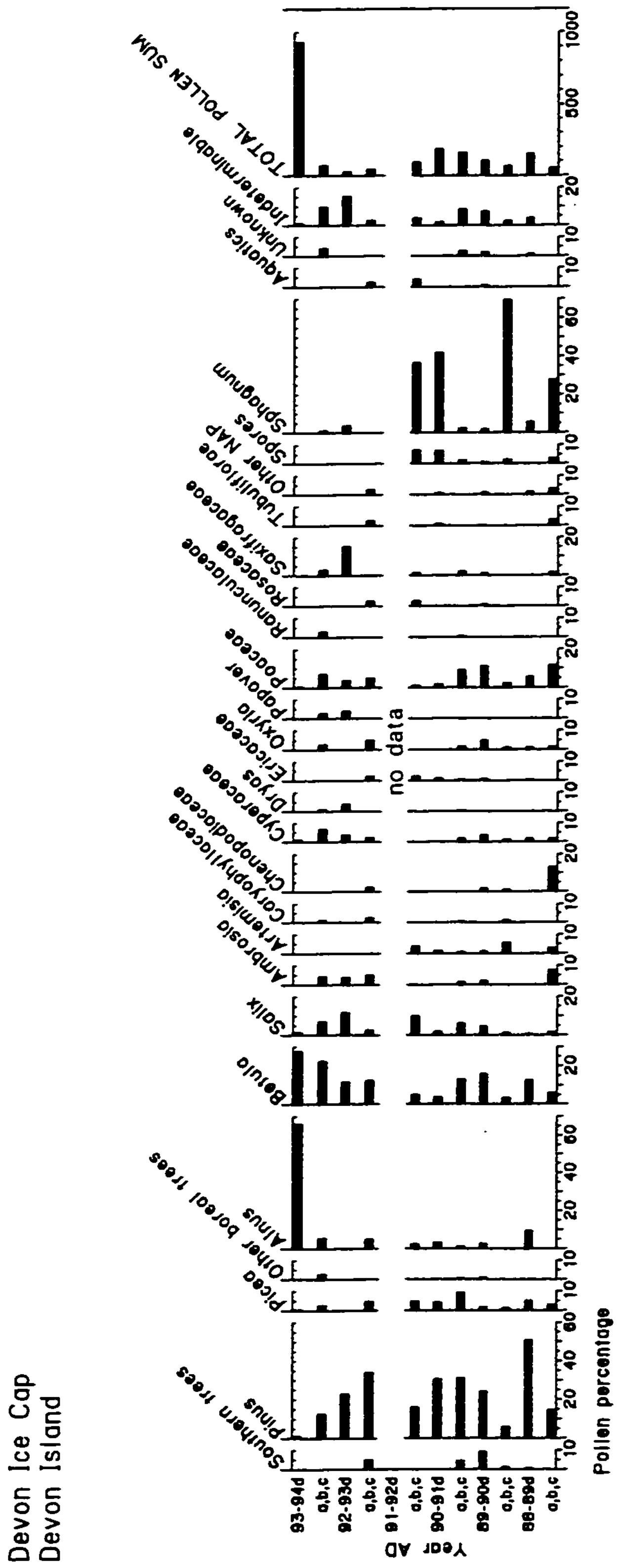

Figure 3.10. Pollen percentage diagram for Devon Ice Cap.

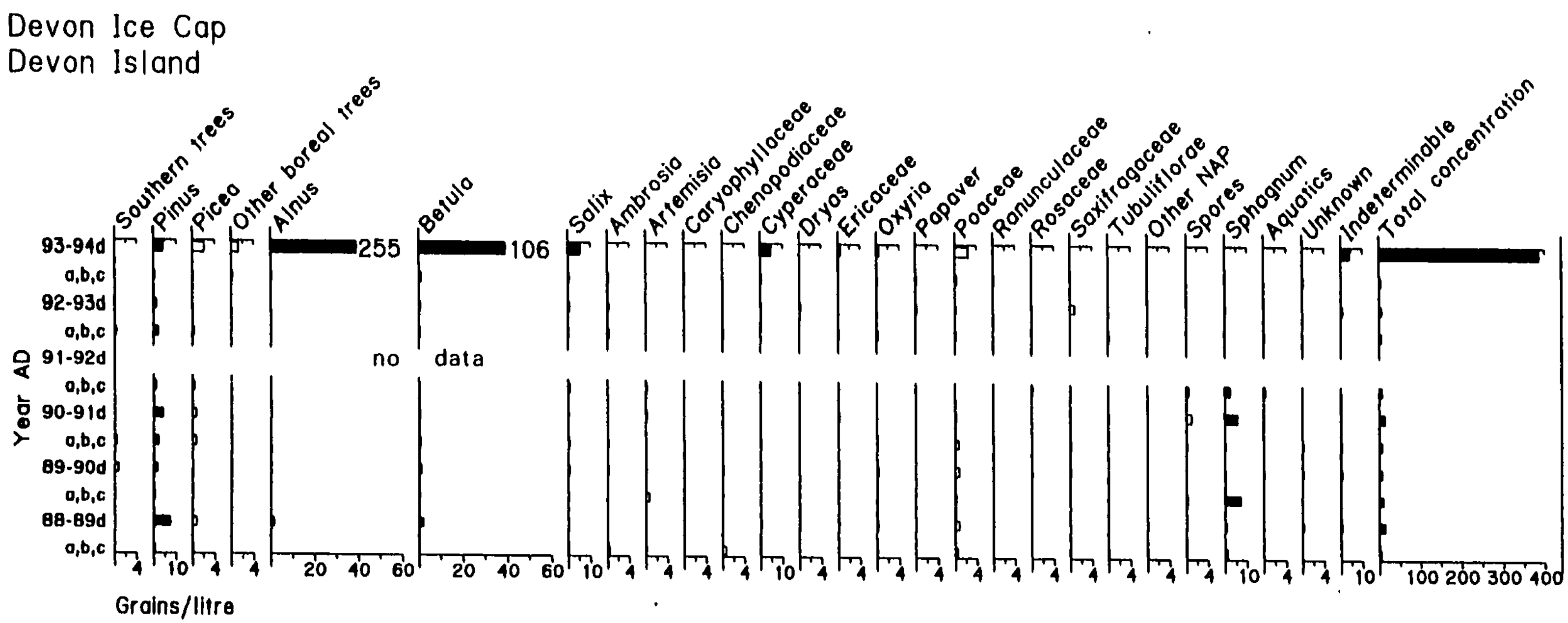

Figure 3.11. Pollen concentration diagram for Devon Ice Cap. Note change in scale.

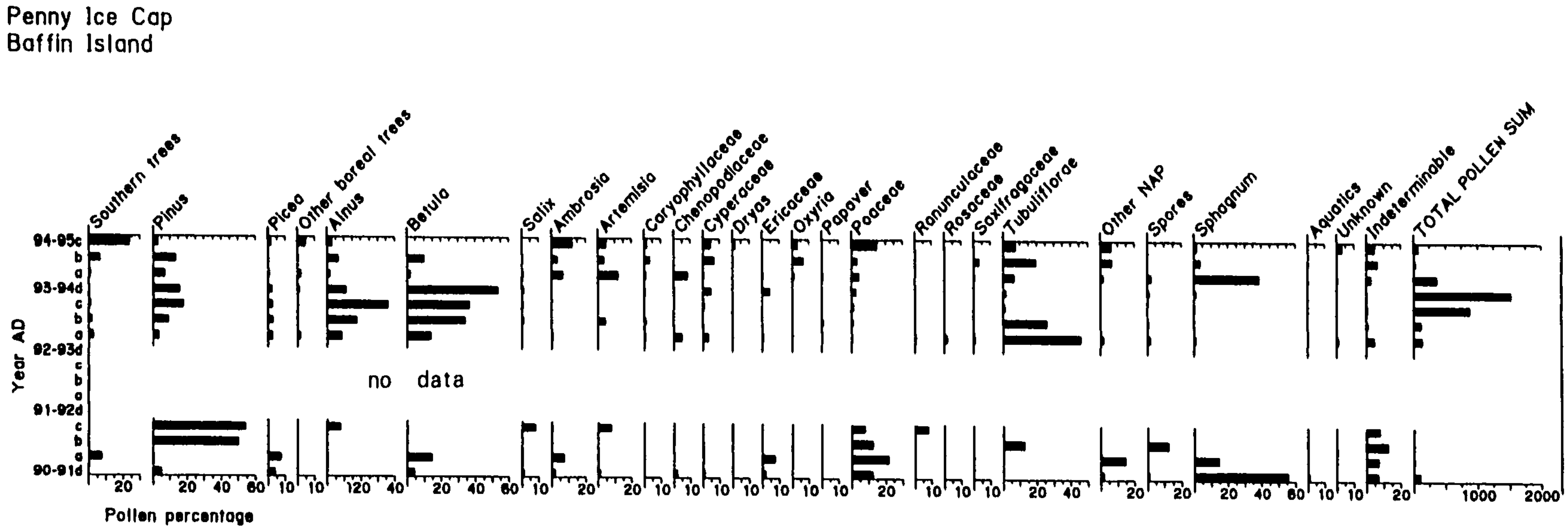

Figure 3.12. Pollen percentage diagram for Penny Ice Cap.

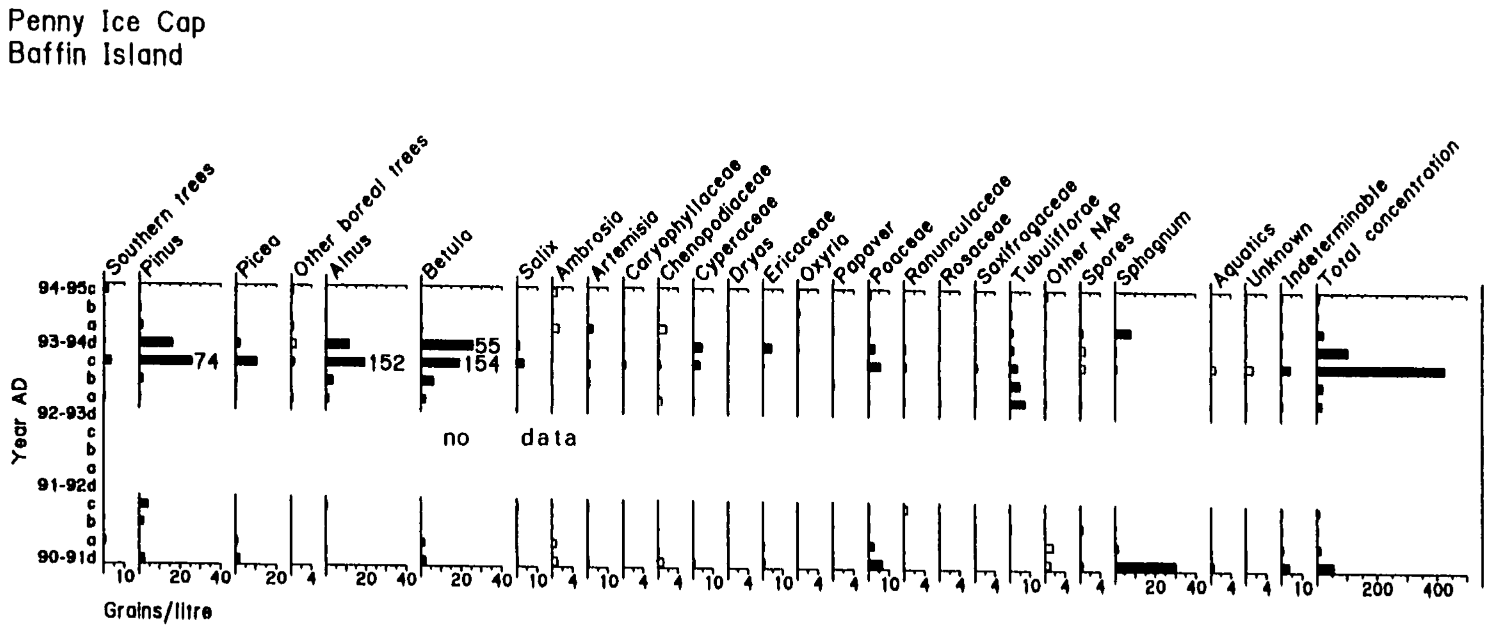

Figure 3.13. Pollen concentration diagram for Penny Ice Cap. Note change in scale.

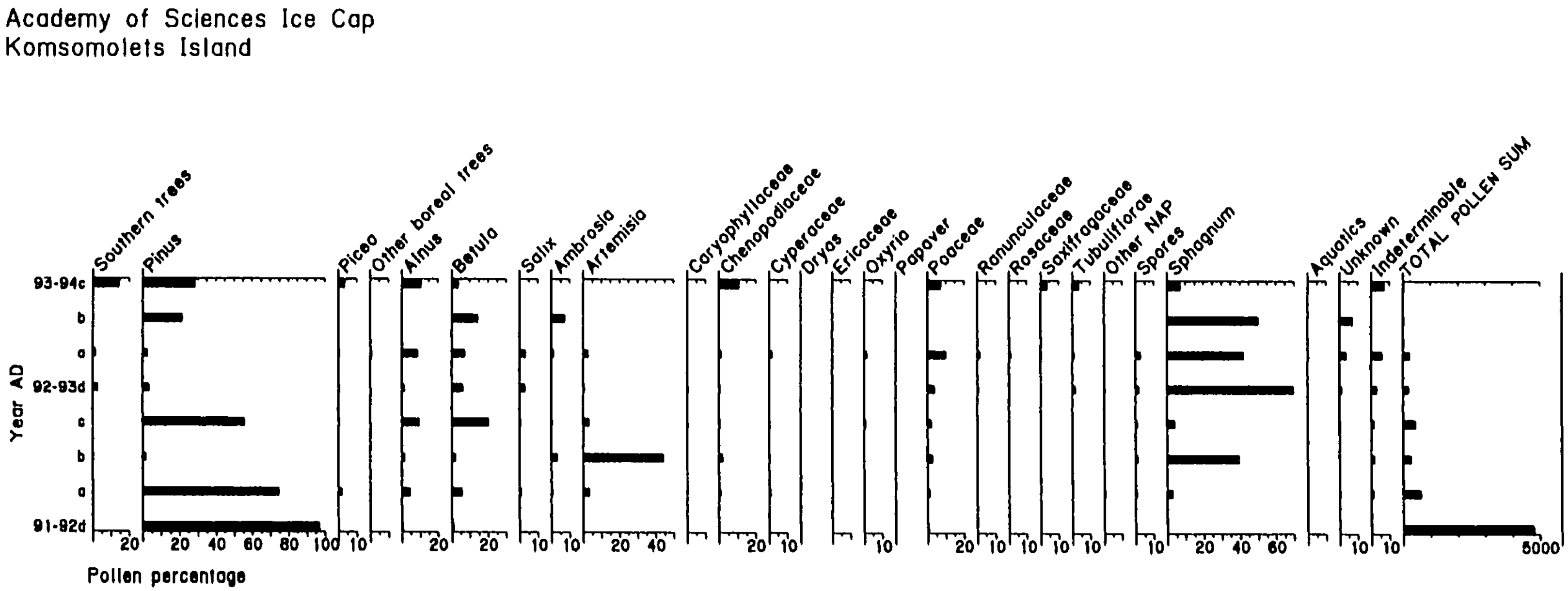

Figure 3.14. Pollen percentage diagram for Academy of Sciences Ice Cap.

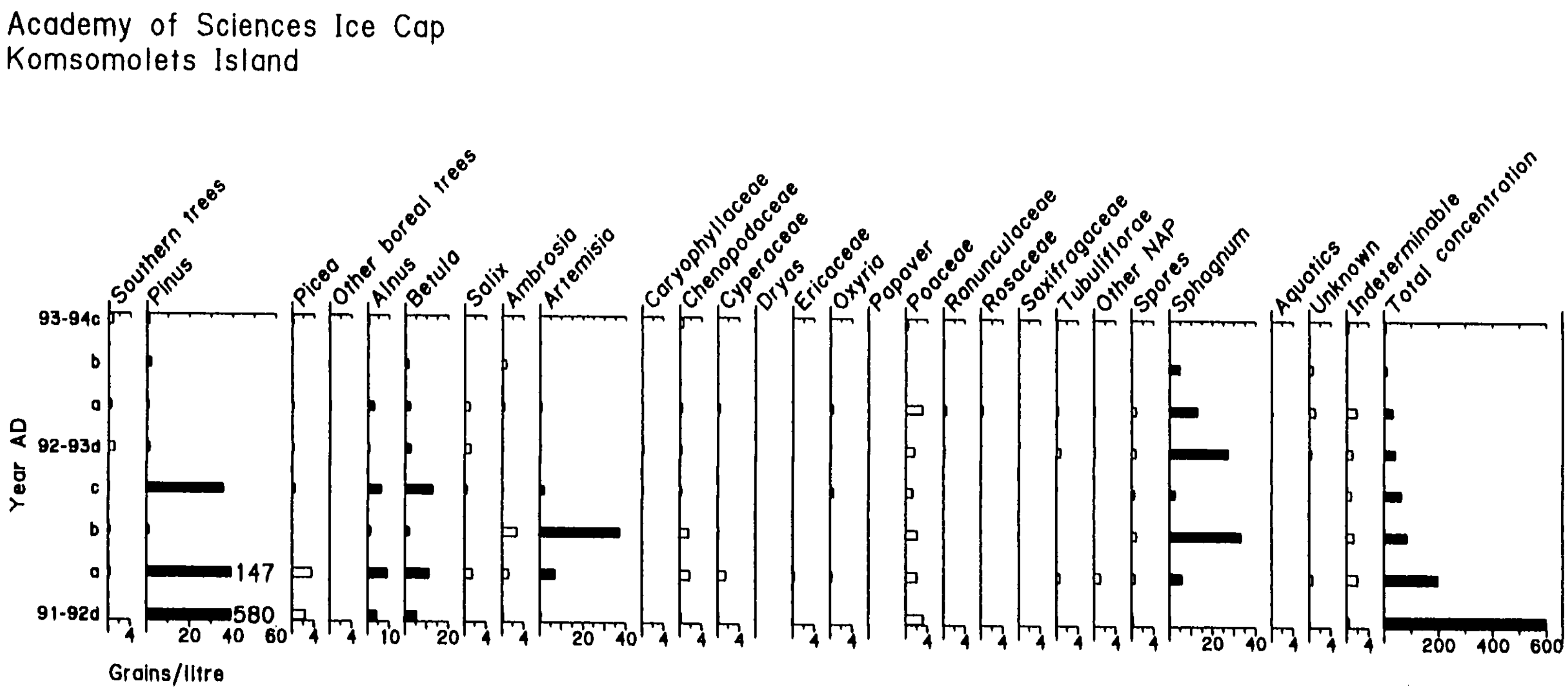

Figure 3.15. Pollen concentration diagram for Academy of Sciences Ice Cap. Note change in scale.

CHAPTER 4

A HOLOCENE ICE CORE POLLEN RECORD FROM ELLESMERE ISLAND, NUNAVUT, CANADA

Abstract

A Holocene record of pollen deposition was obtained from an ice core drilled through the Agassiz Ice Cap. The pollen records long-range atmospheric transport to the ice cap. Pollen concentrations were highest in the early Holocene, decreased in the mid-Holocene and increased after *ca.* 3500 years ago. Before *ca.* 3500 years ago, the pollen concentration curves parallelled the ice core $\delta^{18}O$ and summer melt layer curves, both regarded as temperature proxies. The record suggests that there were regional sources of pollen in the early Holocene, but diversity increased in the mid-Holocene. In the early Holocene, the higher concentration of tree pollen at a time when large parts of Canada were still ice-covered, and when forest zones were more limited and generally farther away, implies that atmospheric circulation was stronger than at present. Following deglaciation, as vegetation migrated north in central and eastern Canada, pollen sources became closer to Agassiz Ice Cap. However, the concentration of tree pollen decreased on the ice cap. This was followed by several relatively rapid changes after 3500 years ago. It is not clear whether these changes are associated with variations in vegetation in the source regions or with changes in atmospheric circulation patterns.

4.1 Introduction

In the Canadian high arctic, Holocene records of paleoenvironmental changes are relatively scarce. Records are often discontinuous or cover only a few thousand years. Ice cores, recovered from large ice caps in the Canadian Arctic Archipelago, are exceptions and provide indicators of paleoenvironmental changes on an annual, decadal, centennial or millennial scale of resolution. The data have contributed to multi-disciplinary efforts in reconstructing Holocene climates of the Canadian high arctic (Bradley, 1990) and circum-arctic of the past 400 years (Overpeck *et al.*, 1997).

The Agassiz Ice Cap (Fig. 4.1), Ellesmere Island, the most studied of the Canadian Arctic ice caps, has been drilled to bedrock six times, and samples from several snow-pits have been collected near the boreholes for environmental and climatological studies. The ice cores have provided records of stable isotope ratios ($\delta^{18}O$) (Fisher *et al.*, 1983, 1995; Fisher and Koerner, 1994), ice-melt stratigraphy (percentage of melt layers in core sections) (Koerner and Fisher, 1990), pH and conductivity (Koerner and Fisher, 1982; Barrie *et al.*, 1985) and solid electrical conductivity (ECM) (Zheng *et al.*, 1998).

The pollen content of the Holocene ice (core A77) was investigated by Bourgeois (1986). However, that pollen diagram was based on the analysis of meltwater produced at the ice/drill-head interface during thermal drilling in 1977. Although the volume of meltwater produced was relatively large, the number of pollen grains recovered was small. Subsequent pollen analysis of snow layers collected near the borehole (Bourgeois, 1990) and of ice samples retrieved from the deepest part of another Agassiz Ice Cap borehole (Koerner *et al.*, 1988), revealed a substantially higher number of pollen grains. A regional survey of modern pollen deposition in arctic snow (Bourgeois *et al.*, 1985) also indicated that more pollen grains are deposited at the surface of the ice cap. The loss of pollen grains in thermal drill meltwater remains unexplained, and experiments to reveal the cause of the loss were inconclusive.

We here present a second Holocene pollen record from an Agassiz Ice Cap ice core. The borehole site (core A87, Table 4.1) is at the top of a local ice divide in the northern part of the ice cap. It is only 30 m away from another borehole drilled in 1984 (A84) and 1 km away from the A77 drill site. Due to the relative abundance of samples from that part of the ice cap, core segments, rather than meltwater, were available for pollen analysis. The objectives of this study are (a) to present a more reliable pollen profile for the Agassiz Ice Cap, and (b) to establish the paleoenvironmental significance of the record. This investigation adds to a short list of contributions on the pollen content of ice cores retrieved in arctic regions (Fredskild and Wagner, 1974; Lichti-Federovich, 1975; McAndrews, 1 984; Short and Holdsworth, 1985; Bourgeois *et al.*, 1985, 1986; Koerner *et al.*, 1988; Andreev *et al.*, 1997).

4.2 Methods

The ice core was recovered with a CRREL thermal drill. Each 1.5 m long ice core segment was split lengthwise, and one half was used for pollen analysis. In the field, each sample was weighed, melted, and filtered through an 8 μm pore size cellulose filter (47mm diameter). Before filtering the sample, a known quantity of *Lycopodium* spores was added to the water. Based on the percentage of *Lycopodium* spores recovered, the percentage of pollen grains lost during filtering and processing of the sample was calculated, and total concentration computed. However, when the percentage of *Lycopodium* recovered was <10% (and usually the number of pollen recovered was extremely low) the data were considered unreliable and were not included in the pollen diagram. This represents approximately 10% of the samples.

In the laboratory each filter was soaked in hydrofluoric acid to dissolve a large portion of the mineral particles. This first step was done in Teflon beakers, and subsequent steps in a filtering apparatus to limit the number of pollen grains lost through centrifuging and decanting. The pollen residue and the partly dissolved filter were filtered through an 8μm polycarbonate filter. After rinsing with water and glacial acetic acid, the remaining cellulose filter and the pollen residue were subjected to acetolysis treatment and rinsed with water. Finally, the transparent polycarbonate filter was removed from the filtering apparatus and placed on a microscope slide with a few drops of glycerine. The complete pollen content of each filter was counted under 250 or 400 magnification. During the laboratory procedures, the air was filtered to limit contamination and blanks, processed along with the samples, showed that contamination was negligible.

4.3 Chronology

The initial timescale for the Agassiz Ice Cap ice cores was based on both a theoretical ice flow model and on measured layer thicknesses. This time scale was verified using the volcanic acid-layer stratigraphy obtained from solid electrical conductivity measurements (Fisher *et al.*, 1983, 1995). These volcanic eruptions are either well-dated from historical records or can be correlated with the high-resolution Greenland ice cores. The Agassiz Ice Cap timescale is considered to be accurate to within 10% for the past 7000 years (Fisher *et al.*, 1983). The cores cannot be dated unequivocally beyond 7000 years ago, but the depth-age curve was adjusted to an abrupt step in the $\delta^{18}O$ profile which corresponds to the late Pleistocene-Holocene transition. It occurs a few metres above bedrock in all Agassiz Ice Cap cores. In the Greenland Ice Core Project (GRIP) core, the transition has been dated at 11,550 years ago (calendar years) by counting the annual layers (Johnsen *et al.*, 1992).

4.4 Results

The results are plotted in 1000 year intervals for the past 7000 years (Fig. 4.2, 4. 3). Between 7000 and 11,500 years ago, the time scale is less accurate and only three intervals were retained. The results presented are averages of the time intervals. These averages are based on a decreasing number of core segments with increasing depth due to the thinning of the annual layers by densification and vertical strain. Consequently, the number of pollen grains counted for each interval decreases with depth thus resulting in some small pollen sums. The concentrations (Fig. 4.3) are computed by adding the corrected number of pollen grains (based on the % of *Lycopodium* spores recovered) in each interval and dividing that number by the total number of litres of water in the same interval. Percentage and concentration diagrams are also computed for the past 1000 years (Fig. 4.4, 4.5), at 50 to 200 year intervals.

The boreal trees *Picea* and *Pinus* (mainly diploxylon type) and the low arctic shrubs

Alnus (mainly *A. crispa* type) and *Betula* are overwhelmingly represented throughout the Holocene. Except for *Betula*, which grows in southeast Baffin Island and southwest Victoria Island, these taxa are restricted to the mainland. However, both *Betula* and *Alnus* grow in West Greenland to latitudes of 75°N and 66°N, respectively. There are also widespread Tertiary deposits in the high arctic that contain tree pollen. When incorporated in lake sediment or peat, these rebedded pollen cannot always be differentiated from contemporary pollen. However, it is unlikely that these pollen grains contribute significantly to pollen assemblages in the higher elevation of the large ice caps (Gajewski *et al.*, 1995). In the present context, *Salix*, Ericaceae, Cyperaceae, *Oxyria*, and Poaceae are considered primarily of regional origin, although the presence of some long-distance pollen is possible.

The Holocene

The percentage diagram (Fig. 4.2) is divided into two pollen zones based on variations of 'Southern trees', *Picea* and *Alnus*. 'Southern trees' comprises pollen of deciduous and coniferous taxa whose modern range lies south of the boreal forest zone. *Pinus* is the most abundant pollen type with percentages ranging between 25-37% for most of the Holocene. Only in the very early Holocene and between 5000-6000 years ago did percentages drop below 15%. In zone I, (11,500 to 6000 years ago), 'Southern trees' and *Picea* peak, whereas *Alnus* is sparse or absent. *Betula* fluctuates between 4-26%. Some of the tundra taxa (Poaceae, Ericaceae and Cyperaceae) have relatively high percentages until 7000 years ago. In zone II (6000 years ago to the present), the percentages of 'Southern trees' decline. *Picea* decreases to <10% from a maximum of 21% in zone I. *Alnus* and *Betula* both increase after 6000 years. *Salix* and *Oxyria*, presently major components of the high arctic tundra, appear consistently in the pollen assemblages after 6000 years ago while other tundra types are generally less well represented.

Pollen concentrations range between 4-17 pollen grains/L (Fig. 4.3). The two pollen zones apparent in the percentage diagram are well defined in the concentration diagram. In zone I, total pollen concentrations are consistently higher than in zone II, except for the past 1000 years. The concentration of 'Southern trees', *Pinus* and *Picea* are higher before 6000 years ago as are Poaceae, Ericaceae and Cyperaceae before 7000 years ago. After 7000 years, there is an increasing number of 'Other herbs" reaching the ice cap. These are all other herb taxa, regardless of their most likely origin. It includes types that are exotic to the high arctic region such as *Ambrosia*, *Artemisia* and Chenopodiaceae and others that are more typical of a tundra environment such as Saxifragaceae, Papaveraceae, Brassicaceae and Caryophyllaceae. *Alnus* appears in the early Holocene, but it becomes more important after 6000 years ago. *Betula*, present throughout the Holocene, has higher concentrations immediately above the late Pleistocene-Holocene transition and just before 7000 years ago.

The past 1000 years

Pinus dominated the pollen assemblages during the past 1000 years. Percentages (Fig. 4.4) varied from 14% (A.D. 1600-1700) to 74% (A.D. 1850-1900). *Picea*, relatively abundant in the first half of the Holocene, did not reach more than 10% during the past 1000 years. The dominant herb types were Poaceae (max. 16%) and *Oxyria* (max. 8%).

The curves of pollen concentrations (Fig. 4.5) differ considerably from the percentages. Total concentrations vary between 4-22 pollen grains/L. *Pinus* concentrations are low before A.D. 1750 and they peak (16 grains/L) between A.D. 1850-1900. All pollen types, except for *Oxyria*, have higher concentrations during the period A.D. 1000-1200 than during the following period of A.D. 1200-1600. At A.D. 1600, most taxa (except *Pinus*) increase and peak between A.D. 1800-1850. Cores were not obtained for the latter part of the 20th century. However, shallow cores recovered approximately 1 km from the drill site in the spring of 1983 (Bourgeois *et al.*, 1985) showed that during the 1970-1983 period, the average, total concentration was 7 grains/L compared to 15 grains/L for the 1900-1950 period (this study) and to 14 grains/L for the 1981-1994 period (Bourgeois, in press).

We did not calculate influx values because of some uncertainties related to the accumulation rate at the beginning of the Holocene. According to Paterson and Waddington (1984), changes to the accumulation rate have probably been <10% on the Agassiz Ice Cap since at least 6500 years ago, but Koerner and Fisher (1990) have suggested that accumulation was possibly higher in the early Holocene. Presently, the influx value (based on the 1981-1994 period) is 0.25 pollen/cm^2/yr. This influx may serve as a low estimate for the early Holocene. By comparison, Holocene pollen influxes in high Arctic lakes in Canada and northern Greenland vary between <1 and 30 pollen/cm^2/yr (e.g. Fredskild, 1985; Hyvärinen, 1985; Funder and Abrahamsen, 1988; Blake *et al.*, 1992).

4.5 Discussion

The number of pollen grains recovered from the A87 ice core is small but is consistent with modern values obtained for the Agassiz Ice Cap, whereas concentrations previously derived from the A77 core (Bourgeois, 1986) were approximately one order of magnitude lower. Another pollen record, obtained from Devon Ice Cap (McAndrews, 1984), was also produced from the meltwater recovered during thermal drilling, and therefore, may have suffered from the same problems affecting the A77 pollen record. Pollen concentration values calculated for the past 5000 years are, however, in the same range (between 2-13 grains/L) as those of the A87 core, whereas concentrations for the 10,000 to 5000 year period average less than 1 grain/L, a value more similar to those of the A77 record. In the Devon core, the low pollen concentrations in the early Holocene have been attributed to the presence of the Laurentide Ice Sheet to the south and consequently, to a greater distance from source areas. If this were so, then the higher pollen concentrations on Agassiz Ice Cap in the early Holocene suggest different source regions and/or different atmospheric trajectories for the two ice caps. Although other ice core parameters have been studied at these two sites and have contributed data on temperature changes during the Holocene, it has not previously been possible to make inferences on past atmospheric circulation.

Comparison with the Agassiz Ice Cap $\delta^{18}O$ and summer melt records

The tree and herb pollen concentrations have been plotted along with the Holocene $\delta^{18}O$ and summer melt profiles of the Agassiz Ice Cap (Fig. 4.6). The records show the least negative $\delta^{18}O$ and the maximum summer melt occurring early in the Holocene but not simultaneously. The lower δ values before 9000 years ago were probably from the melting of the large ice sheets which released massive amounts of freshwater to the surface of the Atlantic Ocean (Fisher, 1992). The summer melt, considered a more direct indicator of past summer conditions on the ice cap, was not as affected by that phenomenon. Maximum melting was probably before 9000 years ago (Koerner and Fisher, 1990) and it remained higher than present until *ca.* 6000 years ago. Tree pollen concentrations were also highest in the the early Holocene and, except for a brief period between *ca.* 6500 to 7000 years ago, concentrations remained high until about 6000 years ago. Herb pollen concentrations peaked during the same period.

After the early Holocene maximum, both $\delta^{18}O$ and melt decreased, indicating a gradual cooling trend. Between 6000 and 3600 years ago, tree pollen concentrations also decreased, thus parallelling the trend of the $\delta^{18}O$ and summer melt records. However, the records diverged after 3600 years; summer melt and $\delta^{18}O$ values remained low but tree pollen concentrations showed more variability and generally increased. The herb pollen profile, relatively constant between 6000 and 3000 years ago, started to vary after 3000 years ago; the largest increase occurred 2000 years ago.

There was a marked decrease in $\delta^{18}O$ values after 1500 years ago and the most negative δ values of the Holocene occurred 100-200 years ago. From 2500-100 years ago, melt was at a minimum. Summer melt and the $\delta^{18}O$ have both increased this century. By comparison, the most recent increase of tree pollen was remarkably high, whereas there was only a moderate increase of herb pollen.

Paleoenvironmental significance of the ice core pollen record

We can only draw broad conclusions about past regional environments based on the small number of herb pollen recovered from the Agassiz Ice Cap ice core. Assuming that most of the herb pollen grains have a regional origin, then the higher pollen concentrations in the early Holocene suggest the presence of a relatively dense vegetation in parts of the high arctic. The most direct evidence of warmer climate than today in the early Holocene comes from the Agassiz Ice Cap ice core records which show less negative $\delta^{18}O$ values and the highest percentage of summer melt for the entire Holocene. A record of summer sea ice conditions in the Canadian Arctic Archipelago, based on remains of bowhead whales (Dyke *et al.*, 1996), also indicates maximum warmth at the time. However, there are few paleobotanical data available for comparison in the early part of the Holocene. Only two long Holocene pollen sequences have been recovered from lacustrine sediments in the Canadian high arctic: Baird Inlet (Hyvärinen, 1985) in east-central Ellesmere Island and southern Somerset Island (Gajewski, 1995). Both have a higher representation of local or regional taxa than the Agassiz Ice Cap ice core. However, the same pollen types (e.g. *Salix*, *Oxyria*, Cyperaceae, Poaceae and Ericaceae) dominate the herb pollen assemblages. The records from these two lacustrine sites also show higher pollen concentrations in the first half of the Holocene. This could suggest a denser vegetation and possibly warmer conditions than today (Gajewski, 1995) although Hyvärinen (1985) suggests Baird Inlet may only be recording a vegetation succession. Other evidences for a warmer climate than today are found in peat deposits from Ellesmere Island (e.g. Garneau, 1992) and from Bathurst Island, in the central high arctic (Blake, 1964). At the latter site, seeds of plants now absent from the region, such as *Vaccinium uliginosum* and *Salix pseudopolaris*, were identified at the base of a 9200 year-old sequence.

On Agassiz Ice Cap, concentrations were higher in the early Holocene, but diversity increased in the mid-Holocene (as seen in the 'Other herbs' category). Other pollen types, representative of high arctic environments, such as *Salix* and *Oxyria*, became more important in the second half of the Holocene. At the lake sites, pollen concentrations began decreasing after *ca.* 6000 years ago and remained consistently low thereafter. In North and West Greenland, pollen records generaly indicate more favourable conditions in the mid-Holocene (eg. Fredskild, 1985; Funder and Abrahamsen, 1988; Blake *et al.*, 1992). However, changes in pollen assemblages, starting at 4000 to 3500 years ago, have been interpreted as a climatic deterioration.

On Agassiz Ice Cap, where there are no local pollen sources, tree pollen has dominated the assemblages throughout the Holocene. Long-distance transported pollen has been found in relatively large amounts in all arctic pollen assemblages, and it has been used to infer changes in past atmospheric circulation (e.g. Nichols *et al.*, 1978; Fredskild, 1984; see also discussion in Barry *et al.*, 1981). In Greenland, the occurrence of tree pollen has been related to the migration of tree species to northern Quebec and Labrador following the late disintegration of the Laurentide Ice Sheet in the mid-Holocene (Fredskild, 1984, 1985). The record from Baird Inlet, east-central Ellesmere (Hyvärinen, 1985), also shows few boreal or low arctic pollen types before *ca.* 7000 years ago. Conversely, Gajewski (1995) found high percentages (max. 50%) and high concentrations of tree pollen on Somerset Island, in the central arctic. The highest concentrations were obtained in the first half of the

the Holocene, but unlike the Agassiz profile, it was dominated by the low arctic shrubs *Alnus* and *Betula* with lesser amounts of *Picea* and *Pinus*.

In the first half of the Holocene, the Agassiz tree pollen profile shows some similarities with pollen sequences from the northwest mainland of Canada, which show an early Holocene thermal maximum (e.g., Ritchie *et al.*, 1983, Ritchie, 1984). *Picea* migrated to the region before 10,000 years ago and displaced the *Betula*-dominated shrub tundra. Treeline may have been 50 to 75 km north of its present limit between 7000 and 10,000 years ago, and *Picea* pollen influx values were up to ten times higher than they are at present (Ritchie *et al.*, 1983; Spear, 1993). After 7000 years ago, *Alnus* became an important component of the pollen assemblages indicating migration to the region . At about that time, the treeline retreated southward and forested zones reverted to tundra.

In the early Holocene, percentages and concentrations of *Picea* pollen differed from modern values on the Agassiz Ice Cap which rarely reach more than 5% and 0.5 grains/L (Bourgeois, in press). It is unlikely that a northward extension of treeline by less than 100 km would have had a direct effect on pollen deposition on the Agassiz Ice Cap, 2000 km to the northeast. The higher *Picea* values on the ice cap in the early Holocene may have resulted from an increase in pollen productivity. But more likely, if the source region was indeed the northwest mainland, they resulted from an increase in wind frequency and/or intensity from the southwest during spring and early summer. That is, the anticyclonic circulation around the Laurentide Ice Sheet (COHMAP Members, 1988) may have transported *Picea* pollen from the Mackenzie Basin far to the north. Another alternative is that *Picea* pollen originated south of the Laurentide Ice Sheet. However, *Picea* forests had a limited extent south of the ice sheet and the distances involved were much greater. Furthermore, when considering the rapid decline of *Picea* pollen north of the treeline reported in modern spatial studies (e.g. Elliot-Fisk *et al.*, 1982; Bourgeois *et al.*, 1985), it seems unlikely to have been the source region. In West Greenland, where the most probable source region was south of the Laurentide Ice Sheet, *Picea* does not appear significantly in pollen diagrams before 7000 years ago (Fredskild, 1984).

The presence of *Pinus* pollen and, to a lesser extent, of southern tree types in the early Holocene does indicate atmospheric trajectories originating elsewhere than the northwest mainland of Canada. At that time, the most likely sources were south of the Laurentide Ice Sheet, in eastern North America or, possibly, in Eurasia. *Pinus,* in particular, is a prolific pollen producer and the pollen can be dispersed extremely long distances. Presently, *Pinus* pollen concentrations in snow that are most comparable to the early Holocene values on the Agassiz Ice Cap are found at approximately 1000 km to the south and on the Arctic Ocean sea ice in the vicinity of the North Pole (J.C. Bourgeois, unpubl. data). Considering that the number of *Pinus* pollen being transported to the ice cap in the early Holocene was higher than today while the distances involved were more considerable, we assume a stronger atmospheric circulation. However, these pollen grains may have originated in North America or in Eurasia.

With the retreat of the Laurentide Ice Sheet during in the mid-Holocene, vegetation migrated northward in central and eastern Canada. Consequently, sources of pollen for Agassiz Ice Cap were closer. Nonetheless, Agassiz Ice Cap pollen concentrations decreased. The glacial anticyclone had dissipated and the westerly flow circulation pattern (COHMAP

Members, 1988) carried pollen from south of Agassiz Ice Cap. Decreasing temperatures are also inferred from both $\delta^{18}O$ and melt records from the same ice core, but these records do not provide evidence for changing atmospheric conditions. Most proxy data from the high arctic, including others from the Agassiz Ice Cap ice cores, suggest a deterioration of climate in the late Holocene (Bradley, 1985; Douglas *et al.*, 1994), a trend that was only reversed in the last 100 years. The ice core pollen concentration record reveals increased variability of arboreal and non-arboreal pollen types in the late Holocene. This trend is difficult to reconcile with other available data from the region, possibly due to the low resolution of some of the other records.

It is difficult to establish if some of the changes seen in the ice core pollen record reflect changes in the regional vegetation that were not detected in other lower resolution records or if they reflect changes in atmospheric conditions affecting the transport of pollen to the ice cap. More pollen records are needed from the high arctic, and other ice core parameters pertaining to atmospheric transport should be studied in the Agassiz Ice Cap cores as well as cores from other arctic ice caps. Until this is done, it may be difficult establishing to what extent some of the changes reported here are related to changes in vegetation cover and pollen productivity in the source region or produced by changes in atmospheric circulation. On the other hand, this study also suggests that ice core pollen records are sensitive indicators of past atmospheric circulation and may provide new information on Holocene arctic environments.

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Table 4.1. Properties of the A87 ice core.

Latitude	Longitude	Elevation (m.a.s.l.)	Thickness (m)	Ice accumulation (m/a)	Summer melt average (%)
80°40' N	73°30' W	1730	127	0.098	5.1

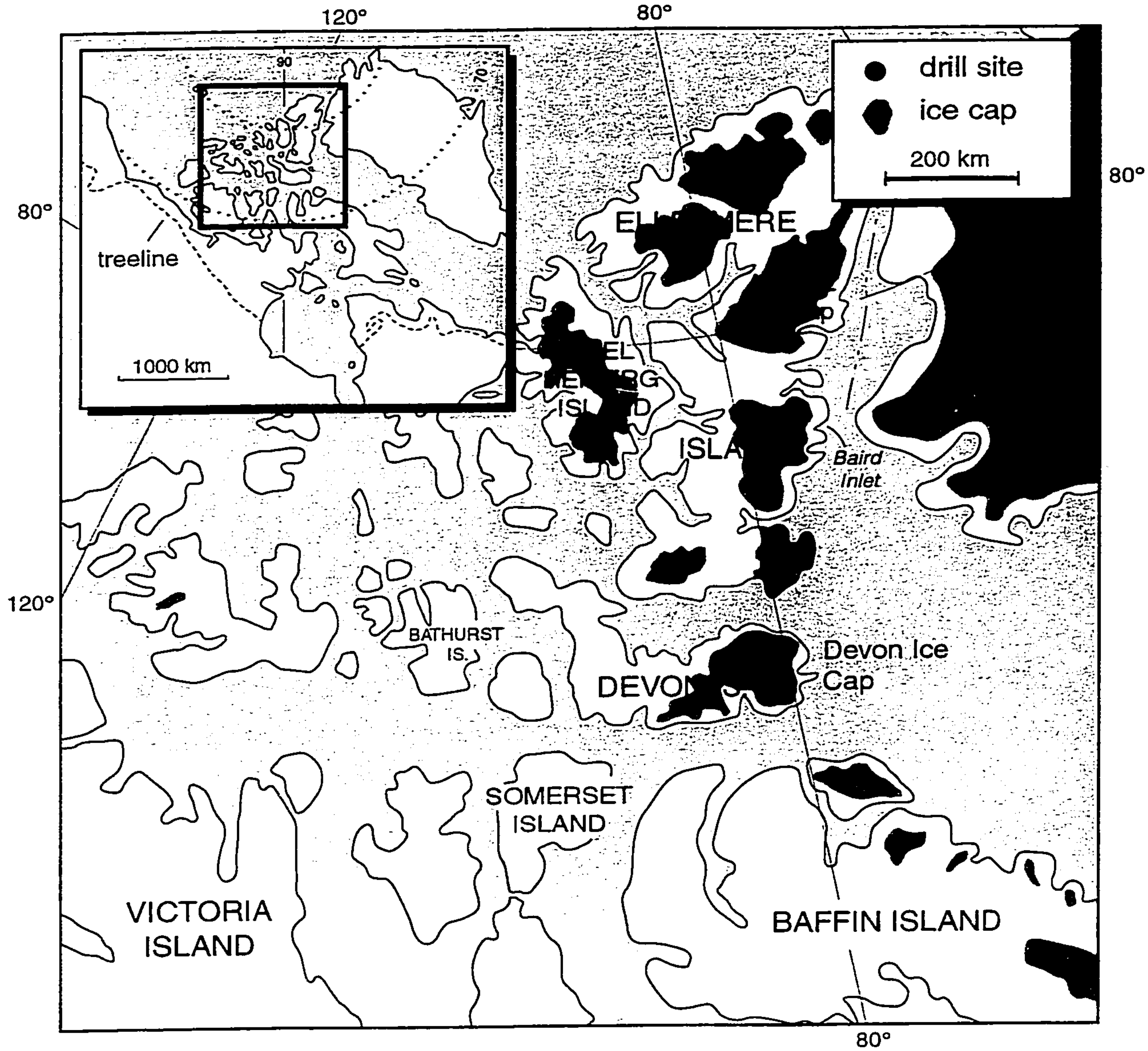

Figure 4.1. Location of the Agassiz Ice Cap drill sites and other sites mentioned in the text.

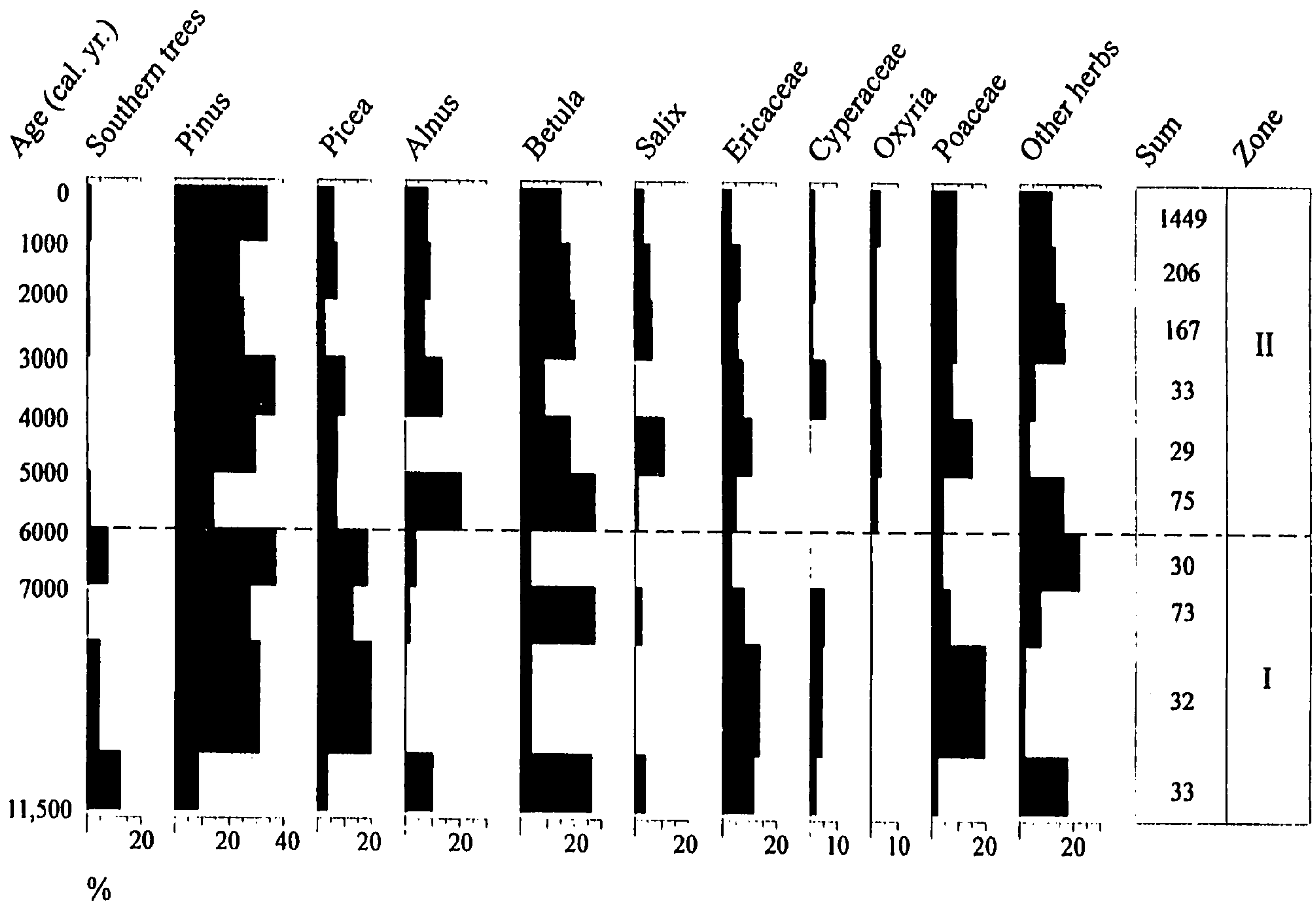

Figure 4.2. Agassiz Ice Cap (A87 core). Holocene pollen percentage diagram plotted in 1000 year intervals until 7000 years ago. From 7000 to 11,500 years ago, the core is not unequivocally dated. The total pollen sum includes all pollen grains counted in the samples, of which approximately 10% are not presented in the diagram. These include the unknown and indeterminable pollen grains.

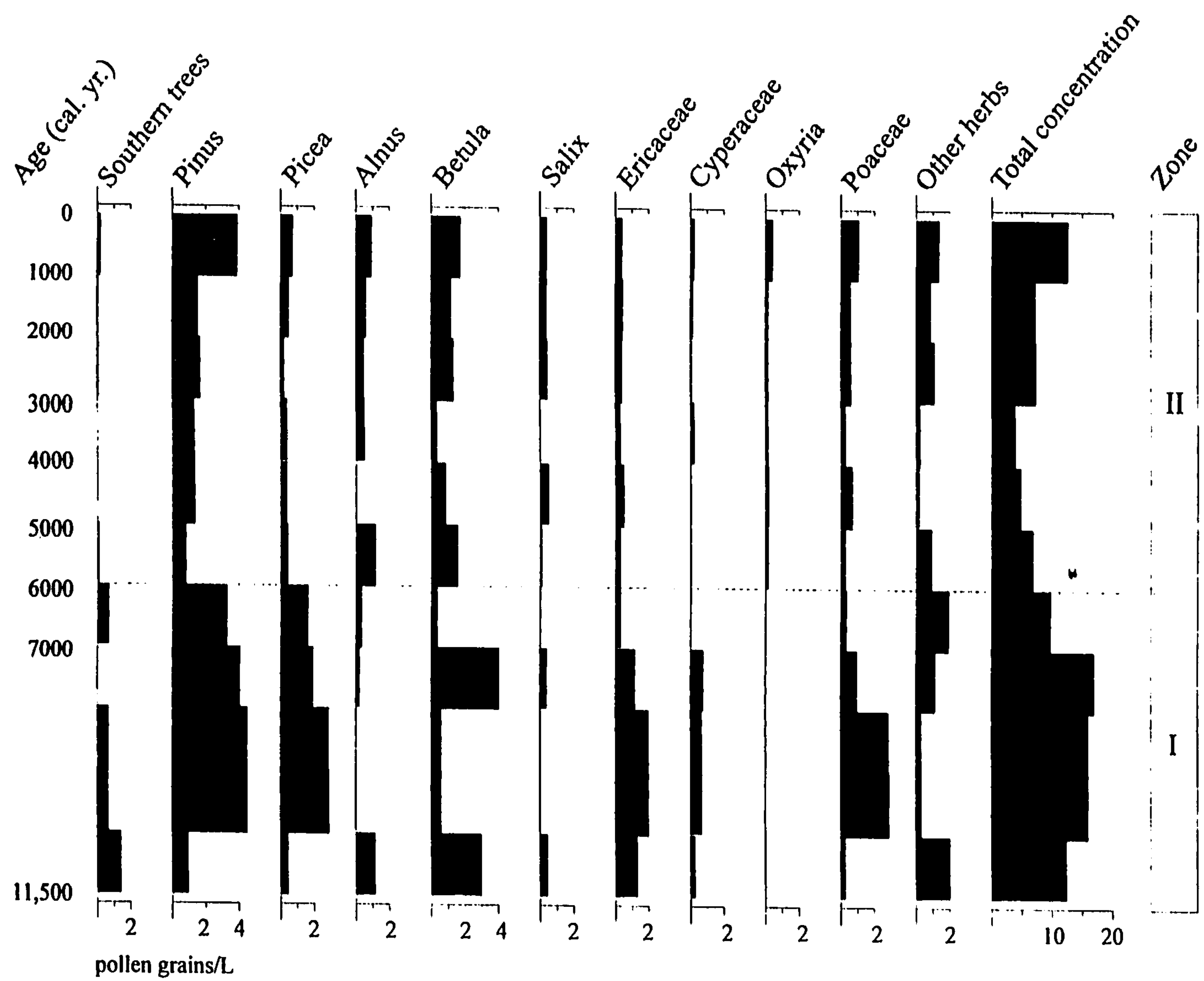

Figure 4.3. Agassiz Ice Cap (A87 core). Holocene pollen concentration diagram plotted on same scale as in Fig. 4.2.

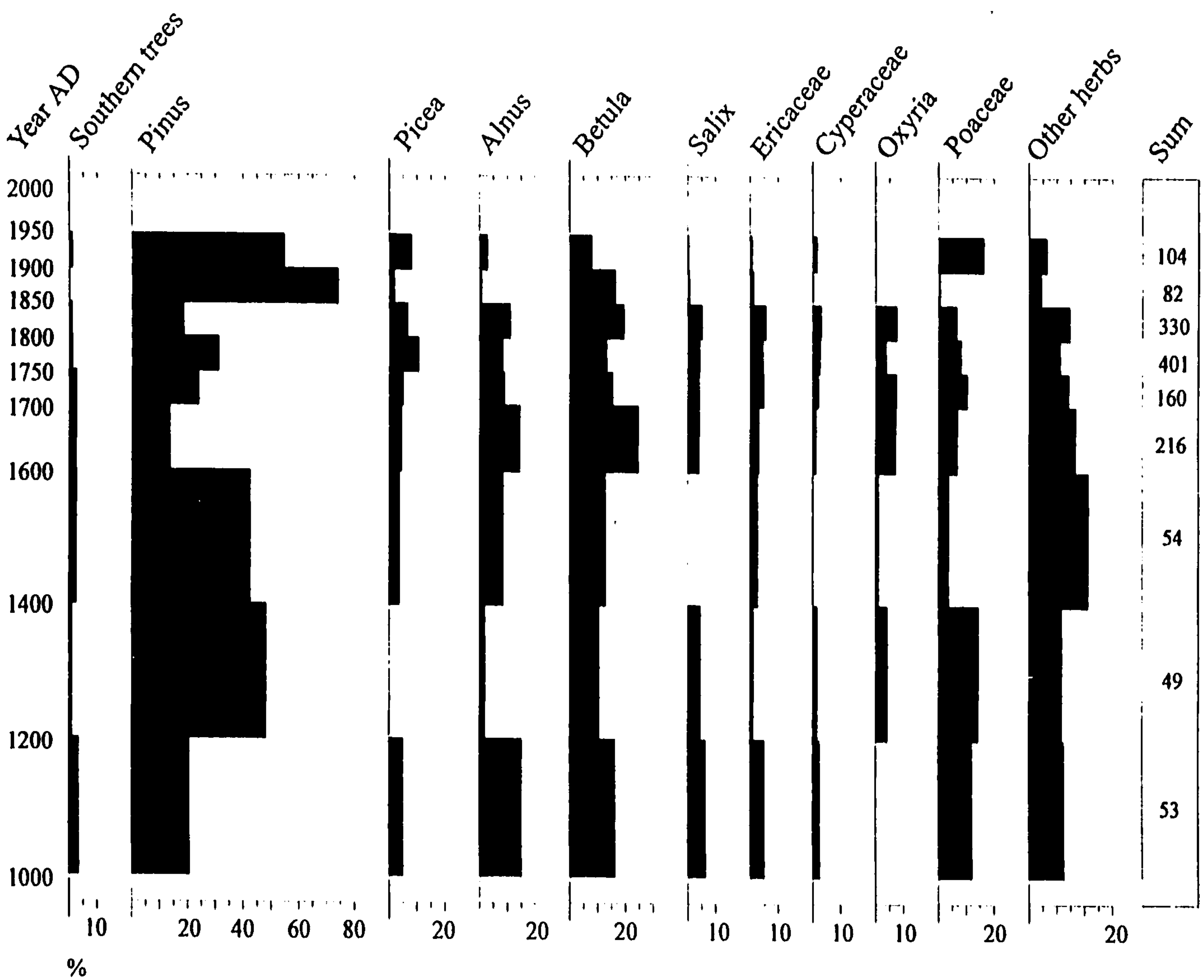

Figure 4.4. Agassiz Ice Cap (A87 core). Pollen percentage diagram plotted for the past 1000 years. The intervals vary between 50 years to 200 years. The total pollen sum includes all pollen grains counted in the samples

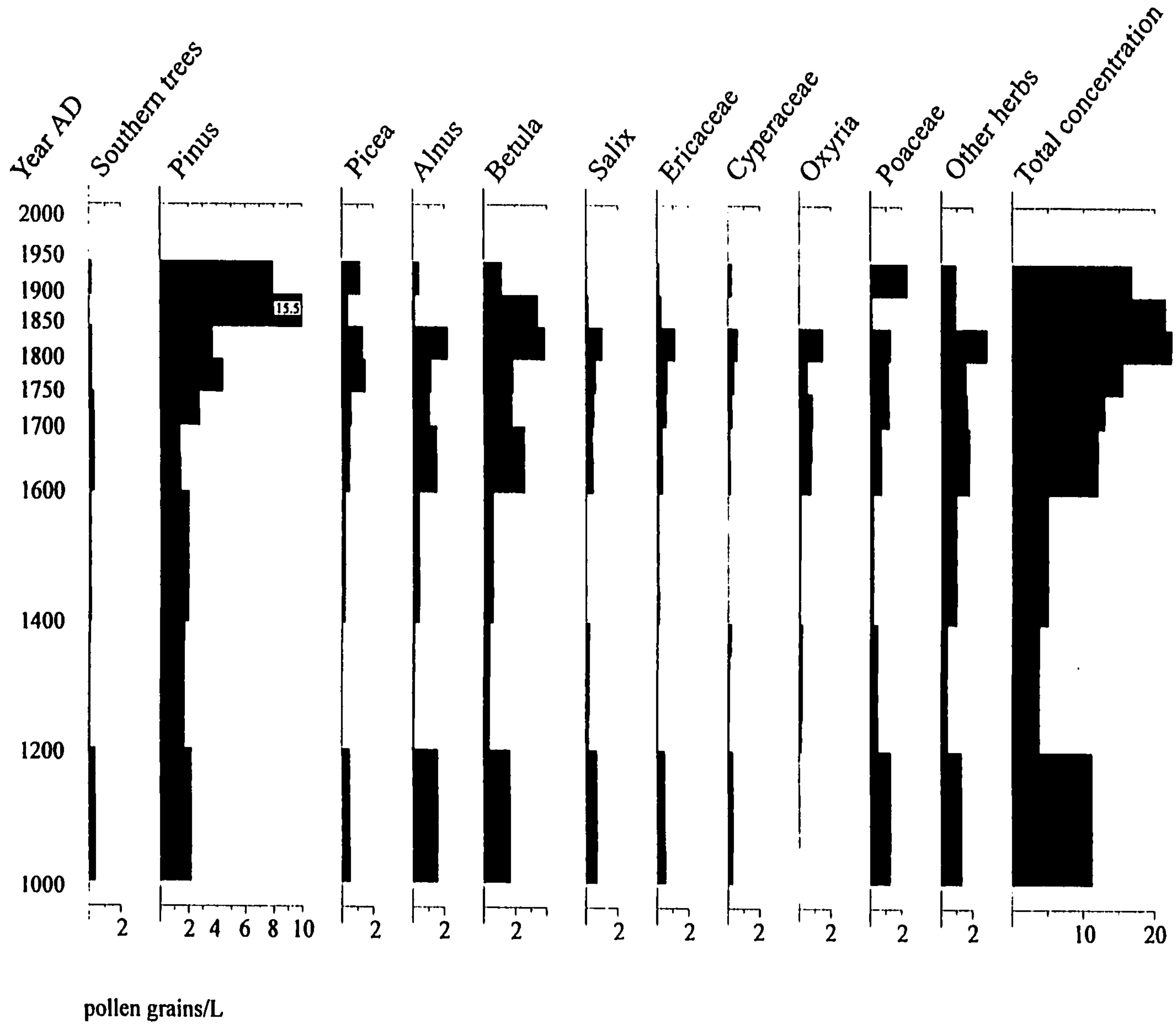

Figure 4.5. Agassiz Ice Cap (A87 core). Pollen concentration diagram plotted for the past 1000 years.

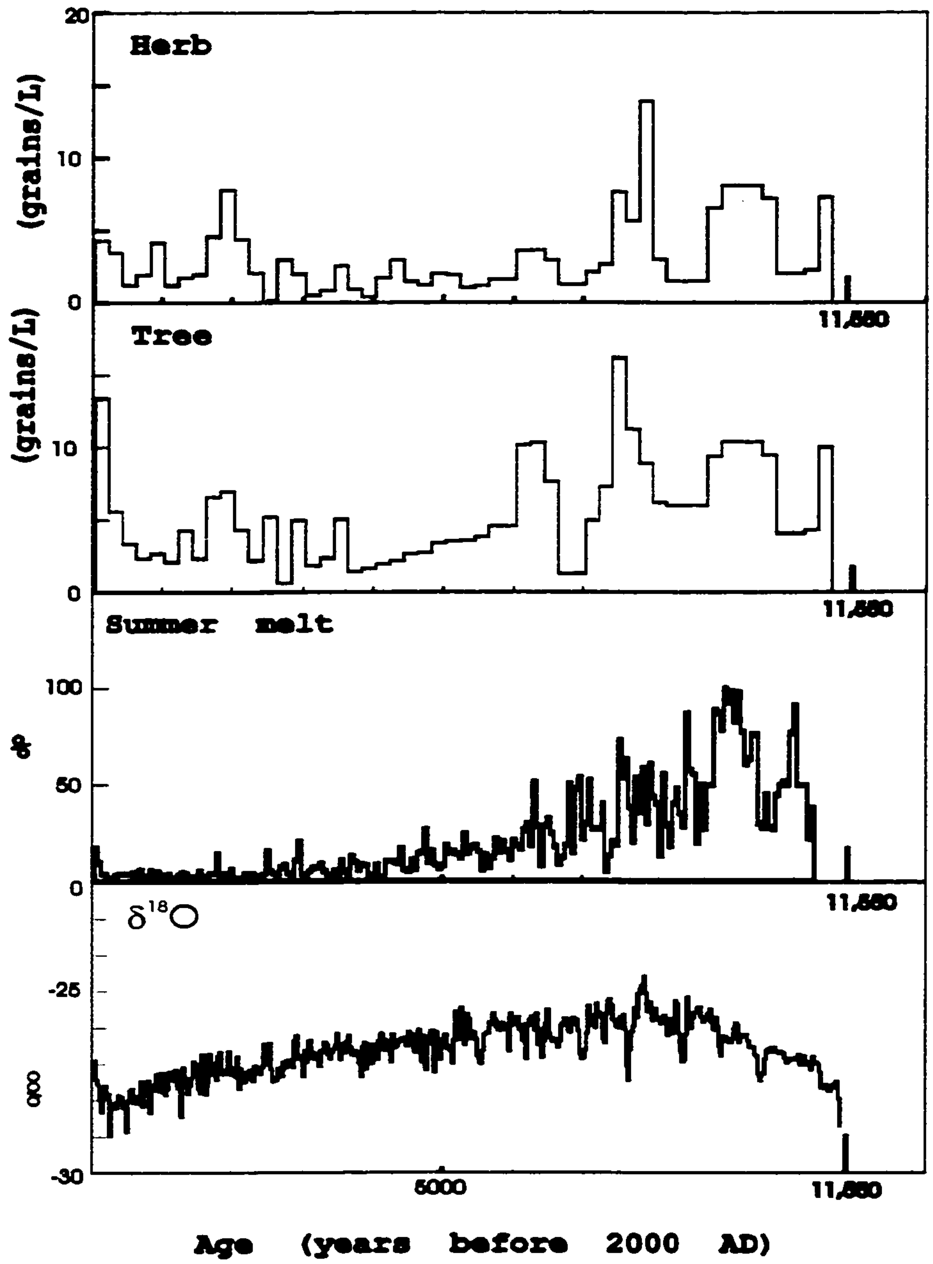

Figure 4.6. Comparisons between the herb and tree pollen concentrations, summer melt, and $\delta^{18}O$ records (Koerner and Fisher, 1990) from the Holocene section of the Agassiz Ice Cap ice core.

CHAPTER 5

SUMMARY AND CONCLUSIONS

5.1 Introduction

This dissertation contributes to our understanding of pollen deposition in arctic regions. It demonstrates how pollen analysis of ice cores can provide valuable information for paleoenvironmental reconstruction. This information is particularly important in the Canadian high arctic where our knowledge of past climates and of environmental change during the Holocene is poor. Pollen profiles retrieved from ice cores provide continuous records for the Holocene and may be extended, at a much lower resolution, into the last glacial and interglacial periods. However, these ice caps are far from vegetation, and the information retrieved from snow and ice cores differs slightly from that of other deposits. Pollen assemblages from arctic ice caps contain pollen derived from the high arctic tundra as well as long-distance transported pollen, primarily tree and shrub pollen from the low arctic and boreal forest. As there is no local source and the productivity of the regional tundra is low, the long-distance component is emphasized in snow and ice core pollen diagrams. In pollen diagrams from the Arctic, the presence of tree pollen is often seen as a form of contamination. In this study it is shown that these long-distance transported pollen can be used to define air mass trajectories in the Arctic. These studies can contribute significantly to the increasingly important field of snow and ice core research, particularly where it concerns establishing past atmospheric circulation patterns. In addition, pollen analysis of snow may help determine pathways of contaminants in the Arctic.

Previous pollen analyses of arctic snow were useful in detecting spatial and temporal variations in pollen transport. The results of these studies, however, were not always comparable. Furthermore, the first Holocene pollen records retrieved from the Devon and Agassiz ice caps were probably affected by the method of extraction, thereby making the reliability of these records questionable. In this study, surface snow and ice core samples were processed using the same field and laboratory procedures. Consequently, this dissertation presents the first comprehensive study of pollen analysis of snow and ice core samples from the Arctic.

5.2 Summary of Major Findings

The data obtained from pollen analysis of snow samples collected at 41 sites in the Canadian Arctic, the Arctic Ocean sea ice and Greenland were used to establish spatial patterns of pollen dispersal in the Arctic (Chapter 2). Pollen assemblages of the major vegetation zones were distinguished despite considerable spatial and temporal variability. Pollen percentages and concentrations are related to the density of the regional vegetation and to the distance to the source of more productive regions. Increased regional productivity generally corresponds to an increase in the concentration of tundra pollen types. In addition, the long-range atmospheric transport of tree and shrub pollen from the low arctic and boreal forest is important at these sites and regional patterns are detected particularly from the concentrations. The presence of air masses originating in Eurasia can be identified by the increase in *Pinus* pollen in the northernmost samples from the Arctic Ocean, whereas incursions of temperate air to the North American Arctic are identified by the presence of taxa typical of boreal regions. An abrupt decrease in the concentration of long-distance pollen occurs at approximately 75°N, which is at the latitude of the Devon Ice Cap. This 'boundary' was found in the first survey on pollen deposition in arctic snow. If it can be shown to be related to atmospheric circulation, this could be relevant for the interpretation of ice core records retrieved from the Devon Ice Cap.

A high resolution study, concerned with the seasonal and annual variations in pollen deposition, was undertaken by sampling the snow layers deposited at the summit of three ice caps and a small glacier in the Canadian Arctic, and at one ice cap in the Russian Arctic (Chapter 3). Pollen assemblages of summer and spring layers differ from those of the fall and winter snow. Pollen succession in the annual snow layer is related to the known phenology of arctic and southern plants, in spite of the great distances involved. The number of tree/shrub pollen reaching the ice caps varies considerably from year to year and this variability increases with decreasing distance to the treeline. Large deposition events of either tundra or long-distance pollen types are registered some years in the snow pack. In the Canadian Arctic, these events are sometimes observed on more than one ice cap. The Agassiz Ice Cap has the longest seasonal record of pollen deposition. At this site, some trends, such as recurring peaks of *Pinus* pollen in the 1990's, are seen and these might have a climatic significance requiring further study.

A Holocene pollen record retrieved from the Agassiz Ice Cap (Chapter 4) reveals considerable changes in pollen percentages and concentrations during the past 11,000 years. As in the modern snow samples, the pollen assemblages are composed of tundra taxa and of a relatively large number of long-distance transported tree pollen. Pollen concentrations, particularly those of *Pinus* and *Picea*, were highest in the early Holocene. Concentrations decreased in the mid-Holocene and there were several relatively rapid changes registered after *ca*. 3500 years ago. Before 3500 years ago, the pollen curves parallel the $\delta^{18}O$ and ice-melt records from the same ice core. The $\delta^{18}O$ and the ice-melt records, both regarded as temperature proxies, indicate warmest summer temperatures and high melt occurring very early in the Holocene. The Agassiz Ice Cap pollen record also suggests that there were regional sources of pollen in the early Holocene contributing, at times, more pollen than today. However, diversity of tundra types increased in the mid-Holocene. The high concentration of tree pollen in the early Holocene, when large parts of Canada were still ice-covered and forest zones more limited and generally farther away, implies that atmospheric circulation was stronger than at present. At the time, the most likely sources of *Picea* pollen were in the Mackenzie Delta area, whereas *Pinus* and southern tree types would have originated south of the Laurentide Ice Sheet or, possibly, in Eurasia. However, the circulation patterns or wind intensity must have changed in the mid-Holocene as the number of tree pollen on the Agassiz Ice Cap decreased, despite the fact that more and closer sources of pollen became available. The fluctuations of pollen concentrations after 3500 years ago and, particularly, the large increase of *Pinus* pollen after A.D. 1850 are difficult to interpret. Until more pollen data are gathered and other parameters are studied from Agassiz Ice Cap ice cores, it may be difficult to determine to what extent some of these changes are associated with variations in vegetation in the source regions or with changes in atmospheric circulation patterns.

5.3 Recommendations for Future Work

This study has shown that pollen assemblages specific to regions could be detected in snow samples and that the long-distance components of these assemblages could provide information on atmospheric circulation in the Arctic. This, in turn, will help to detect pathways of pollutants in the Arctic, a topic that was recently explored in a preliminary study on fluxes of chemical species and pollen in snow of the circumpolar Arctic (Appendix B). These promising results, obtained from that circumpolar study and the spatial survey reported in Chapter 2, indicate that the sampling program should be continued. Presently, snow samples are collected each spring, for pollen and snow chemistry studies, on four Canadian Arctic ice caps but more samples are needed from other locations.

The seasonal records of pollen deposition, presented in Chapter 3, ended in 1995, but on Agassiz and Devon ice caps, samples are being collected every spring. A next step to this study will be to compare atmospheric circulation patterns during the time of deposition as was done in the study presented in Appendix C. There is such a project in place to study the atmospheric circulation in the Arctic for the unusually warm summer of 1998 and to compare the data with various components studied on ice caps, including pollen. This line of study will be particularly useful for the interpretation of the ice core records.

Several ice cores have recently been drilled through the Devon Ice Cap (Devon Island) and the Penny Ice Cap (Baffin Island). During the drilling of each core, ice chips were produced by the electro-mechanical drill as it penetrated into the ice. These ice chips have been recovered for pollen analysis. Some have been analysed from the 350 m long ice core drilled on Penny Ice Cap in 1995. Presently , one of the limitations for obtaining more pollen records from ice cores is the time it takes to count the pollen slides. But more pollen records are needed in order to detect trends or differences between sites and to establish with more confidence the significance of these ice core pollen records. This study has shown the potential of ice core pollen records in arctic paleoenvironmental analyses and in tracing the circulation of air masses in the Arctic. The few pollen data available show results that are largely in accord with other paleoenvironmental records. In some cases the information can extend previous records, such as seasonal input of pollen interpreted from high-resolution analyses of ice layers near the surface. Spatial patterns of pollen can be used to trace air mass movements that are also transporting pollutants onto the arctic environments. Future work will expand our knowledge of the changes in arctic environment at several time and space scales.

APPENDICES

APPENDIX A

Gajewski, K., Garneau, M., and Bourgeois, J.C., 1995: Paleoenvironments of the Canadian high Arctic derived from pollen and plant macrofossils: problems and potentials. *Quaternary Science Reviews*, 14: 609-629.

APPENDIX B

Bourgeois, J.C., Koerner, R.M., and Savatuygin, L., 1997: Snow chemistry and pollen deposition in the circum-polar high arctic. In: *International Symposium on Atmospheric Chemistry and Future Global Environment*. International Global Atmospheric Chemistry (IGAC Annual Meeting), Nagoya, Japan.

APPENDIX C

Alt, B.T. and Bourgeois, J.C., 1995: Establishing the chronology of snow and pollen deposition events on Agassiz Ice Cap (Ellesmere Island, Northwest Territories) from autostation records. *Geological Survey of Canada, Current Research 1995-B*: 71-79.

APPENDIX D

Bourgeois, J.C., Koerner, R.M., Alt, B.T., and Fisher, D.A., in press: Present and past environments inferred from Agassiz Ice Cap ice core records. In: M Garneau (ed.) Environmental Response to Climate Change in the Canadian High Arctic. *Geological Survey of Canada, Bulletin 529.*

APPENDIX A

Pergamon

Quaternary Science Reviews, Vol. 14, pp. 609–629, 1995.

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PALEOENVIRONMENTS OF THE CANADIAN HIGH ARCTIC DERIVED FROM POLLEN AND PLANT MACROFOSSILS: PROBLEMS AND POTENTIALS

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Abstract — Analyses of peat sections, lake sediments and ice cores provide information about Late-Quaternary arctic environments. Palynomorphs and plant macrofossils from each of these three types of sediments record different aspects of the environment with particular spatial and temporal scales of resolution. In the Arctic, the limits to the interpretation of past environments are particularly significant. Problems of low pollen concentrations, long-distance transport, stratigraphic inversions and contamination by fossils from older deposits are more serious in this region due to the biogeographic context that characterizes these high latitudes and influences the particular ecological and geomorphological processes. However, recent work has shown that past environments can be reconstructed from evidence preserved in high arctic sediments and ice cores, if these problems are taken into account before their interpretation.

INTRODUCTION

Holocene pollen diagrams are now available from hundreds of sites across northern North America (Anderson, 1985; Holloway and Bryant, 1985; Ritchie, 1985). Plant macrofossil analyses are less abundant and mostly restricted to boreal and subarctic areas (Couillard and Payette, 1985; Payette, 1988; Nicholson and Vitt, 1990; Nicholson, 1993; Kuhry *et al.*, 1993). Pollen data document the postglacial migration of the vegetation in response to the retreat of the Laurentide Ice Sheet, while macrofossil data offer opportunities to study past plant communities and their associated ecological conditions (Birks and Birks, 1980). These data can be used to indicate the sensitivity of different vegetation formations to past climatic changes at long time-scales. However, one region with little paleoecological data is the Canadian Arctic Archipelago and northern Greenland. With some exceptions (e.g. Fredskild, 1969, 1973; Brassard and Blake, 1978; Bennike, 1983; Funder and Abrahamsen, 1988; Hyvärinen, 1985; Blake *et al.*, 1992), most arctic paleoecological studies are from subarctic and low-arctic environments (e.g. Kelly and Funder, 1974; Funder, 1978; Fredskild, 1973; Ovenden, 1982, 1985; Gajewski and Garralla, 1992; Richard, 1981; Ritchie, 1984, 1987; Short *et al.*, 1985, 1994). Figure 1 shows the biogeographic zonation of the arctic territory into high-, mid- and low-arctic; subarctic describes areas south of treeline (Polunin, 1951).

Twenty years ago, Nichols (1974) summarized the state of arctic palynology in North America. He discussed some of the problems involved in these studies: the problems and expense of logistics, the difficulties of sampling sedimentary deposits in permafrost and the question of pollen data interpretation. Paleoecological studies in the arctic attracted little interest as the methods used at the time could provide only crude results with little taxonomic resolution (Hegg, 1963; Colinvaux, 1967; Jankovska and Bliss, 1977). An exception was the detailed work of Fredskild in Greenland (e.g. Fredskild, 1973). Since the mid-1970s the situation in palynology has changed, as a result of improved coring techniques (Blake, 1964, 1981; Wright, 1980), and laboratory methods (Cwynar *et al.*, 1979), expanded knowledge of arctic pollen taxonomy (e.g. Cwynar, 1982; Ritchie, 1982), and an increased global understanding of the arctic environment. Field work, while still expensive, is less of a problem twenty years later but many questions remain about pollen production, transport, deposition and sedimentation.

Recent work shows the potential of paleoenvironmental studies in the arctic (Bourgeois, 1986, 1990; Fredskild, 1973; Gajewski, *submitted*; Garneau, 1992; Hyvärinen, 1985; Short *et al.*, 1985, 1994), but they have also highlighted problems, especially associated with the high-arctic environment. As discussed in part by Nichols (1974) and Birks (1973), these problems are caused mainly by the presence of permafrost, the composition of the underlying geological deposits, the cold and dry climate and the sparse vegetation cover, as well as limita-

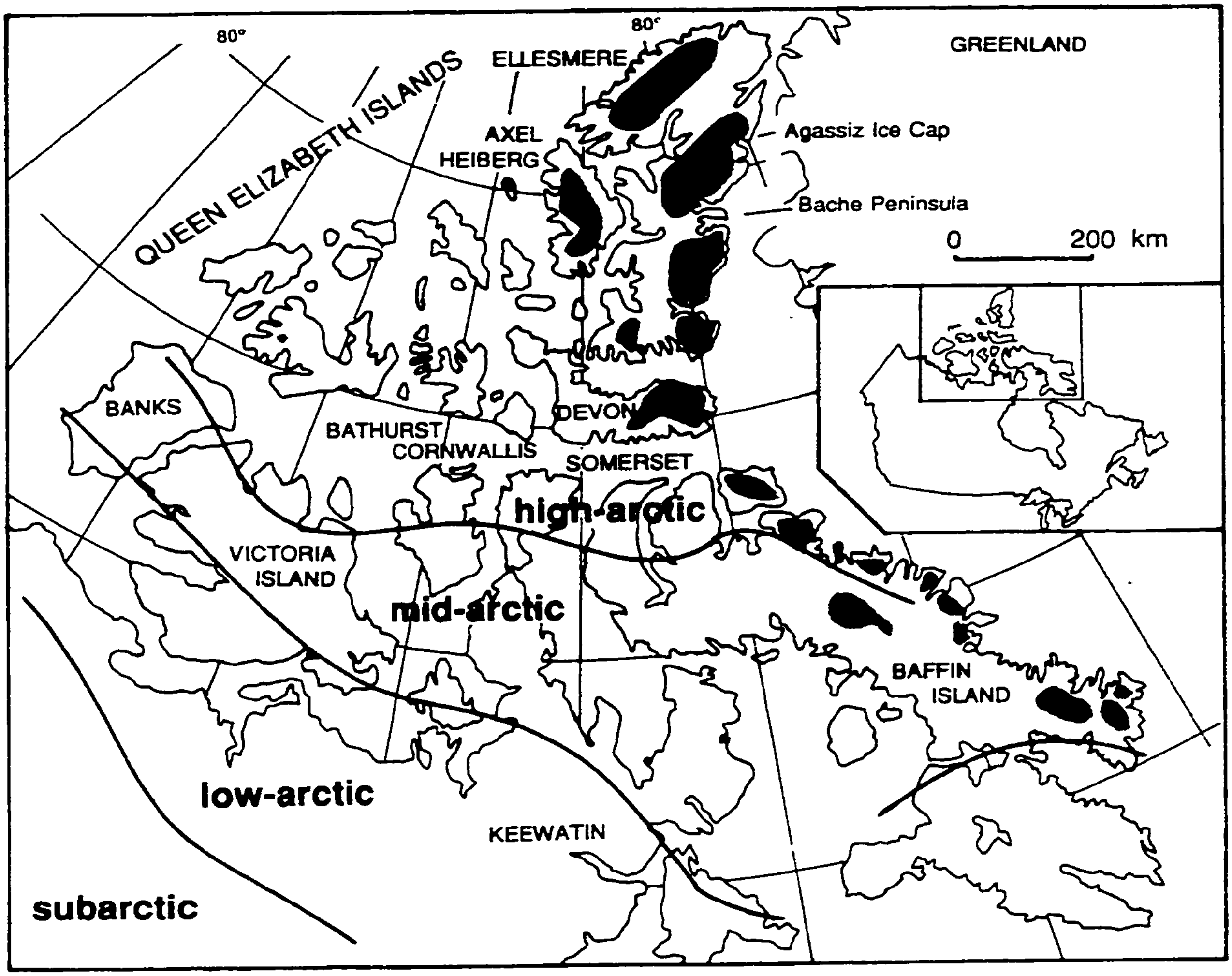

FIG. 1. Map of the Canadian Arctic Archipelago showing the location of sites mentioned in the text. Vegetation zones after Polunin (1951). Shading indicates ice caps.

tions of our methodology in such a context. Much of the landscape is a nearly unvegetated rock surface or covered with mosses and lichens (Bliss and Matveyeva, 1992); these surfaces do not contribute to the pollen rain. Many of the important plant families comprising the arctic vegetation are non-anemophilous and produce few pollen.

In the arctic, palynomorphs are found in a number of different deposits, including peats, lake sediments and ice cores. Each of these environments registers information related to climate or vegetation, whether from local, regional or extra-regional sources. The usefulness of these deposits for paleoenvironmental reconstruction depends on the presence of a sufficient accumulation to provide a history and on a secure chronology. The objectives of research determine which type of deposit to sample, although the limited availability of sites over much of the arctic may be the ultimate determining factor. We will be dealing with records of the terrestrial vegetation and their interpretation. Paleolimnological studies also show great potential (Smol, 1983; Blake *et al.*, 1992), but will not be discussed here.

Pollen assemblages from peats can be used to determine local plant succession as well as hydrological and climatic history. Studies of lakes are used to determine the regional changes in vegetation and climate. Interpretations from ice cores provide a broader regional signal and can possibly reveal past air mass movements. All of these deposits can contribute to knowledge of the glacial history (e.g. Blake, 1987; Blake *et al.*, 1992; Short *et al.*, 1994).

In peat deposits and in lake sediments, pollen accumulate from local, regional and extra-regional areas. In peats, a certain percentage of palynomorphs come from the local vegetation that accumulates in place, while others of these fossils originate from regional or extra-regional sources. Still others originate from Quaternary and Tertiary formations in the watershed; these are released by erosion or alteration from the enclosing deposits and are redeposited in peat and lake sediments by fluvial and aeolian processes. In lake sediments, the origin of a certain percentage of the pollen comes from just around the lake (local), while much comes from a larger area (regional). Tree pollen are found in arctic sediments well north of the treeline; this is pollen which was entrained higher into the atmosphere and carried many kilometers away from the source area (extra-regional). The relative proportion of regional or extra-regional derived grains in a pollen assemblage depends on the nature of the sedimentation processes as well as the type of vegetation surrounding the site.

One unique record of past environments restricted to arctic and alpine regions is the pollen preserved in ice cores. The snow at the top of some large Arctic ice caps can be viewed as a pollen trap whose content can be studied on a seasonal, annual and multi-annual basis. As all of the pollen grains found on ice caps have a distant origin, they can be used to study pollen transport processes in the atmosphere (Bourgeois *et al.*, 1985). Because there is no local input, the regional and extra-regional pollen are emphasized.

METHODS

Sampling

Peats. In frozen deposits, peat cores can be taken with a modified SIPRE corer equipped with tungsten carbide cutters. This allows the collection of frozen sections ranging from 10 to 30 cm long with an inner diameter of 7.5 cm with a recovery of nearly 100%. Coring is continued until the sediment recovered is dominated by mineral material: this distance varies from one site to another. On Bathurst Island, cores of approximately 3 m were collected (Blake, 1964, 1974); in the Fosheim Peninsula of Ellesmere Island (Fig. 2), the depth of the organic deposits varies between 30 and 300 cm (Garneau, *in preparation*).

If exposed faces of the peat deposits are available, sections (roughly 15 cm in height by 10 cm in width by 10 cm in thickness) can be cut from the cleaned face using a knife or saw. Complete sections of the exposed face can usually be obtained, although if still frozen, it may be necessary to sample the deeper portions with a corer to obtain the base of the deposit. Chainsaws have also been used (Nichols, 1974).

Lakes. Sampling lake sediments is normally done with a Livingstone sampler. In deeper lakes, other methods may need to be used (e.g. Reasoner, 1993). Due to the high clay and low organic content of the sediments, a driving rig of some kind is usually needed to obtain complete sequences and a casing is essential to prevent too much bending of the driving rods. To support the driving rig, sampling is usually done from an ice surface in the spring. The ice in many areas north of treeline is typically 2 m thick or more, and sufficient extensions are needed for the ice augers: hand-operated ice augers are usually suitable.

For statistical calibrations, the uppermost sediment is collected from lakes. These modern or surface samples can be compared to the surrounding vegetation or climate to search for modern analogues of the pollen assemblages or to derive transfer functions (e.g. Anderson *et al.*, 1989, 1991). These can be collected with a dredge, (if care is taken to sample only the uppermost sediment), or with various corers designed for taking undisturbed samples of the uppermost sediments. A simple plastic tube fitted with a piston can be used in some sites if a relatively stable platform is available and the water is not too deep (Wright, 1980).

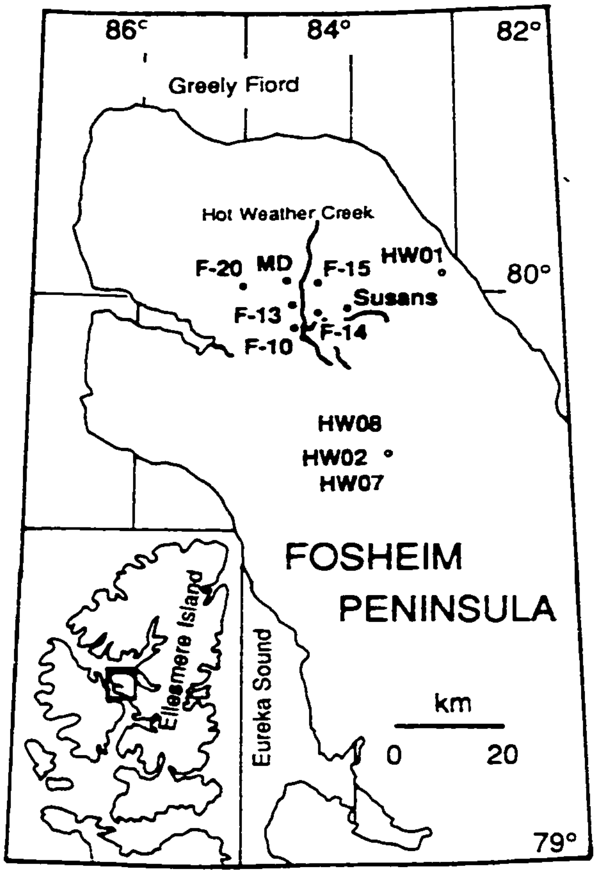

FIG. 2. Map of the Fosheim Peninsula, Ellesmere Island. F-10 to F-20 are sites where ^{14}C dates are presented in Table 1. Pollen and macrofossil dates from peat deposit MD are presented in Figs 4 and 9. Modern pollen assemblages from HW01 to HW08 and Susans Lakes are presented in Fig. 10.

Ice cores. For retrieving surface to bedrock cores, a thermal or electro-mechanical drill is used and, depending on the thickness of the ice, the drilling process takes a few weeks (as in the Canadian Arctic) to 2–3 entire field seasons (as in Greenland and Antarctica) to complete. These cores usually serve for numerous analyses. In studies from the Canadian Arctic, the maximum which has been reserved for pollen analysis is half of a core split lengthwise. For a 1.5–2.0 m increment, this represents a maximum of 8 l of water. The ice chips produced with the electro-mechanical drill can also be recovered and used for pollen analysis, as can the melt-water produced with the thermal drill. However, in the Canadian Arctic, use of the latter has proved to be problematic (Koerner *et al.*, 1988).

In Holocene ice, the pollen concentration is in the order of 5–25 grains l^{-1} (Bourgeois, *unpublished*) and core segments sometimes have to be combined to obtain an adequate pollen sum. As the annual layers get progressively thinner with increasing depth, this procedure is not recommended for Pleistocene ice core increments. For example, on Canadian ice caps, a 1.5 m core segment of Pleistocene ice can represent as much as a few thousand years. To obtain shorter increments in older ice but with a sufficient volume for pollen analysis, Koerner *et al.* (1988) melted, laterally, large ice samples within an existing borehole. Working from a relatively shallow borehole (130 m) for which an approximate age–depth scale and isotopic record had previously been produced, they determined beforehand the levels to be sampled. Starting from the bottom, they filled the borehole with ice chips and pollen-free water to the required depth and left the mixture to freeze so as to form a plug from which to start the melting. A heat source was then lowered down the hole and left to melt the ice for several hours, after which the water was pumped up to the surface and filtered. Using this method, Koerner *et al.* (1988) obtained between 30–40 l of water for each of the 40 cm increments they sampled.

To collect surface samples for modern deposition studies, the snow is sampled from the wall of a pit. From the snow stratigraphy, it is possible to identify the annual and 'seasonal' layers (Bourgeois, 1990). Large samples (15–30 kg) are usually necessary to compensate for the low concentrations found in the snow at the top of the ice caps.

Laboratory Methods

In our studies, standard methods of pollen and plant macrofossil preparation are used, as modified for arctic conditions. This is important, as a standardized pollen extraction methodology permits the comparison of sites from a wide geographic area.

Macrofossils. For macrofossil analysis, constant specified volumes of 50 up to 200 cm^3 (depending on the site and interval analyzed) are used. The material is heated in a 5% KOH solution to disperse the particles, washed and sieved through a 180 μm screen (Garneau, 1987, 1992). Macrofossils remaining on the screen are hand-picked under a stereoscope. Identification of some plant parts (e.g. plant epidermis) is done under the microscope.

Pollen and other microfossils. Palynomorphs are extracted from peats and lake sediments using almost the same methods. Although typically a volume of 1 cm^3 of sediment is used, in arctic studies 4 cm^3 are normally needed because of the low pollen concentration. The sediments are successively treated with hot 10% KOH, 10% HCl, HF and acetolysis treatment. Arctic lake sediments are more inorganic than those from temperate regions and two treatments of HF can be used: the sediment is left in cold HF overnight, centrifuged and decanted, then again treated in hot HF for 12–15 min (K. Hadden, *pers. commun.*). A crucial supplemental step is to sieve the sediment through 7μm nylon screens, after disaggregation in warm 7% sodium pyrophosphate, to remove the clays remaining even after the previous treatment (Cwynar *et al.*, 1979).

Peats are similarly treated, except the samples receive only cold HF overnight, and the sodium pyrophosphate treatment and sieving are not needed. Many peats and lake sediments have a high sand content which must be physically removed. The sample is placed in a clean beaker with distilled water, stirred and rapidly decanted to retain most of the material but allowing the sand to settle to the bottom. However, there may be a substantial loss of material, as shown by the low number of *Lycopodium* 'spike' grains remaining on the slides processed using this treatment.

Extracting pollen from ice cores. All snow and ice core samples are melted inside a large plastic bag and subsequently pressure-filtered through a 8 μm pore-size cellulose filter. A 47 mm diameter filter normally lets through approximately 15 l of melted snow or Holocene ice. Wisconsin-age ice and interglacial ice found near bedrock contain more mineral particles and filters must be replaced after 3 to 5 l. In the laboratory, the cellulose filter is normally soaked in HF for several hours before the acetolysis treatment. The first step is done in a teflon beaker, but the subsequent steps are done in a small filtering apparatus. This is to ensure that the small amount of pollen in the sample is not being lost through centrifuging and decanting. A 47 mm diameter polycarbonate (Nuclepore) filter with a 8 μm pore-size is used for this procedure. After the last rinse, the filter, which is transparent, is cut in half and each half is placed on a microscope slide with a few drops of glycerine. With this method the entire pollen content of the sample is counted. To correct for pollen loss during the field and laboratory procedure, a known amount of *Lycopodium* spores is added to the snow in the plastic bag. The loss of pollen in the sample can then be evaluated according to the number of *Lycopodium* found on the filter (Bourgeois, 1986; Koerner *et al.*, 1988).

Due to the low pollen concentrations in the samples, extreme care must be taken to avoid contamination. When samples must be exposed to the ambient air for any length of time, such as during HF treatment, it is done in a Class 100 clean air cabinet. The air in the laboratory is filtered, but periodically blanks are used for contamination control. This is particularly important for establishing the long-distance tree pollen concentration in the samples.

Dating of sediments. A chronology of peat deposits and lake sediments can be obtained by conventional or AMS radiocarbon dating, depending on the volume of the organic sediment available. As always, the samples must be carefully cleaned to remove roots and rootlets from the sediment (Olsson, 1974). A second step is needed in samples from some areas. Due to the presence and composition of unconsolidated Tertiary rock outcrops in many areas of the high arctic, carbon particles and other old organic material must be removed as much as poss-ible by hand under a stereoscope before sending the sample to the laboratory. Macrofossils can be removed from the sedimentary matrix and dated.

(a)

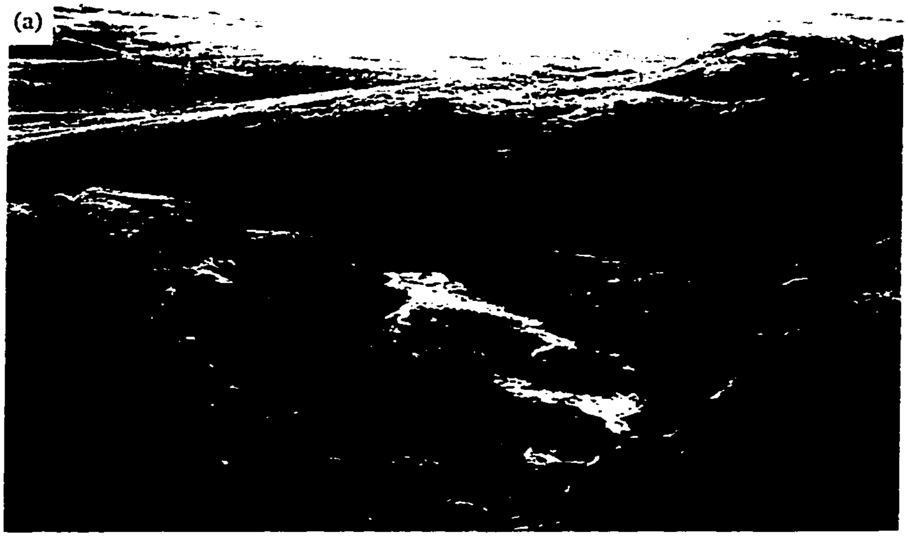

FIG. 3a. Peat sections exposed by fluvial erosion, and/or melting of tundra polygon ice wedges, Fosheim Peninsula, Ellesmere Island.

Ice core dating. Several techniques have been used to establish ice core chronologies, each offering a varying degree of accuracy. Ice flow modeling provides an age–depth chronology, but is rarely sufficiently accurate, except in ideal conditions or when no other methods are possible (Reeh, 1989). The most frequently used methods are based on the seasonal variations of various components such as isotopic composition, dust, acidity and trace substances studied in the ice cores (Hammer, 1989). These methods allow identification of annual layers and thus offer a precision comparable to that obtained by counting tree-rings or varves, but their accuracy largely depends on the amount of accumulation at the site and on the lack of subsequent disturbance. The lower the accumulation, the more likely the annual and seasonal snow layers will be disturbed by surface winds and the more rapidly the annual layers will thin out to the point of becoming indiscernible. On Canadian high arctic ice caps, precipitation rates are relatively low; nevertheless, these cores are felt to be accurate to within 10% for the past 7000 years (Fisher *et al.*, 1983). A major shift in the $\delta^{18}O$, which occurs a few meters above bedrock in Canadian ice cores, is correlated with a similar shift in Greenland cores. In the recently drilled ice cores from the Summit area of Greenland, this transition occurs at 1400 m above bedrock, and has been accurately dated at 11,500 BP (Alley *et al.*, 1993).

RESULTS AND DISCUSSION

Site Selection and Stratigraphic Problems

Peat deposits. A first problem in the analysis of arctic peat deposits is a geographic one and consists of finding suitable deposits. Although peat deposits have accumulated throughout the arctic, finding exposed sections that contain sequences spanning much of the Holocene is more difficult. Many of the reported sites contain only short sections of approximately 2500 years or less (e.g. Boulton *et al.*, 1976; Short and Andrews, 1980; Ovenden, 1988). Bradley (1990) associated this with the duration of a complete tundra polygon cycle where peat accumulates in a water-logged environment, then is subsequently lifted and exposed, resulting in a cessation of peat formation.

However, sections spanning 3000 to 4000 years have been eroded along gullies or exposed by the melting of tundra polygon ice wedges and sampling sites are frequently determined by the presence of these features (Fig. 3a). These can be up to 900 cm thick (Blake, 1974; Lowden and Blake, 1981; Ovenden, 1988; Bradley, 1990; Lafarge-England *et al.*, 1991; Garneau, 1992) and their rate of accumulation seems to have been greater in the past (up to 5 mm year^{-1}) than at present (0.3 to 1.4 mm year^{-1}; Ovenden, 1988, 1990). Ovenden (1988) suggested that these high rates of accumulation during the mid-Holocene indicate changes in past climates, but this interpretation remains controversial (Bradley, 1990; Lafarge-England *et al.*, 1991) principally because the results are derived from the dating of one or only few levels in the sections.

More recent work on the Fosheim Peninsula shows accumulation rates exceeding 1 mm year^{-1} during the mid-Holocene (Garneau, 1992). These rates are calculated from the thickness of the deposit and at least 2 and up to 6 radiocarbon dates per section (Table 1). Alternatively, LaFarge-England *et al.* (1991) calculated the net rate of organic matter accumulation (g organic matter cm^{-2} year^{-1}); thus excluding the seasonal aggradation of permafrost and the contribution of mineral layers which would influence the total thickness of the deposit. The resulting accumulation rate calculations, on 4 peat deposits composed almost exclusively of bryophyte remains and varying in thickness from 2 to 4 m, vary between 0.5 and 3.8 mm year^{-1} (LaFarge-England *et al.*,

TABLE 1. Accumulation of peat from Ellesmere Island. Location of sites is shown in Fig. 2.

Peat section	Depth cm	Age ^{14}C years BP	Lab no.
Fosheim-10	surface	3850± 80	Beta-51994
	surface	4990± 70	TO - 3037
	20–23	4500± 70	TO - 3038
	45–48	3930± 80	GSC - 5547
	100–103	4310± 80	GSC - 5554
	148–151	4900± 70	GSC - 5570
	285–288	7910± 120	GSC - 5066
	300	5070± 70	TO - 3039
Fosheim-13	20–23	930± 60	TO - 3040
	50–53	1890± 80	Beta-51993
	110–113	2650± 70	Beta-51992
	150–153	2930± 60	GSC - 5137
Fosheim-14	surface	4940± 80	Beta-51990
	50–53	3870± 80	UL - 1004
	58–61	2510± 70	GSC - 5575
	85–88	3210± 110	UL - 1036
	100–103	2700± 70	GSC - 5577
	120–125	3140± 110	GSC - 5586
	135–138	3860± 80	Beta-51989
Fosheim-15	surface	6250± 70	TO - 2677
	surface	1910± 60	GSC - 5280
	180–183	5830± 100	Beta-51988
	200–203	5420± 100	UL - 1009
	280–283	7120± 80	GSC - 5180
Fosheim-20	35–38	5560± 80	UL - 1006
	90–93	7910± 100	UL - 1007
	95–98	9220± 140	UL - 1008
	180–183	9260± 130	UL - 1037
	190–193	12,750± 240	UL - 1005
PT-1	21–26	2430± 85	S -2439
	107–122	2995± 85	S -2440
	218–223	3450± 90	S -2441
	ca 323	3760± 90	S -2442
PT-2	23–30	760± 75	S -2449
	100–105	1430± 85	S -2450
	207–212	2800± 200	S -2451
PT-3	17–22	4720± 100	S -2443
	205–210	5165± 100	S -2444
	400–409	6315± 110	S -2445
	400–409	6410± 50	TO - 284
PT-4	12–17	3440± 90	S -2446
	100–105	4250± 95	S -2447
	203–211	5465± 105	S -2448
	203–211	5460± 80	TO - 283

1991: Table 1) for peats accumulating between 6500 and 760 BP on Ellesmere Island. Although these authors observed an interval of high accumulation rates during the mid Holocene (zone PT 3), they conclude that these rates are due to site-specific factors and not necessarily induced by climate changes.

Ovenden (1988, 1990) and Garneau (1992) have shown that younger sections with basal dates less than 3000 BP, have accumulation rates between 0.4 and 1.3 mm year^{-1} in deposits composed principally of sedge remains. The importance of bryophyte remains has decreased in the late-Holocene (Tarnocai, 1978; Garneau, 1992). In a number of poorly drained depressions on the Fosheim Peninsula the sections colonized by bryophytes (principally *Campylium stellatum*, *Calliergon* cf *cordifolium*, *Scorpidium scorpioides*, *Meesia triquetra*, *Drepanocladus lycopodioides* var *brevifolius*, *Drepanocladus* sp and *Bryum* sp) had accumulations of less than 30 cm.

Only few palynological studies have been carried out in peats (Hegg, 1963; Terasmae *et al.*, 1966; Colinvaux, 1967; Jankovska and Bliss, 1977) and the results show few changes through time in the pollen spectra. This might be explained in part by the relatively large intersample interval (10–20 cm) and the short period of time covered by the analyses. When detailed paleoecological analyses of peat deposits are carried out, the results allow the reconstruction of local paleoenvironmental conditions. One of the most important questions is to explain under what conditions such large accumulations of organic matter could have formed, in areas where today little peat is accumulating. The dynamics of this accumulation are associated with a combination of topographic and hydrological factors which are regionally associated with climate (Garneau, 1992). However, there are a number of problems of sampling, analysis and interpretation that must be dealt with before we can use this information for paleoenvironmental reconstructions.

Another problem, associated with the presence of permafrost, is the perturbation of the stratigraphic horizons. One origin for peat accumulation in the high arctic is associated with the temporal dynamics of 'low-centred' tundra polygons (Bradley, 1990; Seppälä *et al.*, 1991). Over time, ridges develop over the ice wedges forming the boundaries of the polygons, with depressions in the polygon centre. As the permafrost partially melts in these depressions, shallow pools are formed. Slightly-decomposed organic matter accumulates in these pools. Over time, the permafrost dynamics change low-centred polygons into high-centred polygons where peat sections are lifted and then exposed (Bradley, 1990). This then leads to the possibility of stratigraphic inversions, caused by the raising of the borders of the polygons (ridge formation), thickening of the centres, followed by a lifting of the centres as they dry. To detect hydrological changes through peat stratigraphy, a detailed analysis of the ecological succession and geomorphological processes is necessary and to interpret regional climate change, this local succession must first be accounted for.

Fluvial erosion processes make sections available, but may also transport allochthonous sediments, possibly with enclosed fossils, and redeposit them among the horizons of the peat sections. Furthermore, as the permafrost melts, slumping can occur causing the originally flat horizons to tilt and, in the extreme, collapse (Fig. 3b). Finally, as the permafrost melts, following exposure of sections by streamflow, lateral movement of meltwater through the deposit may cause redeposition of some of the enclosed fossils from one level to another.

Eolian processes (including niveo-eolian) can also transport some allochthonous material to the peat. Interbedded sections of peat and sand have been described in arctic deposits (Boulton *et al.*, 1976; Pissart *et al.*, 1977; Andrews *et al.*, 1979; Short and Jacobs,

(b)

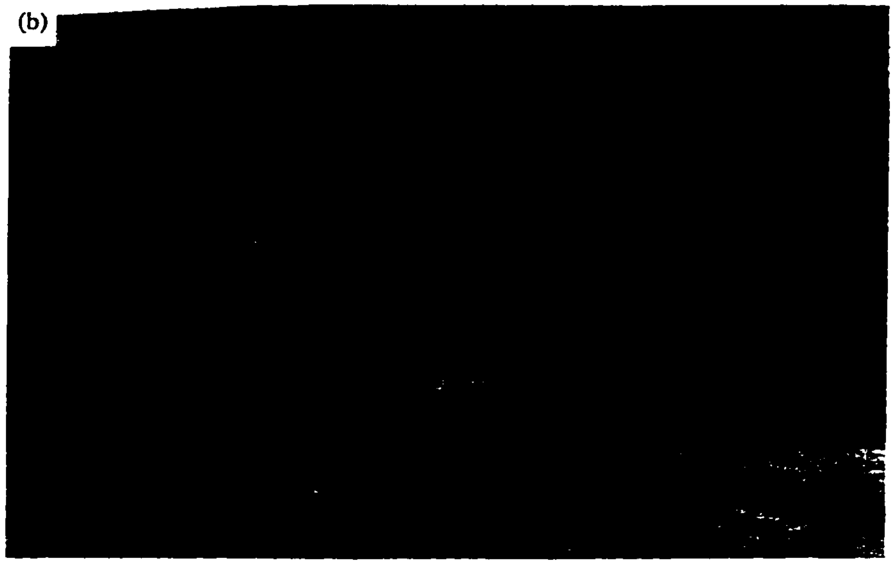

FIG. 3b. Slumping caused by erosion of polygons and tilting of original flat horizons, Fosheim Peninsula, Ellesmere Island.

1982; Ovenden, 1988; Seppälä *et al.*, 1991; Garneau, 1992). Palynomorphs as well as plant fragments can be incorporated into peat deposits at the same time as the sand particles. Recurrent sand layers in peats and lake sediments could be used, if well dated, to indicate dry periods or paleowind direction (Ovenden, 1988).

One alternative to avoid some of these problems of contamination along exposed faces is to collect non-eroded samples with a corer. However, this introduces other problems: it is then not possible to determine in advance the depth of organic deposit, nor to identify stratigraphic perturbations caused by the temporal dynamics of tundra polygons. Furthermore, more elaborate equipment is required for collecting cores (Fig. 3c), and the samples must be kept frozen until sub-sampled in the lab to avoid possible perturbation of the stratigraphy.

When account is taken of these problems, results can provide a detailed reconstruction of the local conditions that influenced the development and accumulation of peat deposits in the arctic. Peat deposits are sensitive to local hydrological variations, which, in turn frequently correspond at a regional scale with climatic changes (for example, an increase in precipitation or longer snow-cover seasons). The combination of ecological conditions permitting continuous peat accumulation is closely related to the combination of moisture from the atmosphere, soil and vegetation cover (Barber, 1981).

Based on a macrofossil analysis of four peat deposits, Lafarge-England *et al.* (1991) have distinguished two origins for peat accumulation: topogenic and soligenic in which the controlling factors are two distinct hydrological regimes, dependent on the local physiography. The

(c)

FIG. 3c. Coring frozen peat deposit with SIPRE corer.

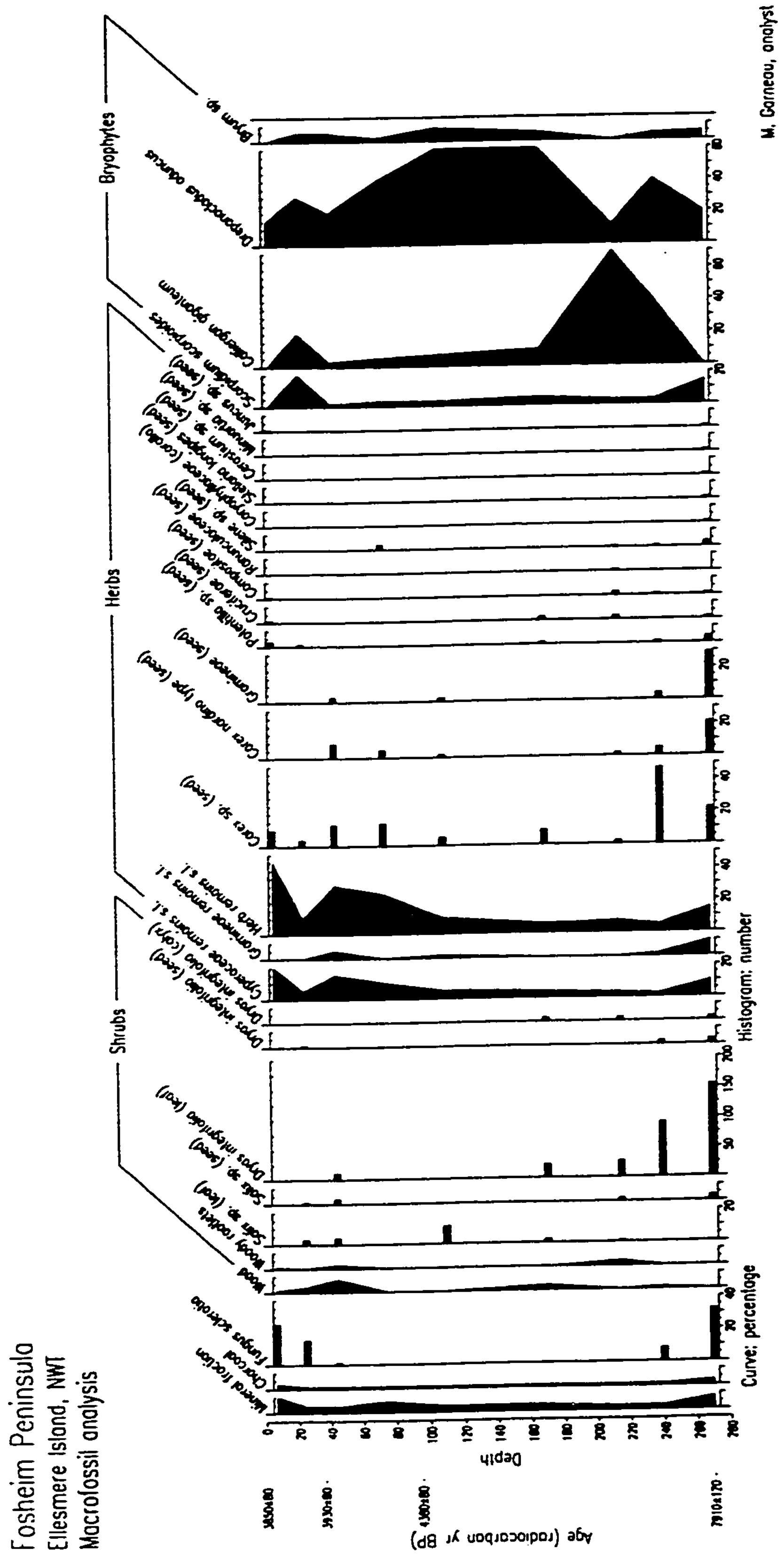

FIG. 4. Macrofossil diagram of peat deposit on the Fosheim Peninsula, Ellesmere Island, after Garneau (1992).

chronology of these deposits indicates a minimum age of 6400 BP for the beginning of peat accumulation in the region. According to these authors, the bryophytes are not as sensitive to macroclimatic conditions as they are to microclimatic (Miller, 1980; Crum, 1972), and the paleobryological record of the soligenous fens supports this conclusion. They associated an alternation from dry to humid phases in the peat horizons with the temporal dynamics of tundra polygons.

On the Fosheim Peninsula, Garneau (1992) associated hydrological and geomorphological processes with peat stratigraphy. Peat accumulation began in a wet depression south of a small lake following the postglacial emergence of the site at around 7500 BP. The cessation of peat growth around 4000 BP resulted from the draining of the lake and the downcutting of the deposit by the lake spillway. A preliminary macrofossil analysis (Fig. 4) reveals an open meadow dominated by grasses and sedges recording the pioneer stage following emergence of the region. The physiognomy of the meadow was subsequently modified by the formation of low-centred polygons. Bryophytes colonized the resulting wet depression and peat accumulated. Other dated, dissected sections reveal an increase in fluvial erosion in the region during this period (Garneau, *in preparation*). These erosion processes thus explain the initiation and termination of peat accumulation, and this provides an alternative interpretation to that of natural evolution of high-centred polygon dynamics (Bradley, 1990).

Lakes. There are numerous lakes in the high arctic, although these are not necessarily distributed throughout the landscape. Large areas contain no lakes. Lakes of all sizes and depths are available, and the selection of sites is a function of the questions to be asked and the equipment available.

As with peat deposits, lake sediments are also affected by periglacial activity. Thermokarst lakes and others with obvious disturbance must be avoided. Due to the long and cold winter the lake ice north of treeline is consistently around 2 m thick (but there are some exceptions; Blake, 1981, 1989). Thus, lake depth must exceed 3 m to avoid freezing of the sediment and possible mixing of the bottom sediment (Nichols, 1967). The entire landscape is underlain by continuous permafrost; in some shallow lakes, the sediment apparently freezes from below, and when cores are taken sections are broken into chunks (see Funder and Abrahamsen, 1988). This seems not to disturb the stratigraphy, however, more work is needed to ensure that this is the case. One solution may be to core even deeper lakes, 10 m or more, as the larger volume of the water mass may keep the sediment under the lake unfrozen. However, frozen sediment in shallow ponds has been sampled and the stratigraphy is apparently not destroyed (Blake, 1978, 1982). Recent work suggests that these ponds may contain undisturbed diatom stratigraphies (Douglas *et al.*, 1994). Coring shallow ponds may provide an alternative to coring small lakes to obtain sedimentary sequences.

Ice cores. For pollen studies of ice cores, it is preferable to be close to areas with a relatively dense vegetation cover as this usually means a higher pollen concentration and a better chance of detecting changes in the pollen record. However, site selection for the location of ice cores is usually determined by the glaciological community involved in the coring, and the best sites are at the top of an ice divide where lateral flow is minimized. The length and sensitivity of the record being sought as well as the probable sensitivity of the site to register various environmental changes are the determinant factors. Bedrock topography, ice thickness and accumulation rates must be considered in the choice of a coring location. The higher the accumulation rate, the longer it will take for the annual snow layer to be smoothed out in the ice core. The amount of melting occurring during the summer is also important as too much melt will obliterate parts of the record. A limited amount of melt, however, can be a good paleoindicator of summer temperatures (Koerner and Fisher, 1990).

There are potential stratigraphic problems in ice sheets. Melting and refreezing at the base of an ice cap may have occurred in the past when the ice was thicker. This may produce a discontinuous record in older ice, as could folding or distortion of the bottom ice layers as they move over bedrock (Koerner *et al.*, 1987). However, because of the multi-disciplinary approach to ice core studies, these unconformities can usually be identified.

Pollen Representation and Transport

Apart from the potential of stratigraphic inversions discussed above, a number of factors related to pollen production and transport to the site need to be considered in pollen analysis of these sediments. These include the composition of the sediment that contains the pollen, the low concentration of pollen and the large proportion of long-distance transported grains (termed 'exotic').

Pollen concentrations. One problem with arctic palynology is the low concentration of palynomorphs in lake sediments and in peats. Total pollen concentrations in mid-arctic lake surface-sediment samples are typically

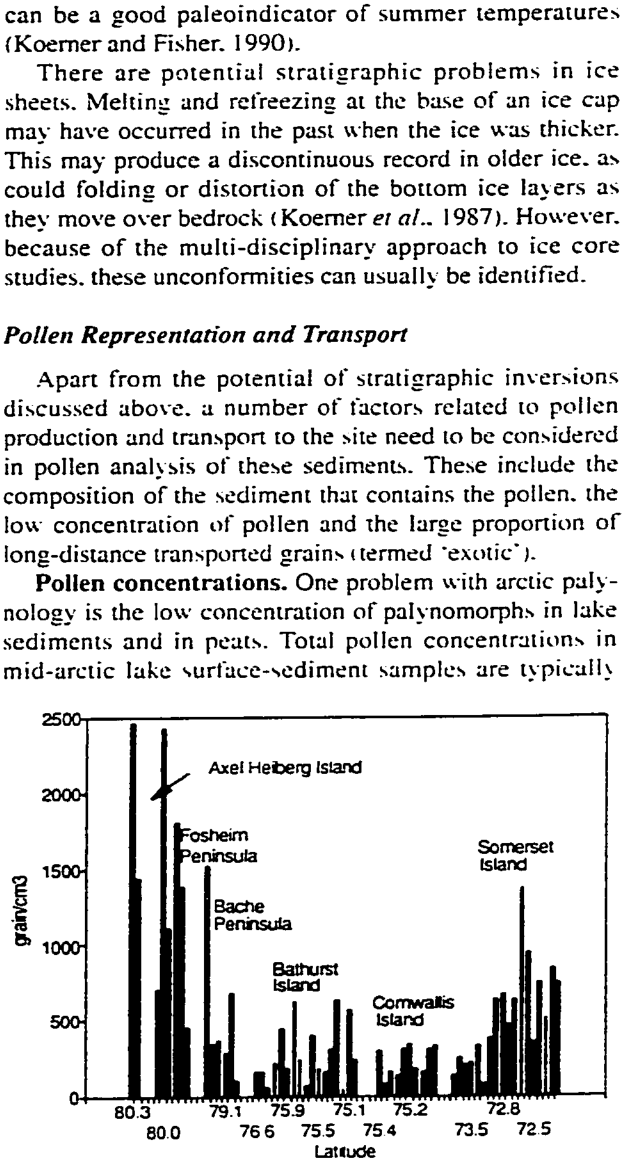

FIG. 5. Pollen concentration from a series of surface sediment samples from lakes in the Canadian arctic. These values are calculated from one slide per site, regardless of the total sum

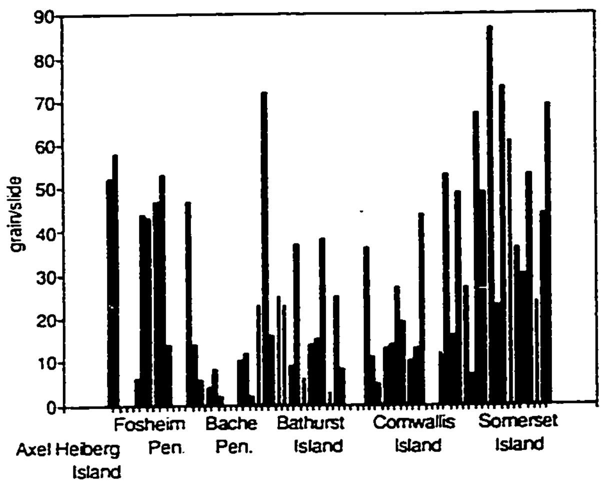

FIG. 6. Number of pollen counted per slide from same sites shown in Fig. 5.

less than 1000 grains cm^{-3}, while in the high-arctic they can be less than 500 grains cm^{-3} (e.g. Blake *et al.*, 1992; Gajewski, *submitted*). Higher concentrations are found in surface samples from the mid-arctic of Banks Island (Ritchie *et al.*, 1987).

Figure 5 shows the pollen concentration (grains cm^{-3}) from a series of lake surface-sediment samples from several areas of the arctic. The pollen concentrations are orders of magnitude lower than those from boreal or temperate sites. This is caused by the low pollen production and the low density of arctic plants on the landscape. Larger sediment samples are thus needed to obtain sufficient pollen.

From a practical point of view this low concentration means that it is difficult or impossible to do palynological studies at many sites. Slides rarely contain more than fifty grains (Fig. 6). By comparison, samples from the boreal zone can easily approach 500 to 1000 grains per slide. Since a 'normal' pollen sum consists of a minimum of 300 to 500 grains, and it takes several hours to count one slide, it is impossible to obtain pollen from many high-arctic sites. Thus, only a small percentage of sites in any field season will produce useful diagrams — from our experience only a third of the total sites. The resultant pollen sums are typically low in pollen diagrams from arctic regions, often as low as 100 grains or less (e.g. Boulton *et al.*, 1976; Short and Andrews, 1980), and this is unavoidable in some cases. Larger sums are well worth the effort, as this gives a greater probability of encountering rare pollen grains, which can provide important paleoenvironmental information. For example, in their analysis of the Beringia steppe–tundra question, Cwynar (1982) and Ritchie (1982) have shown the detailed ecological interpretation that can be made with large pollen sums and detailed taxonomic resolution.

Some preliminary observations suggest that there are differences in concentrations across the arctic (Fig. 5) which are related to the vegetation density. Lowest pollen concentrations are associated with the sparse vegetation on Cornwallis and Bathurst Islands in the central arctic. Areas of more dense vegetation have a correspondingly higher pollen concentration, for example, the high concentrations on Ellesmere Island correspond to the 'oasis' on the Fosheim Peninsula of Ellesmere Island as defined by Edlund and Alt (1989). Along a transect of lakes on western Somerset Island, there is a decrease in pollen concentration at 73°N (Fig. 5; Gajewski, *submitted*) which coincides with the transition from mid-arctic to high-arctic vegetation zones (Woo and Zoltai, 1977).

Changes in pollen concentration over time in some diagrams may be important paleoclimatic signals. Presently, our knowledge of high-arctic Holocene paleoclimate is limited and based almost entirely on evidence other than pollen. Bradley (1990) reviewed Holocene proxy climate records for the Queen Elizabeth Islands and concluded that summer temperatures in the early Holocene were warmer or at least as warm as the present, and minimum temperatures for a late-Holocene cooling were reached between 400 and 100 BP. The different records available do not agree on the timing of this warm period. The few pollen records available show similar tendencies, with decreasing pollen concentrations interpreted as cooling temperatures. In east-central Ellesmere Island, Hyvärinen (1985) found a decrease in pollen concentration starting from around 6700 BP and reaching lowest values after 3500 BP. Similarly, a pollen diagram from Agassiz Ice Cap also shows a decrease in the 'regional' pollen concentration after 3000 BP and a minimum is reached around 250–150 BP (Bourgeois, 1986). Further south, on Somerset Island, sediment older than 6000 BP in two lakes (one from the high-arctic and one from the mid-arctic) contains pollen concentrations twice the present value (Gajewski, *submitted*). To further analyze these data and compute pollen accumulation rates (PAR; influx) requires good chronological control (discussed below), and this has been a consistent problem in arctic paleoecology. Much work remains to determine the variations in pollen concentration and accumulation rates at regional and local scales, and the relationship of pollen accumulation in sediments to plant density on the landscape.

Long distance transport. Another aspect of high arctic pollen analysis is the significant proportion of long-distance transport of pollen grains from forested regions. Long-distance transport of pollen is a feature of all regions, but this is particularly a problem in the arctic, due in part to the small local pollen deposition and also to the large amounts of pollen transport of boreal taxa by the wind (Birks, 1973).

A network of surface sediment samples from across the arctic are not yet available so regional differences in pollen transport cannot be determined. Several studies of snow, moss pollsters and pollen traps have been done, however, and they give an indication of the areal extent of this long-distant transport. Figure 7 shows a preliminary distribution of several pollen types based on four

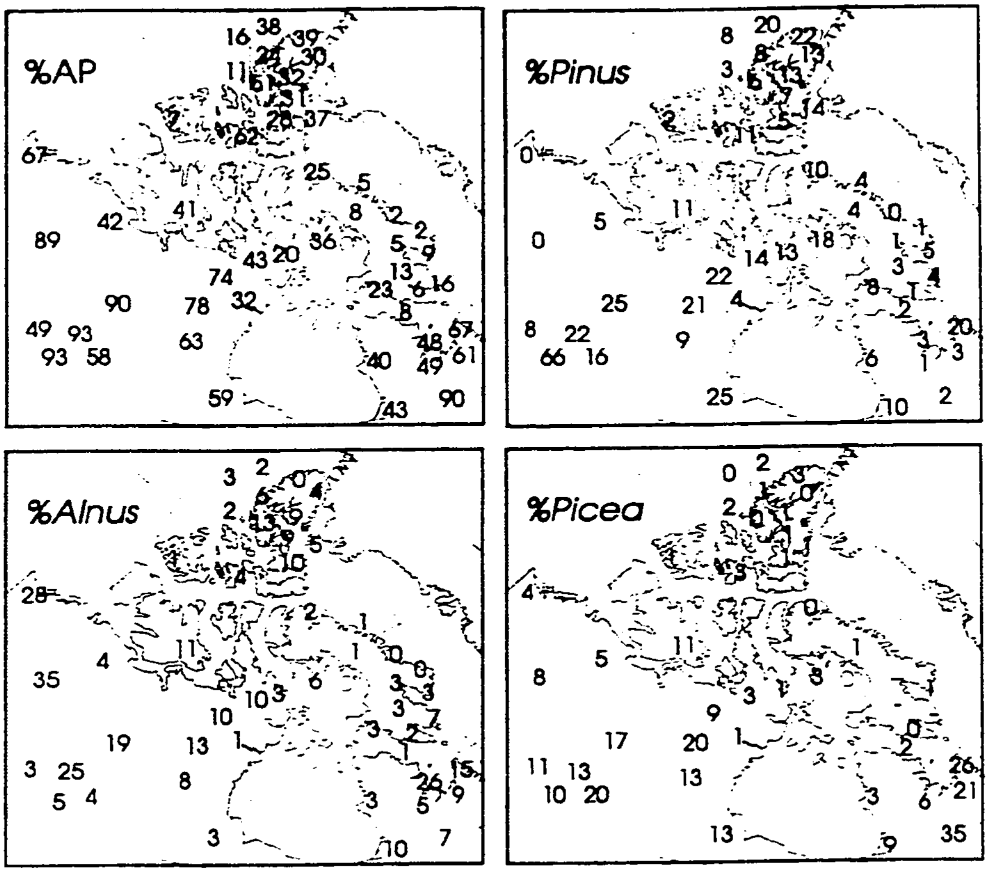

FIG. 7. Modern pollen transport in the Canadian Arctic. See text for explanation.

different studies. These include 15 snow samples (Bourgeois *et al.*, 1985) from the northern portion of the Arctic Archipelago, 20 pollen traps from the boreal zone and central arctic (petri-dishes in Stevenson screens: Ritchie and Lichti-Federovich, 1967), median pollen percentages of moss pollsters from eight regions of Baffin Island (Short *et al.*, 1985), and mean percentages in moss and lichen polsters from Keewatin, Baffin Island and northern Québec (we computed regional means from the 39 values published in Elliot-Fisk *et al.*, 1982). Although these studies are not totally comparable, because of the different trapping efficiencies of the pollen 'sampler' (snow, petri dishes, moss pollsters) and the different sampling intervals (one year for petri dishes; one to several years for the snow samples and moss pollsters), the maps show that arboreal pollen can make up half of the pollen well to the north of the limit of these species.

To further pursue this topic requires a network of samples from across the arctic, analyzed from uniform deposits and using similar methods. This will permit the analysis of the spatial distribution of both the arctic and exotic pollen taxa. This latter may be of interest because of the possibility of interpreting changes in wind direction during the Holocene. Nichols *et al.* (1978) and Fredskild (1985a) suggested that peaks of tree taxa in arctic pollen diagrams indicated periods of increased southerly wind flow, however, this conclusion has been questioned by Barry *et al.* (1981).

Contamination of Sediments by Material from Older Deposits

In parts of the Canadian high arctic, an added problem is the presence of outcrops, frequently unconsolidated, of pre-Quaternary deposits that may be contributing pollen to Holocene and Wisconsin age sediments. This exotic pollen is in addition to the component transported from long-distance.

Vast regions of the high arctic are covered with poorly consolidated clastic rocks deposited during the Cretaceous and the Tertiary (Hodgson, 1985; Fig. 8). These deposits often contain organic beds (coal strata, amber, wood litter) from which macrofossils and palynomorphs can be extracted. Palynomorphs are the most frequently rebedded; macrofossils other than wood are less likely to be preserved (Matthews *et al.*, 1986).

Palynomorphs from the pre-Tertiary are sufficiently different from modern types to be identified in more recent deposits, and these have, indeed, alerted us to this problem of contamination. For example, Table 2 shows the atmospheric pollen collected in traps during four rainstorms at Hot Weather Creek on the Fosheim Peninsula

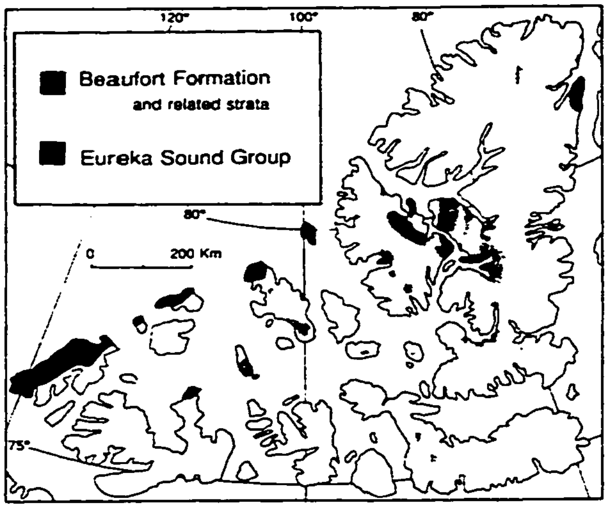

FIG. 8. Distribution of late-Cretaceous and Tertiary outcrops in the Queen Elizabeth Islands. Based on Fyles (1990), Miall (1991) and Kalkreuth *et al.* (1993).

of Ellesmere Island. Ten percent of the pollen and spores could be identified without a doubt as pre-Quaternary (for example, large, dark trilete spores). However, this ten percent does not include pollen such as *Pinus* or *Picea* which could be either modern pollen grains transported from the south, or redeposited pre-Holocene grains. This shows evidence of eolian transport of reworked old palynomorphs that could subsequently be rebedded in modern deposits. In a lake, the number of redeposited pollen could be higher as it could also include redeposited pollen transported to the lake by streams, whose banks may be eroding organic beds of pre-Quaternary deposits. This could also apply to peat sections eroded by streams.

The extent of the contamination of Holocene sediments from Tertiary deposits is more difficult to assess when the deposits contain well-preserved plant macrofossils and pollen from vegetation that still exists today in the forested regions of North America. Outcrops of the Eureka Sound Group (late Cretaceous–early Tertiary) are common throughout the arctic (Miall, 1991; Kalkreuth *et al.*, 1993; Ricketts, 1994) and are particularly widespread in the intermontane zone of Ellesmere and Axel Heiberg Islands (Fig. 8). Palynological evidence suggests that during the Paleocene and Eocene the region supported sub-tropical to temperate vegetation (Miall, 1991; Ricketts, 1994). These assemblages are dominated by pollen of the Taxodiaceae family (i.e. *Metasequoia, Glyptostrobus, Sequoia, Taxodium*) with lesser contributions of Pinaceae pollen and Pteridophyte spores (Kalkreuth *et al.*, 1993; McIntyre, 1991; Ricketts, 1994). The presence of angiosperm pollen (some from extinct species) is variable, with the higher values in younger strata.

Several late-Tertiary (Miocene, Pliocene) sites, containing organic beds rich in wood, other macrofossils and

TABLE 2. Pollen in rain samples from Hot Weather Creek, Fosheim Peninsula, Ellesmere Island during June and July, 1988

Sample #	1		2		3		4	
Date	26 June		28 June		4 July		7–8 July	
Volume H_2O (ml)	80		340		45		75	
	#	%	#	%	#	%	#	%
Betula	1	0.3	9	27.3	2	1.6	–	–
Juniperus	–	–	–	–	–	–	1	0.6
Picea	–	–	1	3.0	1	0.8	–	–
Pinus	1	0.3	1	3.0	2	1.6	1	0.6
Ambrosia	–	–	–	–	1	0.8	–	–
Ericaceae	1	0.3	2	6.1	–	–	10	5.7
Salix	284	91.3	6	18.2	50	40.3	108	61.0
Dryas	–	–	–	–	4	3.2	6	3.4
Cruciferae	1	0.3	–	–	–	–	–	–
Cyperaceae	4	1.3	2	6.1	47	37.9	23	13.0
Equisetum	1	0.6	–	–	–	–	–	–
Gramineae	4	1.3	1	3.0	2	1.6	2	1.1
Oxyria	2	0.6	–	–	3	2.4	4	2.3
Pedicularis	2	0.6	–	–	3	2.4	6	3.4
Rosaceae	–	–	–	–	1	0.8	1	0.6
Saxifragaceae	1	0.3	–	–	4	3.2	2	1.1
Triglochin(?)	–	–	–	–	–	–	2	1.1
Botrychium(?)	1	0.3	1	3.0	–	–	1	0.6
Lycopodium	–	–	1	3.0	–	–	–	–
Monolete spore	–	–	2	6.1	–	–	–	–
Trilete spore	1	0.3	2	6.1	–	–	–	–
Indeterminable	8	2.6	4	12.1	4	3.2	8	4.5
Unknown	–	–	1	3.0	–	–	1	0.6
Pollen sum	311		33		124		177	
Pre-Quaternary	1		34		28		4	
Other spores	84		144		71		178	

palynomorphs, have been located in the Arctic Archipelago (Matthews and Ovenden, 1990; Fyles, 1990; Matthews *et al.*, 1986). The Beaufort Formation, limited to the western arctic (Fyles, 1990), is probably the most well known, but other related sites are found in the eastern arctic (Fig. 8; see also Matthews and Ovenden, 1990). The older of these deposits contain a relatively large amount of hardwood pollen and macrofossils (i.e. *Quercus, Juglans, Carya, Pterocarya, Ulmus, Carpinus*), but boreal taxa such as *Picea, Pinus, Alnus* and *Myrica/Comptonia* become progressively more important in younger deposits. Tundra species also appear towards the end of the Tertiary (Matthews *et al.*, 1986; Matthews and Ovenden, 1990; Bennike and Bocher, 1990). Interestingly, these deposits often contain large numbers of recycled Late-Cretaceous and Paleocene palynomorphs (Matthews *et al.*, 1986; Whitlock and Dawson, 1990; McIntyre, 1991).

Based on the types of pollen (predominance of Taxodiaceae and several extinct species) and on published photographs (Kalkreuth *et al.*, 1993; McIntyre, 1991), it would seem that contamination from early Tertiary deposits would be apparent, or at least suspected, when occurring in a Holocene or modern deposit. However, pollen grains originating from late Tertiary

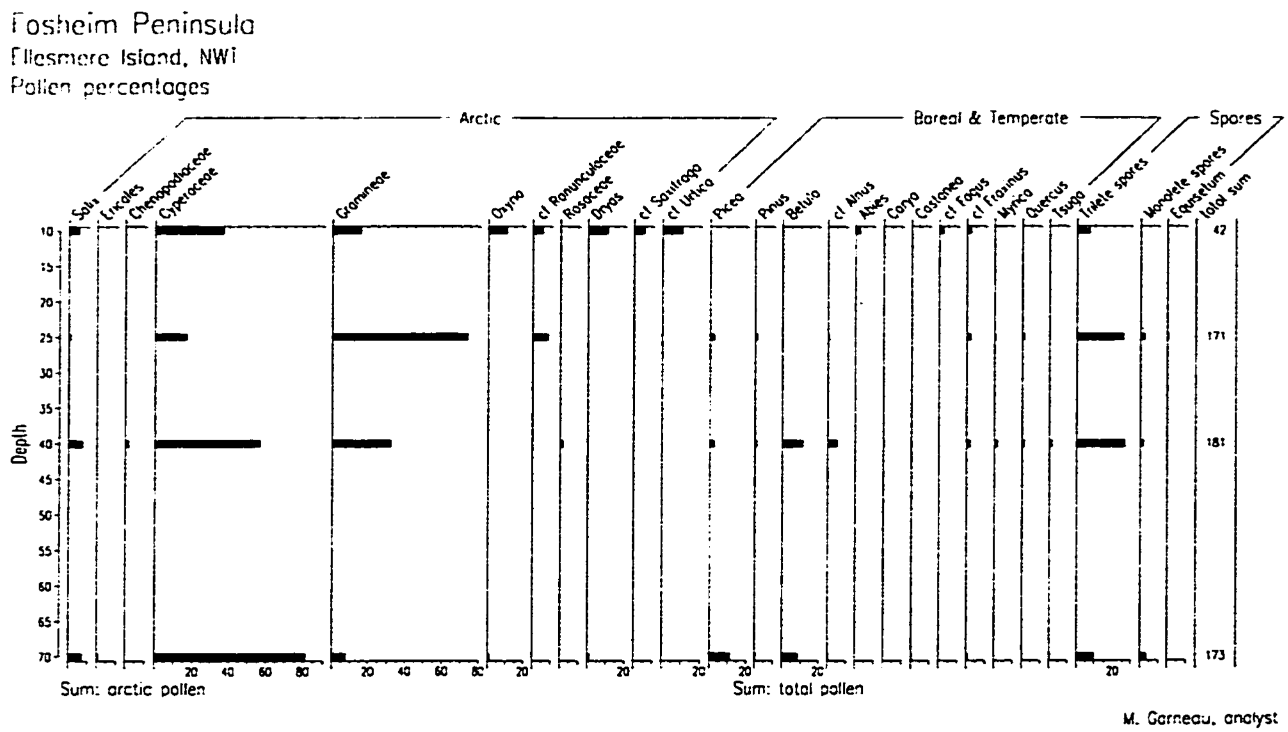

FIG. 9. Percentage pollen diagram from a peat deposit on the Fosheim Peninsula. See also Fig. 4.

deposits from several high arctic localities do not appear sufficiently different morphologically from modern pollen to be clearly identifiable as such (Fyles and Matthews, *unpublished data, pers. observations*). They do not appear 'significantly older' under the light microscope, although this may not be true in all cases. This means that it may be difficult to distinguish old pollen redeposited by aeolian and fluvial processes into a Quaternary deposit from the 'modern' long-distance component.

Some pollen analyses from Ellesmere Island illustrate this problem. Many of the grains of exotic pollen from a Holocene peat deposit (Garneau, 1992) were degraded and are presumed redeposited from pre-Holocene regional deposits (Fig. 9). In surface sediment samples from several lakes on the Fosheim Peninsula of Ellesmere Island, the majority of the pollen grains were tree and shrub taxa not living in the present-day arctic islands (Fig. 10). However, it is uncertain if these grains are modern pollen transported from the south or are ancient pollen redeposited from the Eureka Sound or younger outcrops present in the area.

TABLE 3. Percentage of pre-Quaternary palynomorphs found in the annual snow layer at the top of Arctic ice caps

	Year			
Ice Cap	1982	1985	1987	1989
Agassiz	0%	0%	0%	0%
Meighen	26%	0%	3%	57%
Melville South	46%	33%	-	31%

Presently on large ice caps, contamination by older deposits seems to be less of a problem, but some of the smaller ice caps of the western arctic can show a large proportion of pollen and spores that can be certainly identified as pre-Quaternary and probably pre-Tertiary origin (Table 3). Koerner *et al.* (1988) presented a pollen concentration diagram from the basal ice of the Agassiz Ice Cap of Ellesmere Island. Based on a high concentration and on the type of pollen found in the samples closer to bedrock, together with higher oxygen isotope values, they concluded that the ice cap probably began its growth toward the end of the last interglacial, when the climate was still warm to the south. This would have accounted for the high concentration of hardwood pollen found in the ice samples. Surprisingly these species did not dis-appear in the Wisconsin-age ice samples, although their concentrations diminished slightly, and this was accounted for by a stronger meridional circulation during the ice age, combined with lower ice accumulation rates which concentrated the pollen grains. A percentage diagram of the same samples shows over 50% exotic pollen (Fig. 11). An alternative hypothesis is that some of these grains (i.e. hardwood species) are reworked Tertiary pollen from deposits closer to the ice cap. This does not change the original interpretation, however, as the larger amounts of pollen and the higher dust content in the ice samples closer to bedrock still is evidence of a very small ice cap at the time of their deposition.

Separating long-distance transport from rebedded Tertiary pollen remains a problem. When Cretaceous or Early Tertiary palynomorphs, such as species of

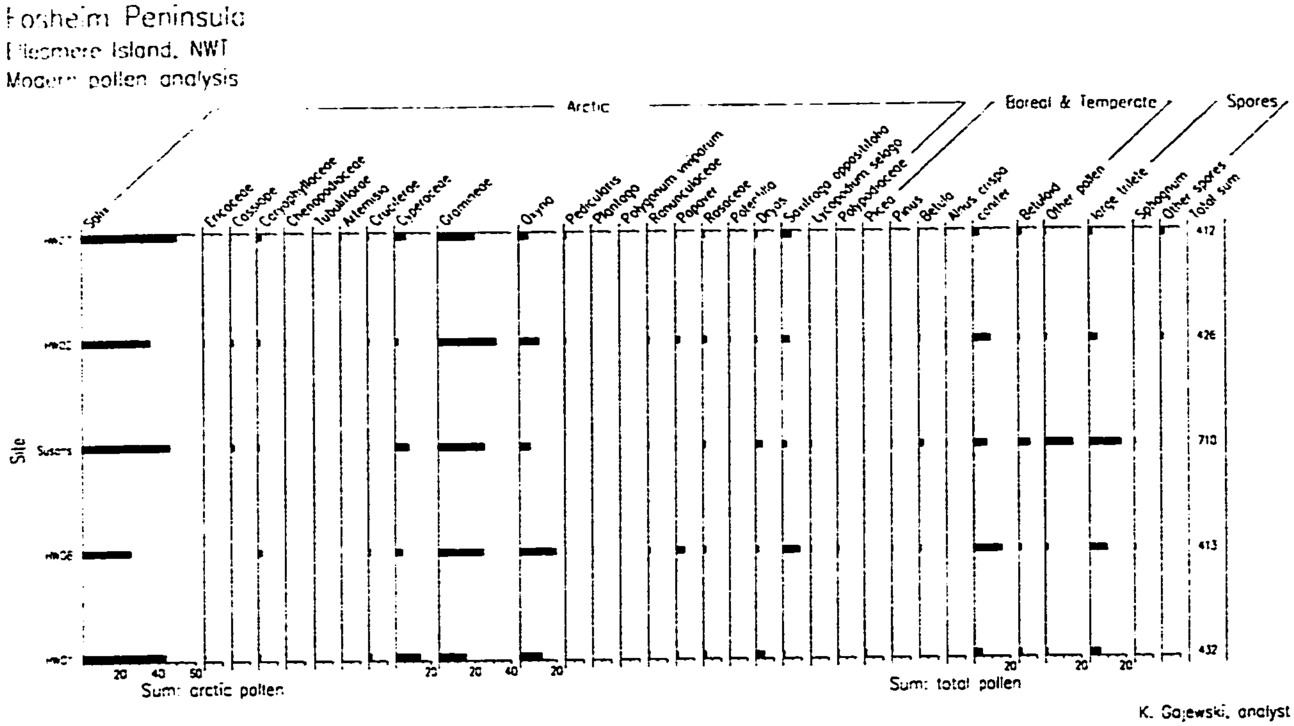

FIG. 10. Percentage pollen diagram of modern pollen samples from lakes on the Fosheim Peninsula. Location of sites is shown in Fig. 2.

Taxodiaceae or large trilete spores are found, they can be interpreted as evidence of contamination. However, their absence does not preclude the possibility of it. The presence of relatively large numbers of 4-pored *Betula* pollen, or of *Myrica*, *Larix* or *Tsuga* pollen which are rarely found north of treeline can be viewed as suspect. Similarly, the presence of more than the occasional grain of hardwood pollen types such as *Ulmus*, *Quercus*, *Carya* or *Fraxinus* should alert us to the possibility of Tertiary contamination.

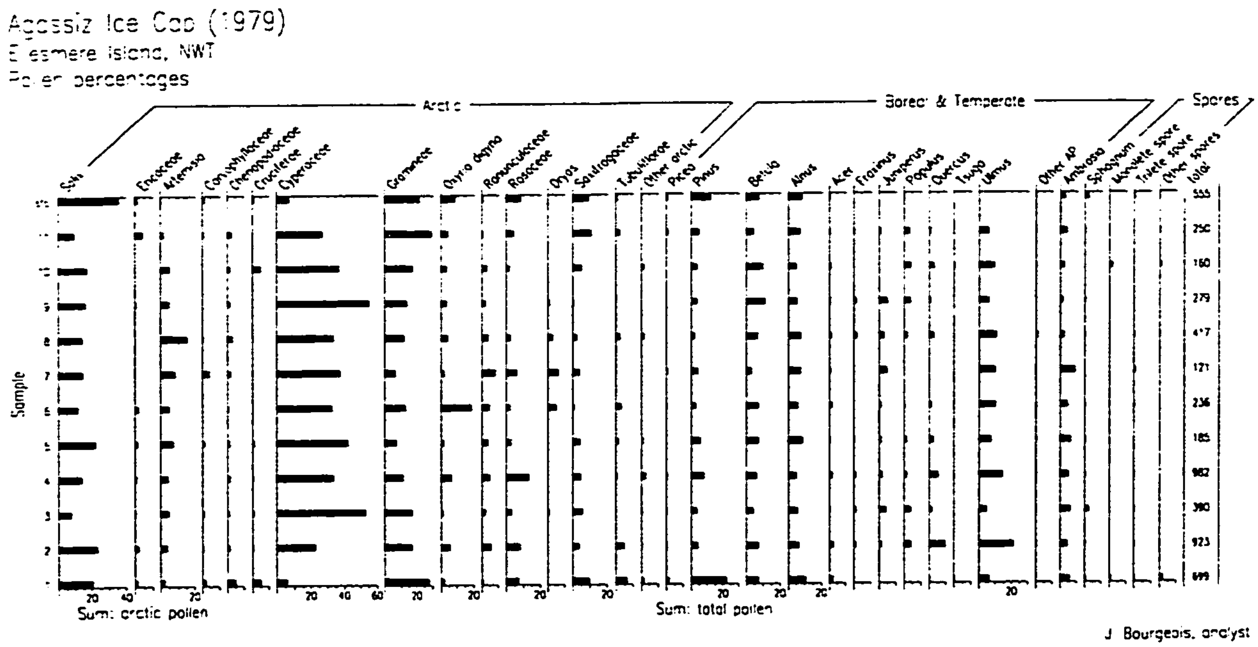

FIG. 11. Pollen percentages from a core from the Agassiz Ice Cap. Based on evidence in Koerner *et al.* (1988), such as oxygen isotopes and pollen concentration, samples 1 to 4 are most likely from the last interglacial, 5 to 8 from the last glacial and 9 to 11 are from the Holocene.

One possible approach to this problem would be to produce synoptic modern isopoll maps of the arctic. Mapping the spatial patterns of arboreal taxa may show anomalous local areas characterized by high percentages of pollen types such as *Picea* or *Pinus*, which may then be assumed to be due to local pre-Holocene formations. Another solution might be to use the *Picea* : *Pinus* ratio in the samples. Studies on ice caps (Bourgeois *et al.*, 1985; Short and Holdsworth, 1985) and in the low-arctic tundra (Elliot-Fisk *et al.*, 1982) have shown a rapid decline of *Picea* pollen north of tree-line, and therefore the relative increase in representation of *Pinus* pollen (Fig. 7). When the opposite occurs in the high arctic, it could be due to local contamination of pre-Holocene pollen rather than long-distance pollen transport.

Not every tree pollen found in regions of Tertiary deposits should be viewed as contamination. There is ample evidence for long-distance transport of tree pollen north of the tree-line. For example, tree pollen have been found in relatively large amounts in snow on the Arctic Ocean thousands of kilometers from any trees (Bourgeois *et al.*, 1985). In snow at high elevations of the Agassiz Ice Cap, large annual and seasonal fluctuations of tree pollen have been observed (Fig. 12) with no evidence of pre-Quaternary pollen or spores. Furthermore, the highest number of arboreal pollen is not deposited at the same time of year as the dust or as the regional pollen types, which suggests a different origin of these grains (Bourgeois, 1990). In most of our samples, the arboreal pollen are dominated by taxa that now grow at or near treeline. If these pollen had been released from local or Tertiary sources, a more varied spectrum would be expected. Until there will be available a more extensive array of sites, it will be difficult to evaluate the extent of this problem. It may be useful in the future to present two pollen diagrams at each analyzed site computed one with and one without tree pollen in the sum to permit the interpretation of alternate hypotheses. Finally, fluorescence microscopy has been suggested as a method to separate older from more recent pollen (van Gijzel, 1967), but we have not yet attempted this.

Dating of the Sediment

Dating of sediments has been a persistent problem in arctic paleoecology. The stratigraphic problems discussed above may cause date inversions, making interpretation of the pollen sequences difficult. Dates on organic material of aquatic origin can be too old due to the 'reservoir effect' or due to contamination of Holocene sediments by redeposited material (Olsson, 1974). All of these problems are exacerbated by the low organic matter in lake sediments.

Older carbon can be incorporated in more recent sediment in several forms; as palynomorphs, as fine fragments or as graphite (Olsson, 1974; Nambudiri *et al.*, 1980; Hyvärinen, 1970). Olsson and Blake (1962), Olsson (1968, 1974) and Namburidi *et al.* (1980) presented curves showing estimates of the age of recent samples as a function of the percentage of included old carbon or fossil palynomorphs.

In the same vein, we attempted to estimate the degree of contamination of peats by old carbon from the Fosheim Peninsula of Ellesmere Island in order to explain the chronostratigraphic inversions observed in certain dated horizons. Two samples of the surface sediment of a peat deposit were dated at the Université Laval radiocarbon laboratory. Bulk samples with no cleaning or other treatment were dated to estimate the potential contamination by ancient carbon or palynomorphs. In principle the date should have been too old, and this correction factor could have been applied downcore to give a first estimate of error induced by redeposition of ancient material. However, in both cases radiocarbon dating gave modern dates with ^{14}C activity greater than one (Sample 1: ^{14}C activity = 1.095 based on 0.7713 g C; Sample 2: ^{14}C activity = 1.208 based on 1.1810 g C; modern dates should give activities equal to one). Thus, although the ages were expected to be too old, due to contamination by pre-Holocene material, the radiocarbon dating gave ages of 'too young'. These results can be explained by an increased concentration towards the polar regions of radioactive fallout associated with atomic bomb tests in the 1950s. The level of this contamination increased and subsequently decreased from 1961 to 1972 and was greatest around 1964 (Lowden and Dyck, 1974). This source of contamination modifies the radiocarbon accumulated by various living organisms, and explain the ^{14}C activity greater than one obtained in the surface sediment. It is thus impossible to use this approach to estimate the old carbon error in this region.

Other sources of error can be introduced by redeposition of material by periglacial, fluvial and aeolian processes as discussed above. Results of dating several horizons of different deposits on the Fosheim Peninsula show a greater frequency of inversion in deposits that have been downcut by streams than in those not so affected (Table 1). One possible solution to obtain reliable dates on certain samples is to date the same stratigraphic level using both AMS and conventional methods, or AMS dates on bulk sediment and separated macrofossils or different chemical fractions of the sediment (Lowe *et al.*, 1988). With conventional dates, the age of the cleaned bulk sediment is obtained, which has a possible incorporation of pre-Holocene palynomorphs and fine carbon. With AMS, it is possible to date one or several specific pieces of material such as seeds, leaves or twig fragments. Since it is conceivable that these fragments were themselves redeposited when the sediments were being laid down, the duplicate dating thus provides a check for the detection of possible contamination.

In lake sediments, organic matter is very rare (weight loss on ignition can be less than 20%; *pers. observation*) and AMS dating is essential. Bulk-sediment dates may give dates that are too old, especially in a region of carbonate bedrock or with older sedimentary deposits in the watershed. Unfortunately, macrofossils may be rare as well.

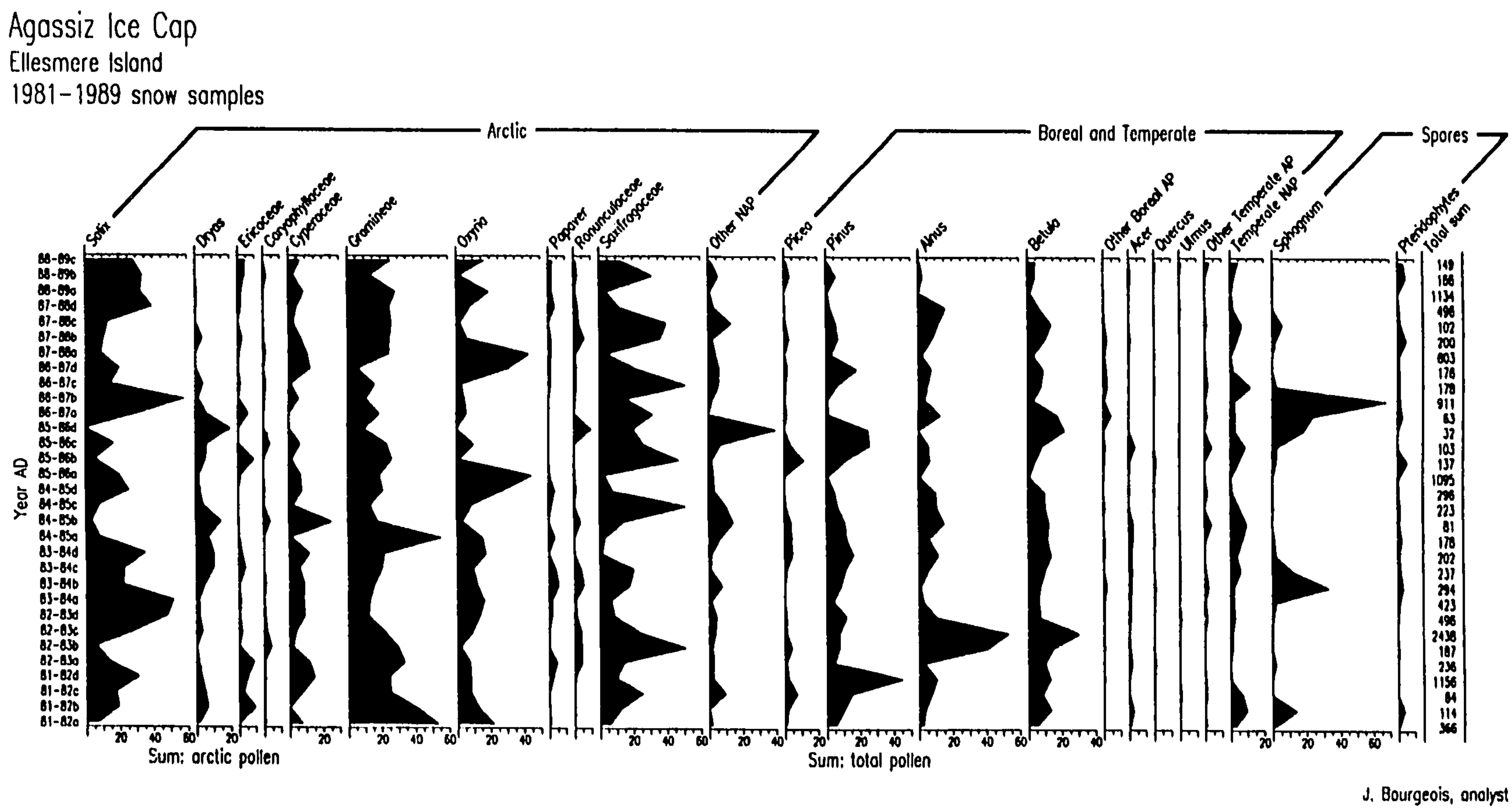

FIG. 12. Pollen assemblages from snow layers on the Agassiz Ice cap. The letters a to d refer to 'seasonal' layers where (a) represents the first snow accumulation of the year and (d) the melt layer of the following summer. Diagram modified from Bourgeois (1990).

Attempts have been made to date various fractions of lake sediments, and results have shown significant discrepancies when different materials are dated (MacDonald *et al.*, 1991). Macrofossils of terrestrial organisms are younger, and both bulk sediment and aquatic macrofossils (mosses, mollusc shells) give significantly older dates. Aquatic organisms absorb carbon from the surrounding water that may not be in equilibrium with the atmosphere, leading to dates that are too old, and bulk dates may include not only this error, but also a further one caused by the presence of redeposited old carbon (MacDonald *et al.*, 1991). Extracting and dating only chironomids is feasible at some sites, and gives younger dates than those obtained from the bulk sediment, but it is not clear if these are more reliable (Jones *et al.*, 1993). Brown *et al.* (1989) and Regnéll (1992) concentrated the pollen in a sediment core by chemical treatments and sieving and compared the AMS date to that of the bulk sediment. They found the pollen dates up to 1000 years younger than bulk dates of the same sediment, and suggested that dating of the pollen fraction of a sediment core could be used to more accurately estimate arrival or decline of species. These studies were done in the boreal forest, where the large concentrations of conifer pollen make this process feasible, if not time-consuming. In the high-arctic, the pollen concentrations are orders of magnitude less, and this solution is probably not practical.

CONCLUSIONS

There is considerable potential for late-Quaternary paleoecological studies in the arctic. None of the problems discussed above are necessarily unique to arctic environments, however, it is necessary to be aware of them. Studies of carefully selected sites have successfully established the history of the vegetation in those regions (e.g. Fredskild, 1973, 1983, 1985b; Funder, 1978; Kelly and Funder, 1974; Short *et al.*, 1985).

The analysis of micro- and macrofossils from peats can provide multiple lines of evidence of local vegetation and hydrological change. Many stratigraphic changes in peat deposits are probably due to site-specific development, but those that can be shown to have occurred at the same time in many basins may reflect the response of these sites to climate changes. Detailed studies of high arctic peat deposits are needed before meaningful generalizations can be made. Lake sediments may present fewer problems of stratigraphy but thermokarst and thaw lakes must be avoided. However, more detailed analyses of the lake sediment may be needed to confirm that they are not disturbed. Dating these sediments has proven difficult.

One of the advantages of using ice cores for pollen analysis studies is such studies are always done in combination with other studies such as $\delta^{18}O$, % melt (indicative of summer temperatures), chemistry and dust concentration. All of these are indicators of paleoclimates or past atmospheric conditions. Pollen diagrams produced from ice cores reveal broad regional changes in the composition of the vegetation cover through time (McAndrews, 1984; Bourgeois, 1986) and may serve, along with other parameters, as paleowind indicators. A pollen diagram produced from pre-Holocene ice samples has been used to date a core and establish the formation of the ice cap (Koerner *et al.*, 1988). One of the problems now is the lack of diagrams from surrounding areas for comparison and correlation. These lacustrine or peat records could then complement the paleoclimatic information from the ice cores.

We have emphasized the reconstruction of past vegetation and surface environments in this review, but past climates can potentially be estimated also from paleoecological data. The relationship of arctic plant physiology to microclimate and soil conditions has been studied in some species (e.g. Chapin *et al.*, 1992) and arctic vegetation diversity has been related to macroclimate patterns (Rannie, 1986; Edlund and Alt, 1989). The application of numerical methods to specify the relationship of pollen assemblages to climate parameters awaits the development of a network of modern pollen sites for calibration purposes and investigation of appropriate techniques (Anderson *et al.*, 1991).

We have not devoted much attention to the question of the different temporal resolution of these various types of records. Ice cores show a response to climate change at seasonal up to millennial scales (Fisher and Koerner, 1983), and there is a very good resolution in ice cores for the last 1000 years. Annual and even seasonal layers can be distinguished and the pollen extracted (Fig. 12; Bourgeois, 1990). In addition to providing a high-resolution record of the past few years, results such as these can help in the interpretation of records from peat deposits and lake cores. For example, *Sphagnum* is a component of pollen diagrams from Somerset Island (Gajewski, *submitted*), although the genus is rare in the Arctic Islands. The two *Sphagnum* peaks in snow layers of the Agassiz Ice Cap in 1983 and 1986 (Fig. 12) indicate that it is being transported in large amounts under certain situations, and a meteorological analysis may help explain the factors involved (cf. Nichols *et al.*, 1978; Barry *et al.*, 1981). Thus, ice cores offer the potential for interpreting high-resolution changes of the recent past. However, due to the low concentration of pollen, large volumes of ice must be processed, sometimes limiting the temporal resolution of the pollen record. In the arctic, the low sedimentation rate in lakes may limit the temporal resolution of the resulting pollen diagram; the entire postglacial period can be contained in less than 2 m (Blake, 1981; Gajewski, *submitted*). Frequently, the lowest sedimentation rates, and thus lowest resolution, have occurred in the past several thousand years, a time when the resolution of the ice cores is greatest. The growth and decomposition of peat and the composition of the enclosed microfossils can respond to rapid climate changes (Aaby, 1976; Garneau, 1993) and these deposits offer the potential of high resolution reconstructions as well.

Studies combining the results from peat analyses, lakes and ice cores from one region offer the potential to complement one another as a method to check for the presence of these problems. Each of these sedimentary deposits accumulates palynomorphs from a different yet overlapping source region and with potentially different resolution. A regional investigation of arctic aeolian deposits, buried peats, lake sediments and ice cores would enable a comparison of these environments to climatic change. Another approach is to integrate different paleoenvironmental proxy data (e.g. sediment geochemistry, diatoms, aquatics invertebrates) found in lakes or peat deposits (e.g. Blake *et al.*, 1992; Garneau, 1992), as is done in ice cores. Finally a synoptic approach, where microfossils or macrofossil values from a network of sites across the arctic are mapped, can identify large scale patterns of abundance of pollen and vegetation and how these have changed in the past. The first step in any paleoenvironmental analysis — the accumulation of the raw data — is only beginning, but will form the basis for an analysis of environmental change in arctic environments.

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APPENDIX B

Snow Chemistry and Pollen Deposition in the Circum-Polar High Arctic

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Introduction

The annual snow cover lying on perennial sea ice in the Arctic Ocean or on ice caps in the circumpolar area contains within it aerosols scavenged from the atmosphere. On the sea ice and some of the ice caps, these aerosols represent the natural and industrial pollutant component of the atmosphere for the period of August to the following April/May. In a region where the number of sites available for direct sampling of the atmosphere is limited, snow sampling might fill the gaps, thereby defining the aerosols pathways and sinks, more precisely. In this paper we concentrate on the ion chemistry and the pollen content of the snow cover.

Methods

In the spring of 1993 and 1994, snow samples were collected on Siberian and Canadian ice caps, on multi-year ice floes of the Arctic Ocean, including a sample site at the North Pole, and at a few land based sites on the tundra.

For the study of ion concentrations, the August-May snow pack lying on multi-year sea ice cover was sampled vertically at 50 to 100 mm intervals. A pre-cleaned plastic scoop was used to take horizontal samples from the wall of a shallow snow pit. The samples were immediately transferred to Whirlpak bags. In 1993, the samples were kept frozen until analysis but in 1994, the samples melted in the Whirlpak bags during transport between Russia and Canada, were subsequently frozen until analysis. All analyses were done at the Geological Survey of Canada, Ottawa.

The pollen samples require considerably larger volumes than the chemistry samples. The samples were collected by shoveling the complete surface to sea ice snow layer into large plastic bags. The samples were melted and filtered in a field laboratory. The laboratory method and analysis follow Bourgeois *et al.* (1985). Pollen fluxes were controlled by snow depth and density measurements at each sample site.

Results and Discussion

Chemistry

The intention of the ion chemistry part of the circumpolar work was to identify the main trajectory of anthropogenic pollutants using the SO_4^{2-} and NO_3^- ions as tracers. About 50% and 33% respectively of these ions found in snow in Northern Ellesmere, are of anthropogenic origin (Koerner *et al*, submitted) of a total maximum concentration (natural plus anthropogenic) commonly reaching 400 ppb (SO_4^{2-}) and 200 ppb (NO_3^-) each spring.

The traverse results indicate that there is an overwhelmingly strong local (marine) component. This masks the anthropogenic SO_4^{2-} input. Additionally, the ion concentrations on land-based snow sites are usually high near the snow surface in spring and decrease with depth, to either the ground or the previous spring snow surface underneath. On the Arctic Ocean ion concentrations frequently do not decrease with increasing depth. This is because inputs from local open ocean sources may dominate at any time during the August-May period. The NO_3^- concentrations are not affected in this manner as, presumably, they bear little relationship with marine inputs. If we use NO_3^- as an anthropogenic tracer (which is questionable) we see a trajectory similar to that of pollen (see below), i.e. a decreasing concentration moving from a presumed Eurasian source, to a Canadian site of deposition.

Pollen

Pollen emission and availability for long-distance transport, vary considerably throughout the year. The suite of samples studied does not cover the entire flowering season. The pollen grains in the snow layers could have been deposited: 1) in late summer of the previous year; 2) in the early spring of the sampling year; 3) during the winter months from sources at lower latitudes or through redeposition. The presence of pollen in the winter atmosphere and in snow has been reported in the Canadian Arctic (Bourgeois, 1990), and in Fennnoscandia (e.g. Janzon, 1981; Hjelmroos & Franzén, 1994) where it is often associated with dust storms.

The pollen samples in this study are dominated by tree species and in particular by pine (*Pinus*) which represents from 3% to 99% of the total pollen in the samples. The lower percentages are found in the tundra sites where local herbaceous species and birch (*Betula*) are better represented. In this study, only the tree species are considered. In order to compare results from different sites and to account for the differences in accumulation, the results were first converted to concentrations (no. pollen grains per litre of melted snow) and then to fluxes (no. pollen grains/m^2) for the eight month period.

Pollen fluxes show considerable inter annual and spatial variations. In the 1993 and 1994 sets of samples the results, with the one exception cited below, show decreasing fluxes (and concentrations) from the Eurasian side to the Canadian Arctic. The number of tree pollen in the 1993 traverse varies from approximately 15 000 pollen grains/ m^2 on the Siberian islands to fluxes as low as 100 grains/m^2 in the Canadian Arctic, for the eight month period. The fluxes start to increase again at the sites nearest to the treeline on the Canadian side. In the 1994 traverse, the fluxes are much lower than in 1993 on the Siberian side (usually less than 2 000 grains/m^2) but nevertheless show the same decreasing trend towards the Canadian Arctic.

At the 1993 Arctic Ocean II site (86°30'N-96°E), a large number of tree pollen was found in the snow. A flux of 149 291 grains/m^2, of which 99% were pine pollen, was obtained. It is possible that the multi-year ice floe originally came near the coast of Siberia and that the pollen came from a nearby source. It is also possible that the pollen were incorporated in the sea ice cover along with sediments in a coastal region of Siberia, a process described in Nurnberg *et al. (1994)*. However, in both cases one would expect a more varied pollen spectrum and particularly an increase of tundra indicator species. The results may well illustrate the phenomenon of an isolated "point" injection at a single locality. Such a phenomenon may also occur with any other components, natural or anthropogenic and needs further investigation.

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APPENDIX C

Establishing the chronology of snow and pollen deposition events on Agassiz Ice Cap (Ellesmere Island, Northwest Territories) from autostation records

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Alt, B.T. and Bourgeois, J.C., 1995: Establishing the chronology of snow and pollen deposition events on Agassiz Ice Cap (Ellesmere Island, Northwest Territories) from autostation records; in *Current Research 1995-B; Geological Survey of Canada, p. 71-79.*

Abstract: Regional pollen concentrations in seasonal snow layers of Agassiz Ice Cap for the years 1981-89 and preliminary results for 1991 are compared with the first quality controlled records from the experimental autostation established at the 1977 borehole site. Emphasis is placed on the distinctive, coarse grained (a) layer which is deposited by the first snow storm of each new mass balance year. These storms are sufficiently well developed to pick up, transport, and precipitate the tundra pollen, and to wash out pollen stored in the atmosphere. The extremes of (a) layer pollen concentrations reflect the climate conditions of the inland areas of the intermontane region to the southeast of the Agassiz Ice Cap station. Accurate dating using the autostation records allows the atmospheric circulation conditions to be examined resulting in valuable information on the source and pathways of the pollen, moisture, contaminants and paleoenvironmental indicators deposited with the snow of the (a) layer.

Résumé : Les concentrations régionales de pollen dans les couches de neige saisonnières de la calotte glaciaire Agassiz pendant les années 1981-1989 et les résultats préliminaires de 1991 font l'objet de comparaisons avec les premières données qui ont été soumises à un contrôle de la qualité et qui ont été enregistrées à la station automatique expérimentale mise en place à l'emplacement du sondage de 1977. L'accent est mis sur la couche (a) distinctive, à grain grossier, qui est déposée par la première tempête de neige de chaque nouvelle année du bilan massique. Ces tempêtes sont suffisamment fortes pour recueillir, transporter et précipiter le pollen de la toundra et pour entraîner le pollen en suspension dans l'atmosphère. Les concentrations extrêmes de pollen de la couche (a) reflètent les conditions climatiques régnant dans les zones intérieures de la région intermontagneuse, au sud-est de la station de la calotte glaciaire Agassiz. Une datation précise, basée sur les données enregistrées par la station automatique, permet d'étudier les conditions de circulation atmosphérique. Il en résulte des informations utiles sur la source et la trajectoire du pollen, l'humidité, les polluants et les indicateurs paléo-environnementaux qui se déposent avec la neige de la couche (a).

INTRODUCTION

The distribution of pollen with time has been recorded in deep ice cores (Fredskild and Wagner, 1974; McAndrews, 1984; Bourgeois, 1986; Koerner et al., 1988) and in snow pits and shallow cores on high arctic ice caps (Bourgeois et al., 1985; Short and Holdsworth, 1985; Bourgeois, 1990a, b). In order to interpret the pollen deposition results from the deep and shallow ice core work and from the seasonal snow over the Arctic Ocean, it is necessary to understand the mechanisms of transport and deposition of pollen in the High Arctic region. The effect of extreme deposition events on the long-term records and their quantitative importance relative to prevailing conditions must also be addressed. To this end, seasonal and annual pollen deposition in snow layers of the Agassiz Ice Cap (Fig. 1) were investigated by Bourgeois (1990a). These studies have been continued and expanded. In 1988 an autostation was established at the drill site on Agassiz Ice Cap (Fig. 2). The autostation records combined with seasonal pollen deposition investigations provide the first opportunity to study the specific atmospheric conditions under which pollen deposition takes place on the ice cap.

The present report makes use of these autostation records for an initial examination of the very distinctive coarse grained late summer/autumn seasonal layer. This layer, which has been labelled the (a) layer because it is the first layer following the end of summer melt, has been found in the annual snow for each of the 13 years sampled to date on Agassiz Ice Cap (Bourgeois, 1990a; Bourgeois, unpublished field reports 1986 to 1994). We have also found this coarse grained layer on Penny Ice Cap (Baffin Island), Devon Island Ice Cap, and on Academii Nauk Ice Cap (Severnaya Zemlya, Russia). On Agassiz Ice Cap, the highest regional pollen (i.e., from the high arctic tundra) deposition was observed in this layer. Bourgeois (1990a) found there appeared to be a good relationship between extremes of these regional pollen values and July mean monthly temperature from the permanent weather station at Eureka. As this (a) layer was found to be only 5 to 15 cm thick, it thus represents a relatively short time period in the seasonal record.

The detailed meteorological records will be used to address the following questions regarding the (a) layer. 1) What is responsible for the distinct nature of this layer and why is it so widespread? 2) Why does this late summer/fall layer rather than the melt layer contain the highest deposition of regional pollen? 3) To what extent does it represent the

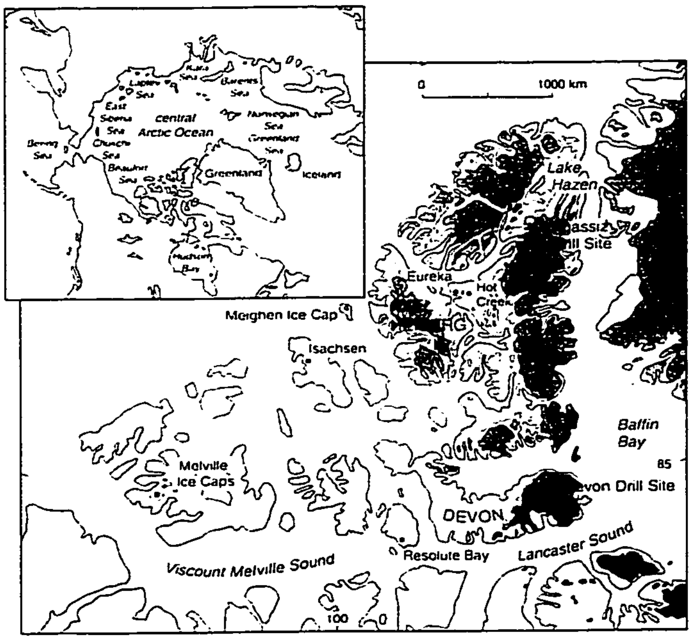

Figure 1. Map of the high arctic islands showing the Agassiz drill site and some other locations mentioned in the text. The dark shaded areas are glacier ice and the light shaded areas represents land above 610 m. a.s.l. The insert shows the circumpolar region.

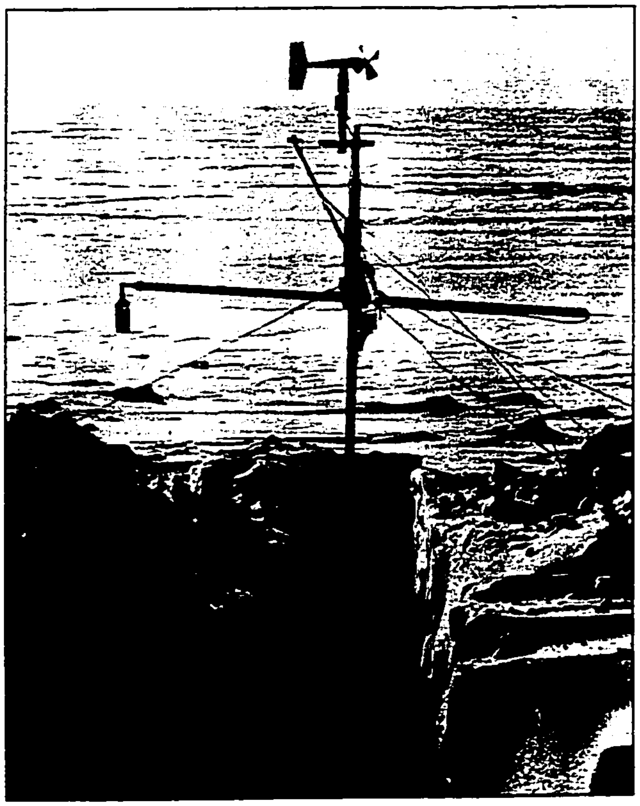

Figure 2. Agassiz Ice Cap 1977 borehole site autostation mast with two snow depth sensors (one hidden behind mast). Gill radiation shield, and wind sensor. The data logger enclosure is buried under the snow. The instrumentation has been provided and maintained in the past few years by Campbell Scientific Canada Corporation as part of a Industrial Partners Program, testing autostation technology in the severe climatic conditions experienced on high arctic ice caps. (photo: M. Waszkiewicz)

climate of the source region and how important are extreme seasons and events in determining the annual and long-term pollen deposition values? 4) How accurately can this layer be dated? Based on this dating, what weather patterns were responsible for this deposition and what were the circumpolar atmospheric circulation conditions producing these local weather patterns?

BACKGROUND AND METHODS

Geographical location

The northern part of the Agassiz Ice Cap, on Ellesmere Island, has been the site of glaciological investigations since 1974. The pollen sampling site (80.7 N - 73.1 W, 1700 m a.s.l.) is located near the location of a deep borehole, drilled in 1977 (A77), approximately 1 km from a local ice divide. The closest unglaciated terrain from the Agassiz Ice Cap sampling site is about 35 km to the north but the most likely pollen sources for the ice cap site are more than 75 km away in the intermontane region of central and northern Ellesmere Island and on Axel Heiberg Island. Some parts of this region, such as the Fosheim Peninsula, support a relatively dense and diverse vegetation cover with more than 100 species of vascular plants (Edlund and Alt, 1989).

Pollen

In 1986, samples were collected from one wall of a 2.5m deep pit located near the A77 bore hole. In each of the following springs the previous year's snow accumulation (as discussed below) was sampled in the vicinity of that pit. The results of the 1987, 1988, and 1989 sampling and some initial results from the 1992 sampling are used in the study. The samples cover the time period from late summer 1981 to late summer 1991.

In order to obtain an adequate pollen sum, from 13 to 35 kg of snow were sampled for each layer. The sampling method and laboratory procedures are discussed in Bourgeois (1990a). The results of the seasonal pollen analysis used in the present paper are expressed as pollen concentration rather than as percentages. Detailed results for the 1981-89 period are also presented in Bourgeois (1990a). Approximately 50 per cent of the pollen grains deposited on the ice cap are tree and shrub pollen from the boreal forest and low arctic region. The other half are from plants that can be found in the high arctic tundra. In the following discussion, only the total regional pollen influx values will be considered.

Stratigraphy

The annual and seasonal layers were determined from the snow stratigraphy in the snow pit (Fig. 3). The glaciological term "balance year" is used to label the annual layers. A glacier mass balance year starts when the snow begins to accumulate at the end of the melt season and ends when melting stops the following summer. The dates of these events vary from year to year and from site to site.

On high arctic ice caps, in most years, the summer melt produces a layer of hard granular snow. These summer surfaces are usually clearly visible in the snow stratigraphy. On Agassiz Ice Cap, they can be recognized on a regular basis to a depth of approximately 10 m, equivalent to about thirty years of accumulation (R.M. Koerner, pers. comm.). This is the final layer of the mass balance year and in this study we label it the (d) layer.

Immediately above this melt layer, in each season sampled to date, a well-defined coarse grained layer has been observed. As the snow in this layer is the first accumulation of the new mass balance year it is labelled the (a) layer. This layer is relatively thin (Fig. 3 and Bourgeois, 1990a). Detailed stratigraphic studies on Devon Ice Cap (Koerner, 1966) showed that the condition of this (a) layer is the result of intense metamorphism of the snow after it is deposited. The coarse grained, loosely packed layer is overlain by a thick layer of high density, fine grained snow. Unless some stratigraphic boundary was evident this winter layer was divided into two equal parts (b) and (c).

Koerner (1966) pointed out that although all snow accumulation undergoes metamorphism, the degree to which this occurs is dependent on a number of variables. The process is dependent on the transport of water vapour, which takes place from warm to cold snow and from surfaces of high curvature (ie., points) to surfaces of lower curvature. The steep reversal of temperature gradient experienced in the surface snow layer in autumn results in strong upward vapour transport and growth of large snow grains. The vapour gradient is twice as great at temperatures between 0°C and -10°C as between -10°C and -20°. Vapour transport is suppressed in the dry, fine grained winter snow which is not resistant to drifting and packing. Autumn snow which falls as wet, complex flakes is more resistant to packing therefore remains more porous (and provides many pointed surfaces), thus promoting strong vapour transport.

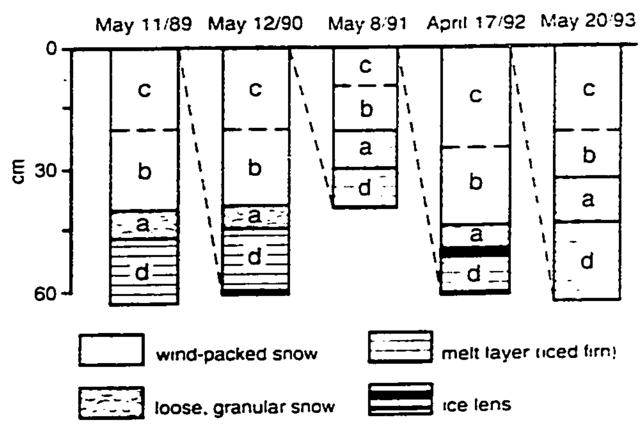

***Figure 3.** Annual snow stratigraphy and sample increments for the summer of 1988 to the spring 1993. The (a) to (d) layers represent a glacier balance year.*

Climate and weather

Long-term climate records are available from two coastal stations each 260 km distant from the ice cap sampling site. Alert, to the northeast, lies on the coast of the Arctic Ocean and is affected by the broad expanse of the polar pack and the proximity of Greenland. Eureka lies on Slidre Fiord to the southwest of the ice cap in the protected intermontane region (Edlund and Alt, 1989; Alt and Maxwell, 1990). Autostation records were begun at Hot Weather Creek in 1988. These data represent the climate of the inland areas of the intermontane region. This inland region is a possible source for pollen reaching the ice cap sampling site. Smaller inland areas and so called "arctic oases" in northwestern Ellesmere Island would experience similar climatic conditions to that recorded by the Hot Weather Creek autostation.

The record from the Agassiz 1977 borehole site (AGA(A77)) autostation (Fig. 2), established and maintained in co-operation with Campbell Scientific Canada Corporation (a GSC Industrial Partner), began in May 1988. These data provide the first opportunity to study actual climatic parameter values associated with the deposition of snow which is being sampled for pollen, atmospheric contaminants and paleo-indicators such as oxygen isotope values. Two summer seasons, 1988 and 1991, were chosen for the present analysis based on the availability of snow depth records, snow stratigraphy, and (initial) regional pollen data. The snow surface heights were measured once daily in 1988 with two CSMALL01 snow sensors (Fig. 4b) and at six-hour intervals in 1991 with two improved UDG01 sensors (Fig. 6b). Daily maximum, minimum, and mean temperatures were obtained from values sampled at one minute intervals (Fig. 4a, 5a, 6a, 7a). The daily wind speed means, available in 1991, are calculated from two minute vector means taken every six hours (Fig. 6b). Daily temperature means from the Hot Weather Creek autostation (HWC(AWS)) are the mean of the daily maximum and minimum temperatures which are obtained from values sampled at one minute intervals

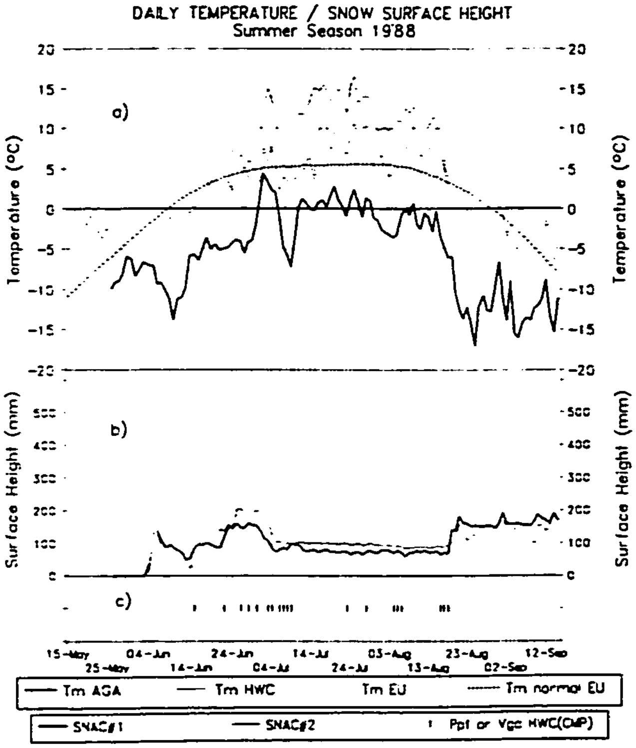

*Figure 4. a) Mean daily temperature (°C) Agassiz Ice Cap 1977 borehole site autostation AGA(A77), Hot Weather Creek HWC(AWS) autostation and Eureka (EU) weather station for summer 1988 and 30 year normal daily temperature for Eureka for 1988. **b)** Snow surface height recorded daily by two CSMALL01 sensors mounted on the same instrument mast at AGA(A77). **c)** Vertical lines represent days when some form of precipitation or virga was recorded on the aviation weather forms from the Hot Weather Creek field camp.*

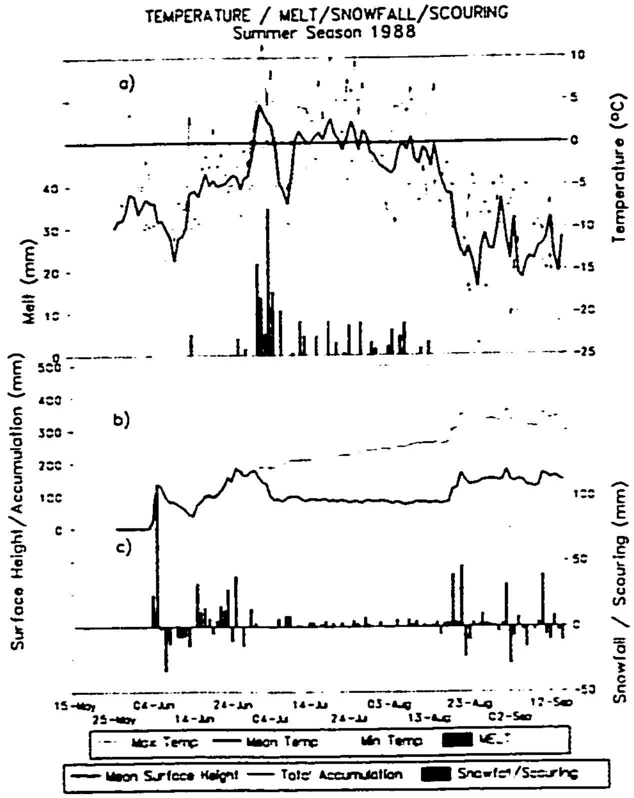

*Figure 5. **a)** Maximum, mean and minimum temperature and melt calculated from surface lowering events and expressed as millimetres of new snow for AGA(A77) autostation from 1988. **b)** Mean snow surface height from two CSMALL01 sensors and calculated total accumulation assuming melt is retained and all scoured snow is lost to the snow pack, expressed as millimetres of new snow. **c)** Snowfall (positive) and snow wind scour (negative) calculated from mean surface height expressed as millimetres of new snow.*

(Fig. 4a, 6a). Detailed site descriptions, instrumentation and quality control procedures are given in Labine et al. (1994) for AGA(A77) and HWC(AWS). The Eureka daily mean values (AES archive) are calculated from the maximum and minimum temperatures (Fig. 4). The remarks section of the Hot Weather Creek field camp aviation weather records were used to obtain the precipitation occurrence information included in Figure 4b. The synoptic weather patterns associated with the deposition events seen in the AGA(A77) autostation records were studied using microfilm records of surface and upper air charts from the AES Central Meteorological Centre in Dorval. Due to the lack of data over most of the Arctic Ocean and the sparse surface data network in the Canadian high arctic, charts often need to be reanalyzed for the regions influencing the study area. The summer atmospheric circulation patterns for the study area are discussed in detail elsewhere (Alt and Maxwell, 1990; Alt, 1987a, b; Bourgeois et al., 1985).

Snow accumulation

The sensors measure the height of the snow surface relative to the position of the sensor. The mean value of the two snow surface height measurements (Fig. 5b, 7b) have been translated into snowfall, melt, and snow scour. The daily snowfall values (Fig. 5c, 7c) are calculated assuming that any rise in the surface height is the result of deposition of snow. Each surface lowering event for the summers of 1988 and 1991 was examined along with the temperature, wind speed and other weather records. It was found that if the mean temperature was above -4°C or the maximum temperature was above freezing (0°C) the event was likely to be a melt event. The exceptions were isolated surface lowering events occurring late in the season when solar radiation is low. In these cases the maximum air temperature must reach 1°C, regardless of the mean temperature, for melt to take place. Surface lowering due to melt was usually gradual while surface lowering due to wind scour was abrupt and normally followed a new

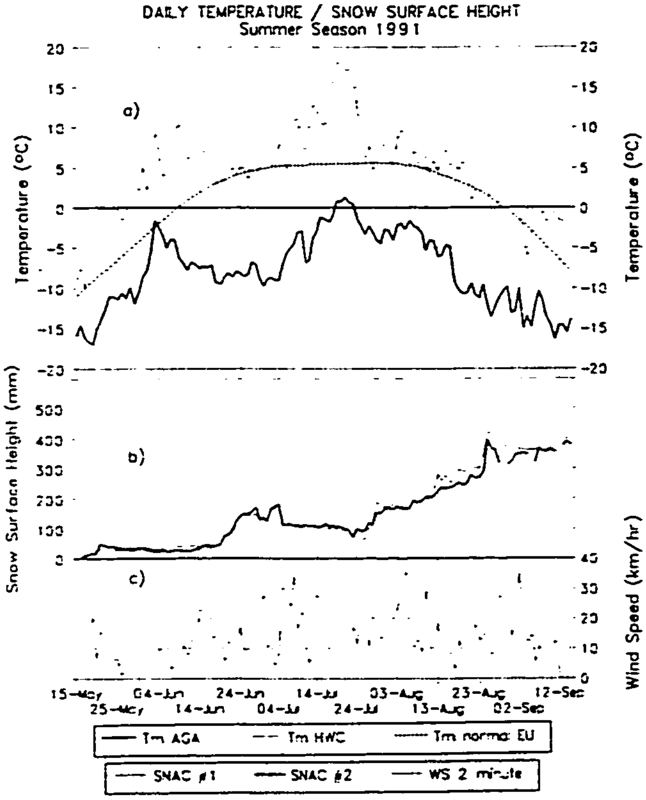

*Figure 6. **a)** Mean daily temperature (°C) Agassiz Ice Cap 1977 borehole site automatic weather station (AGA(A77), Hot Weather Creek HWC(AWS) autostation for summer 1991 and 30 year normal daily temperature for Eureka. **b)** Snow surface height recorded daily by two UDG01 sensors mounted on the same instrument mast at AGA(A77). **c)** Mean daily wind speed from six-hour 2 minute vector means for AGA(A77) autostation.*

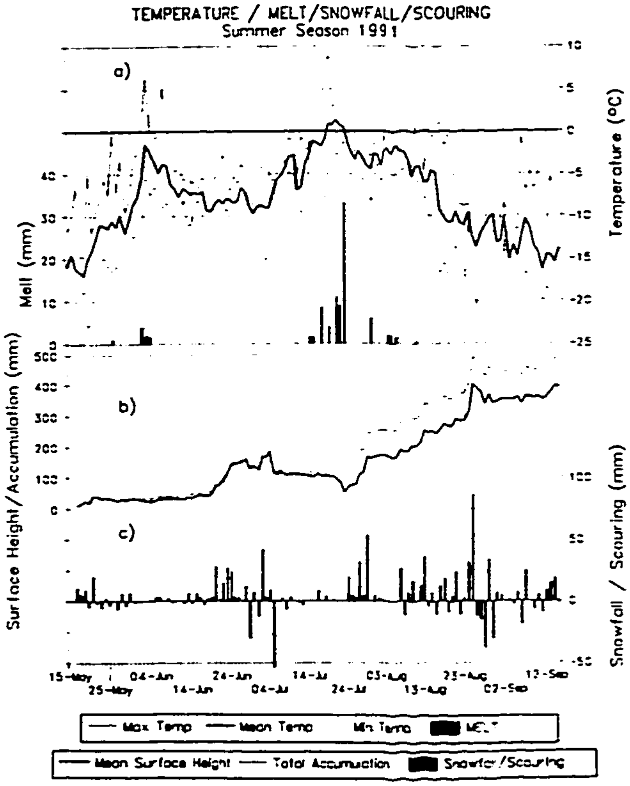

*Figure 7. **a)** Maximum, mean and minimum temperature and melt calculated from surface lowering events and expressed as millimetres of new snow for AGA(A77) autostation for 1991. **b)** Mean snow surface height from two UDG01 sensors and calculated total accumulation assuming melt is retained and all scoured snow is lost to the snow pack, expressed as millimetres of new snow. **c)** Snowfall (positive) and snow wind scour (negative) calculated from mean surface height expressed as millimetres of new snow.*

snowfall. Based on these assumptions daily melt (Fig. 5a, 7a) and snow scour (Fig. 5c, 7c) values have been calculated for the two summers. The total daily accumulation in the snow pack is calculated assuming that all melt is retained by the snow pack and all scouring lost (Fig. 5b, 7b).

RESULTS

Relationship of pollen in (d) and (a) layers to July temperature on the tundra

The relationship noted by Bourgeois (1990a) between maxima and minima of regional pollen influx values for the (a) layer and maxima and minima in the mean July temperatures at Eureka can be seen in Figure 8. The very high concentration for the (a) layer of 1988 is better reflected in the mean temperature record for Hot Weather Creek as this station is truly representative of conditions on the inland tundra areas. To date there is only a brief overlap between the Hot Weather Creek record and the pollen record. The preliminary results from 1991 show far less pollen than would be expected based on the temperatures at Eureka and Hot Weather Creek.

This may reflect poor pollen production due to the deleterious effect of the short early warm period which was followed be three weeks of well below normal conditions in late June-early July. This may have depressed pollen production despite the warm July mean temperatures.

Chronology of the (d) layer in 1988

The 100 mm of gradual surface lowering, in the seven days following 1 July, represents melting of all the snow that fell on the ice cap during the period 20-30 June. Melt ended abruptly on 8 July when the maximum temperature dropped to -4.4°C during a brief cold spell.

Between 10 July and 13 August a net surface lowering of 20 mm took place which would have melted the snow deposited between 16-19 June. In addition, during July and early August, 53 mm of new snow fell and melted. The (d) layer as seen in the pollen pit was 160 mm thick. Assuming a density for this melt layer snow of 0.37 g/cm^3 and for new-fallen snow of 0.15 g/cm^3 (Koerner, pers. comm.), the (d) layer contains the refrozen melt water of 120 mm of spring snow plus 53 mm of summer snow and another 220 mm of spring snow deposited before 16 June into which the melt water percolated. Thus of the 400 mm of (d) layer deposition only 53 mm (or 14%) fell during the pollen production season in the tundra.

The last significant melt probably occurred on 13 August although the surface lowering of 6 mm on 16 August may have been due to melting. The melt season came to a definite close with the deposition of 40 mm of new snow on 19 August.

Chronology of the (a) layer in 1988

The (a) layer in the pollen sampling pit was only 70 mm thick. Given a density for new-fallen snow of 0.15 g/cm^3 (Koerner, pers. comm.) and for the (a) layer of 0.23 g/cm^3 (Koerner, 1966), the (a) layer required 107 mm of new snow accumulation. Assuming that melt ceased and the (a) layer deposition began on 14 August this amount of accumulation was reached by 31 August but much of it was scoured the following day. The required accumulation was definitely received by 19 September. Well over half of this snow, however, was deposited between 14 and 22 August.

The first 51 mm of (a) layer snow was deposited directly on top of the melting snow surface over a six day period 14 to 20 August. Mean air temperatures at this time are in the -5°C range. These conditions would produce a very steep negative 5°C temperature gradient in the first 51 mm of new snow. This same precipitation fell as rain at Hot Weather

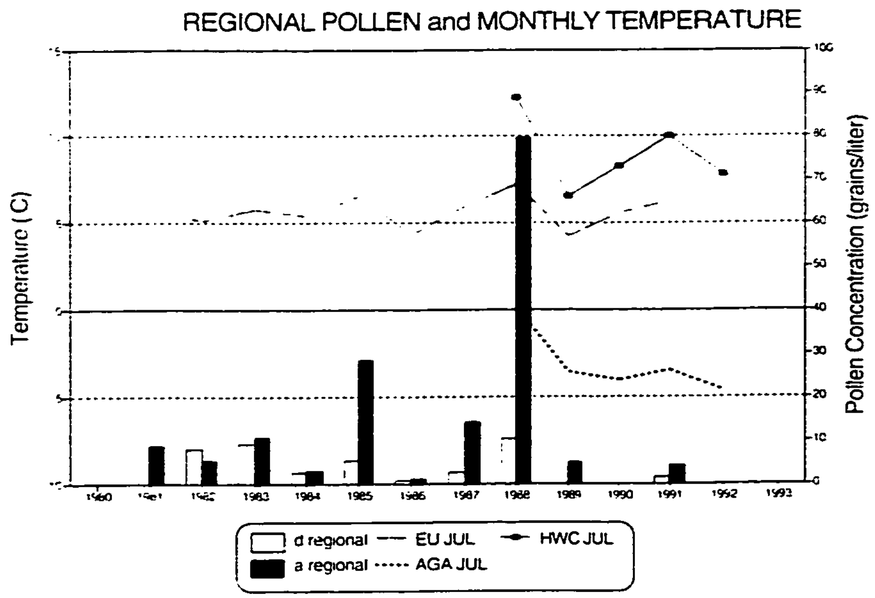

Figure 8.

Pollen concentrations in the (d) melt layer and (a) metamorphosed layer with mean July daily temperatures from Agassiz Ice Cap (AGA), Hot Weather Creek (HWC), and Eureka (EU).

Creek. There is no evidence of scouring of this snow. All these factors suggest that this 51 mm of new snow meet all the criteria suggested by Koerner (1966) for very rapid metamorphism. The synoptic charts for this period show passage of two surface lows over the northern Ellesmere area. Both these systems travel along the strongly baroclinic zone north of Alaska. At this time of year a band of open water bordered by warm land to the south and cold sea ice to the north provides the necessary temperature gradient and moisture source to provide support for typical mid-latitude low pressure systems with well defined frontal systems.

The first of these systems dissipates as it reaches the high arctic islands and moves south around the blocking high pressure system which dominated the preceding month. Deposition associated with this system is in the 2 mm/day range similar to that experienced during the melt season when weak troughs passed over the area depositing a dusting of snow. The second of these systems (Fig. 9) tracks from northern Siberia into the Bering Sea reaching the Beaufort Sea by 15 August. Here it is regenerated and moves northwest along the edge of the high arctic islands with strong upper air support from the 50 kPa circumpolar vortex which moves into the Beaufort Sea sector of the central Arctic Ocean. The surface frontal system occludes and the trowal (trough of warm air aloft) is over Eureka early on the morning of 19 August and is responsible for the 40 mm of snow deposited on Agassiz Ice Cap later that day. This upper trough contains relatively warm air and considerable moisture. The moisture falls in the form of wet snow which melts before reaching the ground over the tundra but is deposited as snow on the ice cap. Upward vertical motion associated with the trough would provide a mechanism for entrainment of the pollen from the tundra regions it passes over.

On reaching the eastern side of Ellesmere Island the surface feature appears to split, the southern portion moving into Baffin Bay while the northern portion moves north to sit under the upper polar vortex in the central Arctic Ocean north of the islands. Subsequent systems move into the islands from the Arctic Ocean This is reflected by the rapid drop in temperatures at the ice cap station between the 19 and 22 August.

The snow deposited on 22 August and scoured (or packed) the following two days would be colder and drier than that which fell on 19 August. However a strong temperature gradient would still exist in the upper layer of snow as air temperatures are in the -10°C range while the top of the melt surface would not have cooled below -5°C in such a short time. Thus metamorphism would still take place although probably at a reduced rate. Air temperatures remain in the -10°C range for over a month. During this time the warm melt surface layer now buried under some 70 mm of new snow would be cooled from above and below and the negative temperature gradient in the snow which is conducive to rapid metamorphism would slowly disappear. Snow deposited in late September would not undergo rapid change and thus would appear as winter snow in the stratigraphy the next spring.

Chronology of the (d) layer in 1991

At the very top of the melt layer (d) of summer 1991 the stratigraphy shows a thick ice layer. The only melt event strong enough to create such an ice layer was that which ended abruptly with the seasons maximum melt of 33 mm on 23 July. The 110 mm depth of the (d) layer represents 270 mm of new fallen snow (using density of 0.37 g/cm^3 for the melt layer and 0.15 g/cm^3 for the new fallen snow). As most of the new snow deposited in late June - early July was scoured, the 70 mm of melt occurring between 15 and 23 July was snow deposited in mid June well before pollen production would have begun on the tundra. This and the early June melt percolated 190 mm into the snow pack deposited prior to mid June.

Deposition occurred on only one day during the melt period when a low moved into northern Ellesmere Island from the central Arctic Ocean. This weak Arctic Ocean low would be responsible for the only precipitation deposition of regional pollen to the ice cap during the main melt period of 1991.

Chronology of the (a) layer in 1991

When the first snow of the (a) layer was deposited, the melt water produced the previous day was in the process of refreezing at the surface. The melt was responsible for lowering the

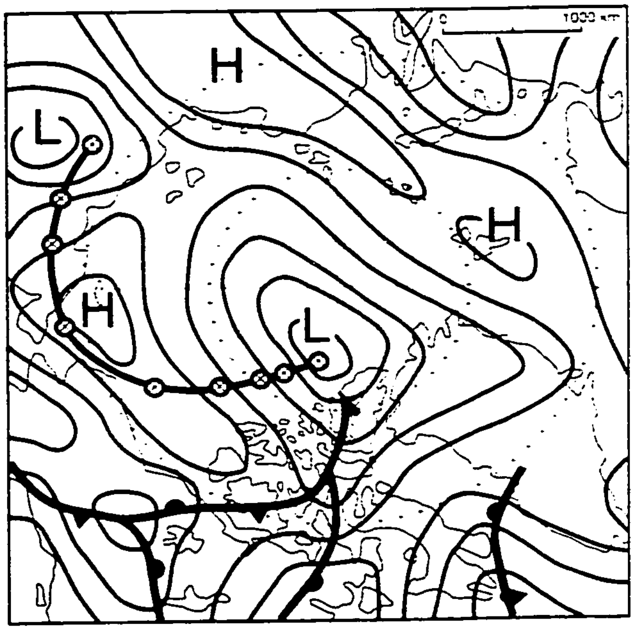

Figure 9. *Surface atmospheric pressure contours (in millibars) for 12h00 UCT on 19 August 1988 showing passage of trowal (trough of warm air aloft) over Eureka. Track of low pressure system is shown by joined circles denoting the centre position for consecutive 24 hour intervals.*

surface by 33 mm. The snow fell with air temperatures in the -2°C to -5°C range and winds between 10 and 20 km/h. Most of the 63 mm of new snow, required to make up the 50 mm of observed (a) layer, fell during the 4 days following the maximum melt event of the 1991 summer. A total accumulation of 116 mm was reached by the 6th day after the melt event. No scouring was observed in the daily surface lowering record during this time although there is possible evidence in the six-hour record of some scouring or packing on 28 July. A three to four degree temperature gradient would have existed in this relatively new snow to support rapid metamorphism of this layer. There is the possibility of several very minor melting events after the deposition of this layer but the mean air temperature remains below -2°C. These minor melt events, in fact, may have contributed to the formation of large snow grains of the (a) layer.

The first 58 mm of new snow accompanied the passage of a well developed surface low pressure system. This system moved into the islands from north of the Beaufort Sea having stalled there for several days (Fig. 10). Prior to stalling north of the Beaufort Sea this system can be traced back to the East Siberian Sea but the lack of data in the ocean makes its exact history uncertain. Satellite imagery might provide more information. The system appears to have had a frontal system associated with it but the surface fronts passed south of northern Ellesmere Island. There is evidence, in the 85 kPa chart (which is the approximate altitude of the ice cap sampling site), of the upper warm trough reaching as far north as Alert. Winds at the ice cap station become southerly on 26 July during the major period of deposition.

By contrast the 53 mm deposition on 29 July was accompanied by northerly winds at the ice cap station. These are consistent with the northerly flow on the surface and upper air charts at this time. It is not clear what produced this high accumulation value. It may have been orographic lift of the moist Arctic Ocean air mass or there may be a small trough or low pressure system which, due to the lack of data over the central Arctic Ocean, cannot be located . One further possibility is that much of the surface change on that day was due to drifting of snow to the autostation site from elsewhere on the ice cap. In this latter case the (a) layer would consist entirely of snow from the low pressure system discussed above.

Air temperatures at the ice cap station remained in the -2°C to -5°C range for another 20 days after the deposition of the (a) layer. Temperatures on the tundra were still above freezing and the ground was not yet snow covered. Snow deposited during this period would not be significantly colder than the snow surface which would now be in equilibrium with the ice cap air temperature. This isothermal gradient in the top layer of snow would not promote rapid metamorphism in the new snow. The (b) layer accumulation for this year (1991-92) can be taken as having begun with the deposition event of 7 August.

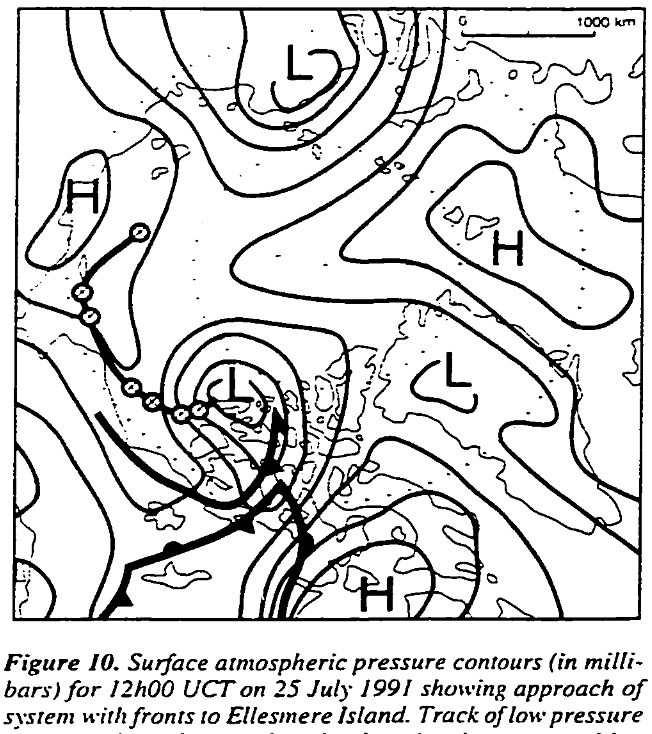

Figure 10. Surface atmospheric pressure contours (in millibars) for 12h00 UCT on 25 July 1991 showing approach of system with fronts to Ellesmere Island. Track of low pressure system is shown by joined circles denoting the centre position for consecutive 24 hour intervals.

CONCLUSIONS

The results of this initial study of autostation records and the (a) layer in the pollen seasonal record provide the following answers to the questions posed in the introduction.

1. The loosely packed, coarse grained (a) layer is a result of the deposition in a short period of time of a significant amount of snow on to the melting or recently melting summer snow surface. Only the first significant snowfall deposited over the melting snow surface develops a sufficient temperature gradient to undergo rapid metamorphism. This accounts for the relatively thin depth of this layer. The widespread distribution of this distinctive layer suggests that on arctic ice caps the melt season is usually culminated by an accumulation event.
2. Only a small portion of the (d) layer deposition occurs during or after the pollen production season on the tundra. The first storm of the accumulation season, which is responsible for the (a) layer deposition, has two sources of pollen to draw on. The first is the pollen that has accumulated in the atmosphere during the pollen production (melt) season but which has not been washed out. Some dry deposition doubtless takes place and some instances of light precipitation occur but these are often local events as shown by comparison of the Agassiz and Hot Weather Creek records in 1988 (Fig. 4b, c). The second is the pollen deposited during the pollen season on the dry, snow free ground of the tundra. In both summers studied, the first storm of the accumulation season was a

well developed system with good upper atmosphere support. Both passed over the tundra areas to south and west of the ice cap. These systems had the necessary upward vertical motion to pick the pollen up from the ground and carry it to the altitude of the ice cap. They also contained sufficient moisture to cause significant precipitation when the upper trowal passed over the mountains.

3. Pollen deposition in the (a) layer on the ice cap appears to better represent the continental climate of the inland tundra regions than that of either of the coastally located permanent weather stations. There appears to be an almost exponential increase of pollen reaching the ice cap with increasing temperature. Extreme situations are better represented than are the more normal seasons. It is particularly interesting to note that proximity of the date of deposition of the (a) layer to the date(s) of pollen production does not seem to influence pollen concentration in the layer. The extremely pollen rich (a) layer of 1988 was deposited almost a month later than the pollen poor (a) layer of 1991. The pollen production in the source region may itself not always accurately represent the climate of the source region as suggested by the preliminary 1991 results. The overlapping record, presently available, is too short to make a quantitative assessment of the various relationships or of the impact of extreme events on the paleo record.

4. Using the autostation records, it was definitely possible to date the two (a) layers studied. In both years, the well developed surface systems, responsible for most of the snow in the layer, originated in the Siberian sector and tracked around the central Arctic Ocean, entering the high arctic islands from the west or southwest. The 1991 system followed a more northerly track than the 1988 system which travelled along southern boundary of the permanent sea ice. In both cases the regional pollen source(s) was probably west or southwest of the ice cap. The exotic pollen source for the layers might differ somewhat between the two years, as might the moisture and contaminant source. The preliminary results indicate that all would, however, have come from Siberia and Alaska rather than eastern North America.

These results show the importance of continuing the simultaneous collection of seasonal pollen and autostation records from Agassiz Ice Cap and the Hot Weather Creek tundra site. Detailed analysis of the (a) layer of each year in the overlapping record is necessary to verify the results of this preliminary study. Study of the other seasonal layers and of other constituents of the snow pack are also warranted.

ACKNOWLEDGMENTS

We would like to thank all those who took part in the autostation and pollen sampling work in the field on Agassiz Ice Cap and at Hot Weather Creek. In addition to members of Terrain Sciences Division, this includes personnel from Campbell Scientific Canada Corporation and Atmospheric Environment Services, Arctic Division. Polar Continental Shelf Project has provided flying time and logistic support to the Agassiz Ice Cap and Hot Weather Creek programs since their inception. R.M. Koerner provided valuable discussion regarding the snow stratigraphy. Ana Djuric helped in the pollen analysis and diagram preparation and Tracy Barry prepared the maps.

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APPENDIX D

PRESENT AND PAST ENVIRONMENTS INFERRED FROM AGASSIZ ICE CAP ICE CORE RECORDS

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ABSTRACT

Six surface-to-bedrock ice cores have been recovered from the Agassiz Ice Cap, providing long paleoenvironmental records. The cores have been analyzed primarily for oxygen isotope ratios ($\delta^{18}O$), percentage of melt layers, volcanic acidity layers, and pollen concentrations. The pre-Holocene section is limited to a few metres above the bed, but nevertheless contains ice from the last glacial period and, at the very bottom, from the last interglacial. The absence of older ice suggests that the present ice cap started to grow during the last interglaciation. Melt concentration and $\delta^{18}O$ records indicate that summer temperatures were 2 to 2.5°C warmer in the early Holocene than at present. A gradual cooling has occurred in summer since approximately 8000 years ago with a minimum around 100-200 years ago.

Short cores and snow-pit samples retrieved in the borehole area have been used primarily for chemical and pollen analyses. The chemistry records reveal an increase in the concentration of contaminants in the snow since the middle of this century that is associated with a marked increase in SO_2 emissions in Eurasia. Long-range transport of tree and herbaceous pollen are also evident from pollen studies of the snow layers of the Agassiz Ice Cap. The pollen assemblages show seasonal and annual variations and are compared with meteorological records obtained from automatic weather stations. These meteorological records are considered essential for transfer functions for both the ice-core results and the long-term mass-balance survey.

INTRODUCTION

The northern part of the Agassiz Ice Cap (Fig. 1), on Ellesmere Island, has been the site of glaciological investigations since 1974. Six surface-to-bedrock ice cores have been drilled, providing long records of past environmental changes. The cores are referred to as A77, A79, A84, A87, A93.1, and A93.2, the number representing the year they were drilled. In addition, several shorter cores have been extracted near the deep ice cores. The area around the borehole sites has been monitored extensively over the years and snow-pit samples have been analyzed for contaminants (both natural and anthropogenic). Mass balance measurements, which determine the difference between the amount of accumulation (usually snow) and the amount of ablation (melting of ice and snow), have been taken every spring since 1977 near the borehole sites and on a nearby glacier.

Because of the substantial amount of information available on the Agassiz Ice Cap, the site has been used to test new instruments or methods. In 1988, an experimental automatic weather station (autostation) was installed near the A77 borehole site and two more stations have since been placed at other locations. Many problems inherent to remote arctic locations remain to be solved, but the available data are being used to relate the various parameters being studied on the ice cap to climate variables.

This paper summarizes some of the modern and paleoenvironmental records obtained thus far from the Agassiz Ice Cap ice cores and snow surveys. Some of these records may be regarded as climate proxies whereas others reveal changes in the composition of the atmosphere over the ice cap through time. The paleoenvironmental records discussed are primarily (1) stable oxygen isotope ratios, presented as $\delta^{18}O^{(1)}$, in which the variations in the records are essentially interpreted as changes in temperature; (2) ice-melt stratigraphy (percentage of melt layers in core sections), which provides a record of past summer conditions at the site and through relation to mass balance survey results over the ice cap in general; (3) the acidity record, determined by solid electrical conductivity measurements, that provides seasonal acidity variations in which high-acidity peaks are associated with major volcanic eruptions; and (4) pollen spectra, composed of both regional tundra types and extraregional types (mainly tree pollen), which are used to trace the possible origin of the air masses reaching the ice cap and can provide additional climate information when combined with other core parameters.

This review of the Agassiz Ice Cap snow and ice core records is divided into three sections; the pre-Holocene, the Holocene and the modern record. The pre-Holocene refers to the section between the late Pleistocene-Holocene transition and the bedrock. It represents less than 10 per cent of the total core length, but probably close to 90 per cent of the time period covered by the ice cores. The resolution is low in the early Holocene, but improves in younger sections. The most recent part of the Holocene (the modern record) is presented separately. With the exception of the meteorological and mass-balance records, the data were obtained primarily for monitoring levels of contaminants and the samples were recovered from snow pits and shallow cores.

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(1) The ratio of the concentrations of heavy and light isotopes in a sample is measured in terms of its deviation from the ratio of standard mean ocean water(SMOW), which is expressed as δ and measured in parts per thousand (‰).

THE BOREHOLES AND ICE CAP

The Agassiz Ice Cap, approximately 16 000 km^2, covers central Ellesmere Island between latitude 79°45'N and 81°N. It is drained by glaciers that penetrate the surrounding mountains, the larger ones reaching sea-level. The boreholes are in the northern part of the ice cap, at latitude 80°40'N, longitude 73°30'W. The first core (A77), drilled in 1977 to a depth of 338 m, is about 1 km downslope from a local ice divide at an elevation of 1670 m (Fig. 2). Core A79 (139 m deep and located about 30 m higher) is 200 m down the opposite side of the ice divide; cores A84 and A87 (both 127 m deep) are 30 m apart at the top of the flow line, which is a local ice dome. In 1993, two more cores (A93.1, A93.2) were drilled, only a few metres apart, at the top of that local dome. However, with one exception (Zheng *et al.*, 1998), the results obtained from cores A93.1 and A93.2 are unpublished and since these two cores are only tens of metres away from the A87 borehole site, they are not discussed in this paper. Information on the various core parameters is presented in Table 1.

The Agassiz Ice Cap lies in the Eastern Canadian Arctic within the Arctic Mountains and Fiords climate region as defined by Alt and Maxwell (1999). It faces the anomalously warm, dry intermontane region to the west (Alt and Maxwell, 1999) and the contrastingly wet regime of northern Baffin Bay to the southeast as reflected by Koerner's regional accumulation map (Koerner, 1979) reproduced in Alt and Maxwell (1999). It is somewhat protected from cold, wet Arctic Ocean air masses by the mountains of northern Ellesmere Island. The automatic weather station (autostation), installed near the site of the A77 borehole in 1988, is in the area where these three climate regions would be expected to converge. The station records temperature, changes in relative snow surface height, wind speed and direction, relative humidity, and incoming solar radiation. Preliminary reports on the data and on instrumentation problems associated with remote Arctic locations are presented by Alt *et al.* (1992) and Labine *et al.* (1994). At the A77 site, during the six years of record (1988-93), monthly mean screen-level temperatures for July varied from +0.5°C to -4.9°C and for January from -32.4°C to -39.3°C (Table 2). Mean annual temperatures for this period lie in the -22.5 to -21.5°C range. The highest screen-level temperature recorded was 12.7°C, but this was undoubtedly produced by overheating of the screen. The lowest recorded temperature of -51.2°C may be a result of radiative cooling of the sensor. Above freezing temperatures were experienced every summer in the record. The resulting positive degree day (PDD) totals ranged from 47.3°C to 0.1°C (Table 2).

In the northern part of the ice cap, annual precipitation is about 0.175 m/a, ice equivalent (Fisher *et al.*, 1983). Most precipitation falls as snow in early winter. However, due to the presence of a ridge in the borehole area, wind turbulence is common and causes the removal or scouring of the soft winter snow (Fisher *et al.*, 1983; Fisher & Koerner, 1988). Borehole sites A79, A84, and A87 are currently most affected by this process, but not the A77 site. As a result, accumulation at the three scoured sites is about 65 per cent of that at the A77 site. South of the A77 site, accumulation rates vary much less (Fisher *et al.*, 1983). The removal of some winter snow also affects the $\delta^{18}O$ record since some of the cold winter peaks (most negative δ values) are truncated. The cores from the top of the ridge (A79, A84, A87) give on average 2.5‰ less negative δ values than the core drilled 1 km down the ridge (A77) (Fisher *et al.*, 1983, 1995). These differences must be considered in the interpretation of the Holocene records. At the A77 site, ice migration down the flow line presents another problem because some of the oldest ice comes from the scouring zone.

Melting is minimal at high elevations on the ice cap and when it does occur, it is recorded as ice layers in the firn. It currently represents approximately 3 per cent of the annual precipitation at the A77 site and 5 per cent at the top of the ridge. The difference is due to the thinner annual layer at the top that results from removal of some of the winter snow on the ridge. The thickness of the summer melt layers in the ice cores provides a good indicator of past summer temperatures and is used by Koerner & Fisher (1990) to produce a Holocene summer temperature record for the Agassiz Ice Cap.

TIME SCALES

Several techniques are used to establish ice-core time scales. These and the related problems are documented in a series of papers edited by Oeschger & Langway (1989). Some are based on ice-flow modelling, which provides an approximate age-depth chronology, whereas others are based on annual and seasonal variations of various components found in snow and ice, the most frequently used of which are variations in isotopic composition, dust, acidity, and trace substances. However, the lower the accumulation, the more likely the annual and seasonal snow layers will be disturbed by surface winds and the more rapidly the annual layers will thin out to the point where they cannot be identified. Calibration between cores can also be made using well known volcanic events. These are detected in ice cores as layers of high acidity.

On the Agassiz Ice Cap, the precipitation rate is relatively low; nevertheless, the time scales are felt to be accurate to within 10 per cent for the past 7000 years (Fisher *et al.*, 1983). Note that the chronology used in this paper is in calendar years (cal. yr.) and not in ^{14}C years. The initial theoretical time scales were checked by comparison with the volcanic acid layers in the cores (Fisher *et al.*, 1983, 1995). These volcanic events are well dated in the finer resolution Greenland cores. A major shift in $\delta^{18}O$ values, which occurs at a few metres above bedrock on the Agassiz Ice Cap, is correlated with a similar shift in the Greenland cores. Assuming that the more negative the $\delta^{18}O$ values, the colder the period, then this shift marks the transition between the late Pleistocene and the Holocene. In the recently drilled ice cores from the Summit area of Greenland, this transition occurs at 1400 metres above bedrock and it is assigned an age of 11 550 cal. yr. (Johnsen *et al.*, 1992). The Agassiz Ice Cap ice cores are not dated unequivocally beyond this transition.

THE PRE-HOLOCENE RECORD

In all Agassiz Ice Cap ice cores, the pre-Holocene record is limited to the bottom 8 to 11 m of ice near bedrock. Studies of ice properties (Koerner *et al*, 1981; Fisher & Koerner, 1986; Fisher, 1987) and of the origin of the basal ice (Gemmell *et al.*, 1986; Koerner *et al.*, 1987, 1988; Koerner, 1989) have been conducted. The latter studies are of importance for establishing the timing of the initial growth of the present ice cap. The much compressed record does not allow for high-resolution studies, but because of the shallow ice depth, the pre-Holocene ice is easily accessible for bulk sampling (Bourgeois, 1995; Koerner *et al.*, 1988, 1991).

In the pre-Holocene $\delta^{18}O$ records shown in Figure 3, a section with very negative δ values (cold climate) is found below the late Pleistocene-Holocene transition and another relatively warm section is found near bedrock. Core A79 and to a lesser extent the other cores at the top of the flow line have a layer of very "dirty" ice near the bed. Radioecho sounding (Walford & Harper, 1981) shows that this dirty ice layer is characteristic of the basal ice over most of the bedrock hill underlying the ice divide and extends for over 1 km northeast from hole A79. At the A79 hole, core recovery was poor in that particular section (only about 10 per cent) and most of the ice was brought to the surface as meltwater produced by the thermal drill. The origin of this particle-rich bottom layer is problematic and two hypotheses are proposed for its formation.

Koerner *et al.* (1987) consider this dirty ice to be of Ice Age origin. The warm δ values (normally indicative of a warm climate) are attributed to basal melting in a deep bedrock valley south of the borehole site at a time when the ice cap was much thicker than at present, such as at an early stage of the Sangamon Interglaciation. The warmer δ values could result from fractionation during refreezing as the ice flowed up the bedrock slope towards the borehole area. The ice could have incorporated the dirt at that time, as suggested by Gemmell *et al.* (1986).

On the basis of results of pollen analyses of the pre-Holocene ice (summarized in Fig. 4), Koerner *et al.* (1988) had to reconsider this hypothesis. They analyzed several large ice samples from early Holocene and pre-Holocene sections in core A79 core to resolve the stratigraphy in the older, undated section of the core. To obtain an adequate pollen sum, they melted, laterally, 30 to 40 kg ice samples from the bottom of the hole and pumped the water to the surface. However, they could not obtain samples from the bottom 3 m and instead used the meltwater samples recovered with the thermal drill in 1979. It has since been established that this type of sample (from the tank of the thermal drill) produces abnormally low pollen concentrations (Koerner *et al.*, 1988). Even so, the pollen concentration from the meltwater is as high as the modern values. A second set of pollen samples, retrieved from the A87 borehole using the same procedure (Bourgeois, unpub. data, 1991), shows higher concentration than today in the bottom ice samples. The results obtained from pollen analyses, together with the oxygen-isotope signatures, are for Koerner *et al.* (1988) evidence that the ice near bedrock was formed in a relatively warm period, probably towards the end of the last interglaciation. Consequently, the ice cap must have melted completely or extensively in an earlier stage of the same interglaciation. From similar ice properties found at the bottom of Camp Century and Dye 3 ice cores in Greenland, Koerner (1989) concludes that these sites must also have been ice free during part of the last interglaciation.

Other oxygen isotope variations that occur in the pre-Holocene section of the cores have not been studied in detail, but a comparative study of $\delta^{18}O$ values for the bottom part of all Agassiz Ice Cap cores is in progress. The characteristics of the pre-Holocene Agassiz Ice Cap ice have been studied from cores and from borehole measurements (Fisher & Koerner, 1986; Fisher, 1987). The studies reveal several differences between Wisconsinan and Holocene-Interglacial ice. Ice Age ice contains more microparticles, is much "softer" than interglacial ice, and deforms more readily in bore-hole tilt and closure measurements. Furthermore, because of the alkaline nature of the Ice Age ice, the solid conductivity, which is used as an indicator of volcanic acid layers in cores, is extremely low in this type of ice, compared to Holocene ice.

Other than the relatively high pollen concentration found in the lowermost ice samples of cores A79 and A87, there are no significant pollen variations in the pre-Holocene section (Koerner *et al.*, 1988). Furthermore, the difference between the glacial period and the Holocene (based on the results of pollen analysis) is not well defined. To what extent this lack of variations is due to a different (i.e. stronger) atmospheric circulation during the Ice Age and/or lower snow accumulation and therefore higher pollen concentration is still to be determined. Results obtained from the chemical and particle analyses of Agassiz Ice Cap pre-Holocene ice should help elucidate this problem.

THE HOLOCENE RECORD

Holocene records of oxygen isotopes, melt percentages, electrical conductivity, and pollen concentration for the Agassiz Ice Cap are reported by Fisher *et al.*, (1983, 1995), Bourgeois (1986), Fisher & Koerner (1988), Koerner & Fisher (1990), and Zheng *et al.* (1998). Fisher & Koerner (1994) present a detailed record of the last 1000 years of melt, $\delta^{18}O$ values and electrical conductivity; Taylor (1991) compares, for the same period, $\delta^{18}O$ (temperature) profiles from the Agassiz Ice Cap ice cores with deep ground temperature profiles from three wells southwest of the Agassiz Ice Cap.

All Agassiz Ice Cap ice cores, including the two A93 cores, were analyzed for $\delta^{18}O$ (at the University of Copenhagen). The record, usually presented in 50 year averages, is interpreted as a proxy temperature record, where a change in the isotope ratio of 0.62‰ is equal to a 1 °C temperature change (Dansgaard *et al.*, 1971). However, in core A77, the winter snow contributing the most negative $\delta^{18}O$ values was apparently removed by scouring until about 8000 years ago, by which time the borehole site had moved from the scouring zone into the non-scouring zone lower down. Consequently, the early Holocene part of the record is mainly a summer-dominated record of snow accumulation whereas the most recent part is representative of all seasons; therefore, the average δ values in the latter are 'colder'. The differences between ice cores and the distinction between the climatic signal and noise attributed, for example, to snow drifting are considered by Fisher & Koerner (1994) and Fisher *et al.* (1995) in the interpretation of these records.

Following the large abrupt change in stable-isotope ratios marking the end of the last glaciation, Agassiz Ice Cap ice core $\delta^{18}O$ profiles (Fig. 5a) show a rapid warming trend that reaches a maximum between approximately 9000 and 8000 years ago. This is followed by a gradual cooling that steepens after 1500 cal. yr. The coldest values of the entire Holocene were reached approximately 100-200 years ago. This was followed by a pronounced warming into the present century. Koerner & Fisher (1990) consider that since the early Holocene, the change in the $\delta^{18}O$ values corresponds to a cooling of the average temperature of approximately 2.5 °C. Changes in the elevation of the ice cap are not considered in this estimate as Koerner (1979) shows that in the Canadian Arctic, $\delta^{18}O$ values are unaffected by elevation.

The changing concentration of melt layers with depth is a good indicator of past summer conditions (Koerner & Fisher, 1990). Melt layers, recognized in the cores by their bubble-free texture, are formed by melting on the snow-pack surface, which produces water that percolates down and refreezes in the colder snow layers below. The warmer the summers, the greater the number of melt layers produced. The melt percentage is established by measuring the thickness of melt layers in a core and expressing it in terms of a varying percentage of the annual layer ice-equivalent thickness. Although a warmer climate would most likely lead to an increase in the amount of precipitation on the ice cap, it would probably also lead to an increase in the amount of melt.

The Holocene melt record is presented in Figure 5b. The resolution is poor in the oldest section, but nevertheless four features are significant in this record: very high melt from approximately 9500 to 8500 cal. yr.; decreasing melt from 8500 cal. yr. to about 2500 cal. yr.; very low melt from 2500 to 100 cal. yr.; and a moderate increase during the last 100 years. Presently, melting occurs almost every summer at the Agassiz Ice Cap drill sites. The meltwater percolates down into the snow, refreezes, and forms an ice layer. This represents only about 5 per cent of the annual accumulation (Koerner & Fisher, 1990). By comparison, in the early Holocene, melt layers often formed 100 per cent of the annual layer, which means that the entire annual layer melted during the summer. In the absence of runoff, this would be equivalent to the formation of a 10 cm ice layer in the snow pack each year. If there was runoff, some years would have had a negative mass balance at the drill site. Presently in the Queen Elizabeth Islands, 10 cm melt (the equivalent of early Holocene melt at the top) occurs each year at an elevation 200 m lower than the borehole site (Koerner, 1979). This can be interpreted as a minimum summer cooling of 2°C from the early Holocene to the present (Koerner & Fisher, 1990) and is comparable to the drop of 2.5°C indicated by the oxygen-isotope record.

A comparison of Holocene $\delta^{18}O$, melt, and electrical conductivity records for four of the Agassiz Ice Cap cores (Fisher *et al.*, 1995; Zheng *et al.*, 1998) reveals significant differences in δ records in the early part of the Holocene. For the same period, the volcanic acid (electrical conductivity) record is noticeably flat (Fig. 5c), whereas the amount of summer melt is high. According to Fisher *et al.* (1995), the very high summer melt and noise in the δ values between c. 8000 and 11 550 cal. yr. could indicate that at that time or possibly earlier, the site had a negative balance and, therefore, some very early Holocene ice might be missing. However, the presence in all four cores of the Younger Dryas oscillation suggests that the site was not in the ablation zone long enough for the Wisconsinan ice to be eroded. In a comparative study of circumpolar Arctic ice cap $\delta^{18}O$ and melt records, Koerner and Fisher (1995) conclude that glacier retreat was widespread in the early Holocene and some smaller ice caps (e.g. Meighen and Melville) disappeared at that time. These findings are consistent with several other proxy records from the High Arctic showing maximum warmth at the beginning of the Holocene (e.g. Bradley, 1990; Dyke *et al.*, 1996). However, they are more difficult to reconcile with some glacial geology records from Ellesmere Island which show limited glacier retreat in the early Holocene (e.g. England, 1983, 1990, 1996). According to Koerner and Fisher (1995), the cooling that followed in the second half of the Holocene promoted the expansion and regrowth of glaciers and ice caps.

Pollen analysis was carried out on the meltwater tank samples produced during drilling of core A77. For every 1.5 m increment drilled, the heating element of the thermal drill melted between 12 to 15 L of water. Field and laboratory procedures were developed to minimize the loss of pollen grains (Bourgeois, 1986). Even so, the pollen content of the melt tank samples was extremely low (usually <1 grain/L). A pollen concentration diagram was constructed for the Holocene by combining samples to produce 500 year increments. The pollen grains found in the samples were assigned to two groups on the basis of their most probable origin; regional (from the High Arctic tundra) and exotic (long distance), pollen from the latter group consisting predominantly of tree and shrub (alder, birch, pine, spruce) pollen from the boreal forest and Low Arctic tundra.

The resulting pollen concentration diagram (Bourgeois, 1986) shows that the maximum concentration of regional pollen types lags behind the temperature maximum (as recorded in the melt and $\delta^{18}O$ records) by several thousand years. A small increase occurs approximately 6000 cal. yr., but the maximum is reached 3000 years later. After 3000 years ago, the number of tundra taxa decreases slightly. The concentration of pollen from tree taxa is also very low in the early Holocene. By 7500 years ago, the concentration of alder and birch pollen starts to increase. The two taxa reached a maximum 1500 years ago, for which there is no adequate explanation. Pine and spruce pollen, although regularly present in small amounts throughout the Holocene, does not follow the same trend as pollen from the above taxa.

The A77 pollen diagram was the first produced from ice cores on the Agassiz Ice Cap. Subsequent analyses of surface samples (Bourgeois *et al.*, 1985; Bourgeois, 1990) and pre-Holocene bulk samples obtained by melting down the borehole and pumping the water to the surface (Koerner *et al.*, 1988) consistently showed higher pollen concentrations than those obtained from the melt-tank of the thermal drill. Pollen concentration is always extremely low on the Agassiz Ice Cap, but the difference between the melt tank and other types of samples is significant (<1 grain/L in melt-tank samples compared to average values of 2 to 20 grains/L for other types of samples). We do not know the cause for this, but the difference also applies to the chemical analyses of samples obtained from melt tanks and from core segments. Because of these significant variations between the pollen content of melt-tank samples and other types of snow or ice samples from the same ice cap, it is impossible to validate the pollen record obtained from the A77 site. Furthermore, other pollen diagrams produced from analysis of the A87 ice core and selected sections of the A84 core reveal high pollen concentrations in the early Holocene, with profiles similar to $\delta^{18}O$ and melt records for the same periods (J.C. Bourgeois, unpub. data, 1996).

MODERN RECORDS

A large number of surface (and near surface) samples have been collected on the Agassiz Ice Cap to 1) monitor levels of contaminants in the snow, including seasonal and spatial variations, 2) monitor climate variability (using proxy data), and 3) provide modern values for various components being studied in the ice cores. Studies related to the chemical content of snow, based on accumulation over a few years, are provided by Duchesneau (1992), Nriagu *et al.* (1994), and Goto-Azuma *et al.* (1997). Koerner & Fisher (1982), Barrie *et al.* (1985), Gregor *et al.* (1995, 1996), Peters *et al.* (1995), and Koerner *et al.* (1999) contribute longer modern records. Peters *et al.* (1997) investigated the annual deposition of ^{210}Pb to the ice cap for the 1963 to 1993 period to determine the temporal trend and to test the accuracy of ^{210}Pb dating methodology. A detailed, 35 year record of radioactivity related to nuclear bomb tests has also been developed from the firn layers of the Agassiz Ice Cap (A. Kudo, unpub. data, 1996). In addition, the biological content (pollen and bacteria) of the snow is the subject of studies by Bourgeois (1990) and Handfield (1992). The data on temperature and surface elevation changes during the summer months are used by Alt & Bourgeois (1995) to determine the timing of precipitation events and pollen deposition on the ice cap. This information is also being used for similar studies on stable isotopes and snow chemistry. Some data obtained from the Agassiz Ice Cap are used in spatial studies of High Arctic glacier mass balance (Koerner & Lundgaard,1995), pollen influx (Bourgeois *et al.*, 1985), and contaminants (Koerner, 1994). The findings, particularly those related to ice core studies, are reviewed below.

Chemical analysis

Koerner and Fisher (1982) were the first to notice increasing acid levels in Arctic ice cores based on analyses of Agassiz Ice Cap ice samples. Their measurements of pH and conductivity on core A77 and shallow cores reveal a significant increase in acidity and conductivity beginning in the early 1950's, compared to the preceding 5000 years. Barrie *et al* (1985) reviewed the same data, dating back to 1912, along with a 10 m core from Mount Oxford area in northwest Ellesmere Island. They correlated the concentrations with those obtained from aerosol samples collected at Alert on the northern coast of Ellesmere Island. Results show that since 1956, there has been a 75 per cent increase in Arctic air pollution that can be associated with a marked increase in SO_2 and NO_x emissions in Europe. More recently, Koerner *et al.* (1999) measured the influx of those two major pollutant ions in two shallow ice cores retrieved from the Agassiz Ice Cap. The cores, one recovered from a snow scoured zone and the other from an unscoured zone, cover close to 100 year of snow deposition. The record, which ends in 1989, shows no evidence of decreasing concentration of these aerosols (Fig. 6), despite decreasing emissions due to improved industrial fuel emission technology in western Europe and North America in the last two decades. According to Koerner *et al.* (1999) this suggests that the dominant pollutant source to northern Ellesmere Island is Russia. Indeed, a spatial study on snow chemistry conducted in the Canadian Arctic (including the Agassiz Ice Cap), the Russian Arctic, and on the Arctic Ocean sea ice effectively shows a decreasing concentration of pollutants from the Russian side to the Canadian Arctic (Koerner, 1994).

In 1993, an 8.3 m pit was excavated at a site approximately 1.7 km north of the Agassiz Ice Cap drill sites to study the deposition trends of polycyclic aromatic hydrocarbons (PAHs), elemental carbon (also referred to as black carbon, graphitic carbon, charcoal, or soot), and polychlorinated biphenyls (PCBs). The thirty year record shows that the flux of PAHs to the Agassiz Ice Cap has remained relatively constant for the period 1972-73 to 1992-93, with a mean value of 11 (±6) $\mu g/m^2/year$ (Gregor *et al.*, 1995; Peters *et al.*, 1995). A greater flux (mean value of 74 (±20)) was obtained for the period prior to 1972. This change in flux magnitude may reflect changes in the global use of fossil fuels (Peters *et al.*, 1995). Elemental-carbon flux shows no trend during the same period and there seems to be no relationship between elemental carbon and PAH concentrations or deposition. However, as mentioned by Peters *et al.* (1995), problems were encountered during the analysis of elemental carbon and these results should be regarded with caution.

Similarly, the flux of PCBs to the Agassiz Ice Cap shows no well defined trends and records obtained thus far disagree. In the study by Gregor *et al.* (1995), fluxes show large interannual variability during the 30 year period. Deposition was generally higher in the 1960s (maximum of 930 ng/m^2 /year in 1967-68), decreased in 1968-69, increased to 1979-80 and reached a minimum in 1980-81 (91 ng/m^2 /year). The flux has since increased, although interannual variations are still considerable. In 1987, a 17 year record was obtained at another Agassiz Ice Cap location. Gregor *et al.* (1996) conclude that at the site, PCB deposition has been decreasing since the early 1970s. However, the results obtained thus far demonstrate the need for more research on the analytical methods and post-deposition processes affecting certain types of organic contaminants in the Arctic.

In the first chemical analysis of ice samples from the Agassiz Ice Cap, Koerner and Fisher (1982) noticed a seasonal cycle in which higher acidity levels are reached in late winter-early spring. This seasonality and the spatial variations in snow chemistry (Cl^-, SO_4^{-2}, NO_3) on the Agassiz Ice Cap has been studied in detail by Goto-Azuma *et al.* (1997) and by Koerner *et al.* (1999) from snow-scoured zones and from the unscoured zone. Seasonal ion peaks are clearly evident in both zones even though a large part of the winter snow is removed from the scoured zones. This fact suggests that on the ice cap, dry deposition of pollutants, as opposed to wet deposition (i.e. in snowfall), is the primary form of deposition in mid-winter. Furthermore, the SO_4^{-2} and NO_3 peaks in late winter/early spring are enhanced by anthropogenic inputs.

Pollen

Bourgeois (1990) studied annual and seasonal variations in pollen grains in the snow layers of the Agassiz Ice Cap to gain a better understanding of the relationship between pollen concentration in snow/ice and climate. An eight-year pollen concentration record (Fig. 7) was obtained from pits excavated near the A77 borehole site. Tree pollen shows less variation than regional pollen and exceptionally high (low) concentrations are apparently related to particular synoptic conditions at the time of flowering. For example, the high concentration of tree pollen in the 1982 and 1983 spring and summer snow layers occurred when synoptic conditions promoted a strong, direct flow of air from the south to the Arctic Islands. Conversely, a strong northwesterly flow over the islands in the spring of 1986 resulted in very low tree pollen concentrations in snow deposited at that time. Regional pollen types show larger seasonal and interannual variations. A higher concentration is normally found in a loose, granular snow layer formed on top of the summer melt layer (referred to as the 'a' layer for representing the first snow of the balance year). Bourgeois (1990) considered this layer to be formed in late-summer or early autumn. However, Alt & Bourgeois (1995) studied the timing of deposition of that particular snow layer on the basis of meteorological records from the nearby autostation and found that it could be deposited as early as mid-July. At that time, melting may have ceased on top of the ice cap, but plants are still flowering on the tundra. If the first snowstorms of the season are sufficiently well developed (as in the summer of 1988 and 1991), they could potentially transport relatively large amount of tundra pollen to the ice cap as well as scavenge pollen in the atmosphere above the ice cap. According to Alt & Bourgeois (1995), extreme regional pollen concentrations reflect the conditions of the intermontane region southeast of the ice cap. However, longer records are needed to confirm this relationship.

Mass balance

Since 1977, mass-balance measurements have been made each year on the northwest part of the Agassiz Ice Cap. The measurements form two data sets, but they cannot be used to compile a real mass-balance record as the area for each elevation interval remains unknown. The first data set is from a line of 18 poles set between the drill sites and a ridge that forms the eastern part of the ice cap in this general area. It represents accumulation in the uppermost regions of the ice cap, which is in the percolation zone (where summer melt refreezes within the current surface layer). Pole measurements on their own introduce errors because the snow between the surface and the bottom of the pole settles whereas the pole cannot. Therefore, in the accumulation zone, a board is placed on the surface each spring. The depth and density of the snow overlying this board is then measured the following spring. The second set begins just above the firn line, above the 'Drambuie Glacier', and runs down that glacier to the east-northeast and into the ablation zone. In the ablation and superimposed ice zones, the distance from the top of the stake to the surface of the ice is measured each spring to determine the ice gain or loss since the previous spring.

The record from the Agassiz Ice Cap is only 20 years long. No trends are appearing, which is consistent with what has been found for four other ice caps in the High Arctic (Koerner and Lundgaard, 1995).

Synoptic Climate

Preliminary results from the A77 site autostation show that, in summer, the ice cap is approximately 12°C cooler than the interior of the Fosheim Peninsula to the west (cf. Table 2 and Alt *et al.*, in press). The mean July atmospheric lapse rates between the two stations range from 0.6°C to 0.9°C/100 m, which shows the cooling effect of the ice cap relative to the "standard atmospheric" lapse rate of 0.6°C. In winter, the lapse rate is reversed (-0.2 to -0.5°C/100 m), as would be expected from the persistence of temperature inversions in the radiosonde records for High Arctic stations during this season (Bradley *et al.*, 1992, 1993; Alt *et al.*, in press). An examination of the daily temperature records for Hot Weather Creek (Alt *et al.*, in press) and site A77 on the Agassiz Ice Cap (B.T. Alt, unpub. data, 1998) shows that the inversion breaks down during warm intrusions, which are frequent some winters and not others (see Alt et al., in press, for further discussion). The potential breakdown of the winter inversion conditions has obvious ramifications for the interpretation of the paleotemperature record relative to sea-level tundra locations. An absence of temperature inversions in paleorecords would suggest more direct transport of pollen and pollutants to the ice cap with the mid-latitude systems.

The ice cap snow depth records for Agassiz Ice Cap core site A77 (Fig. 8), as well as Hot Weather Creek and Eureka on the Fosheim Peninsula (Fig. 1), show considerable variation in accumulation rates and patterns from year to year. The snow depth records, combined with the analysis of synoptic weather conditions, provide vital information regarding source, timing, and method of deposition of the various components measured in detailed snow-pit studies (Alt & Bourgeois, 1995), such as chemical composition (Goto-

Azuma *et al.*, 1997). The autostation records also provide the data necessary to calculate transfer functions for summer melt values (Mass balance above). Comparative studies of the modern record are essential to interprete the deep ice core results.

SUMMARY

Several ice cores and sets of snow-pit samples have been collected on the Agassiz Ice Cap for environmental/climatological studies. The records obtained thus far suggest that 1) the borehole area was free of ice for part of the last interglaciation and the present ice cap probably started to grow towards the end of that interglaciation; 2) Pleistocene ice is limited to a few metres near the bed, but this ice shows the major $\delta^{18}O$ steps seen in higher resolution cores; 3) the early Holocene was apparently the warmest part of the Holocene; 4) since 8000 years ago, summer temperatures have cooled by 2 to 2.5 °C; 5) the coldest period of the Holocene occurred approximately 100 to 200 years ago; and 6) the concentration of pollutants in snow on the Agassiz Ice Cap has been increasing since the middle of this century.

The drilling of four ice cores within a few tens of metres of one another and of two other cores approximately 1 km away has allowed us to study the effect of wind scouring and summer melt on ice-core paleoclimatic records. It has also allowed us to experiment and compare results of pollen analysis obtained from meltwater, core segments, and downhole melting devices. Spatial and temporal chemical surveys from shallow cores and snow pits have provided valuable information on the reproducibility of data and on the process of dry deposition. Automatic weather stations installed at several locations on the ice cap are providing meteorological data sets that are being used for mass balance, snow chemistry, pollen analysis, and stable isotope studies.

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TABLE 1 Properties of Agassiz Ice Cap ice cores.

Core number	Elevation (m a.s.l.)	Core length (m)	Ice accumulation (m/a)	Melt average (%)
A77	1670	338	0.175	3.1
A79	1700	139	0.115	-
A84	1730	127	0.098	5.1
A87	1730	127	0.098	5.1
A93.1 and A93.2	1730	123	0.098	5.1

TABLE 2 Preliminary values of positive degree day (PDD) totals for summers in the A77 autostation record and mean monthly July and January screen-level temperatures calculated from 24 hourly values (Tm24)* and by the traditional method of maximum+minimum/2 (Tmxn).

YEAR	PDD (°C)		July mean (°C)		January mean (°C)	
	Tm24	Tmxn	Tm24	Tmxn	Tm24	Tmxn
1988	31.7	47.3	0.1	0.5		
1989	11.0	11.5	-3.6	-3.5	-39.3	-39.1
1990	8.2	16.1	-4.0	-4.5	-32.4	-32.3
1991	4.1	3.3	-3.5	-3.8	-38.3	-37.9
1992	0.1	1.9	-4.6	-4.9	-35.8	-35.6
1993					-35.7	-35.7
1994	6.3	9.1	-3.2	-3.2		
1995					-35.1	-35.0

* Differences between the Tm24 and traditional Tmxn methods of calculating daily mean temperatures are discussed in detail in Alt and Maxwell (in press). Note that for ice cap conditions, the difference in total PDD between the two methods can be quite significant (see for instance 1990).

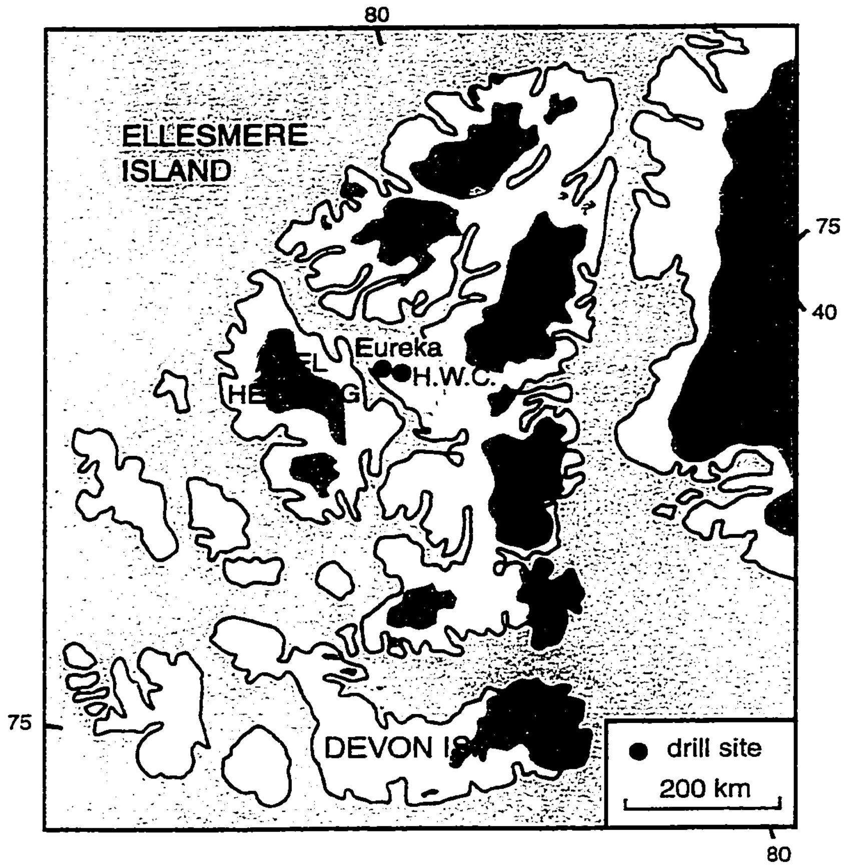

Figure 1. Drill sites on the Agassiz Ice Cap. H.W.C. = Hot Weather Creek.

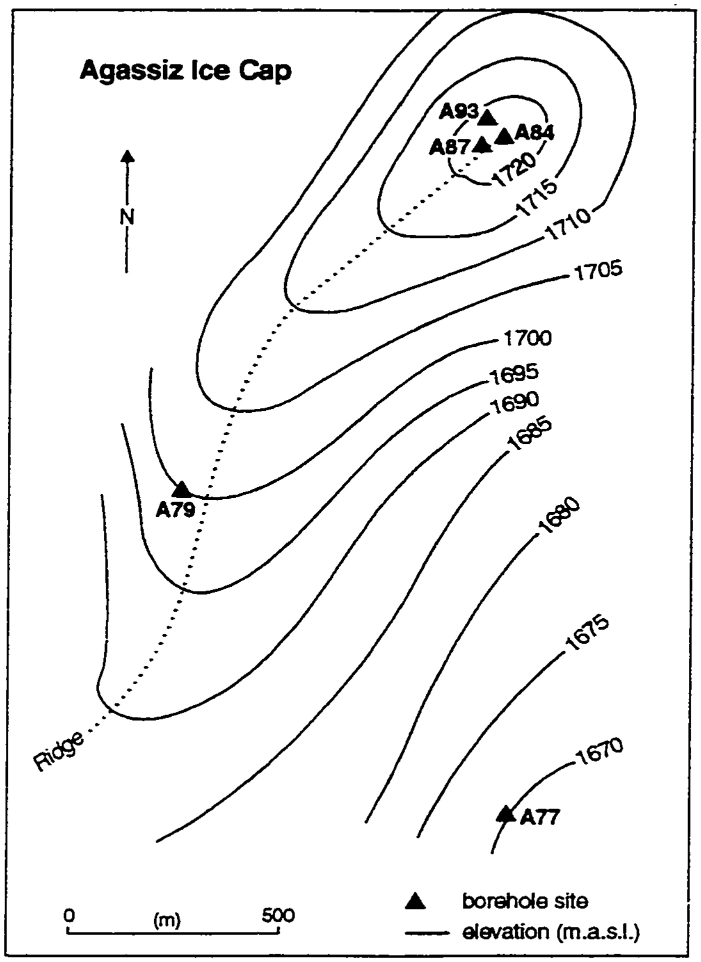

Figure 2. Borehole sites on the Agassiz Ice Cap. The site number indicates the year of drilling.

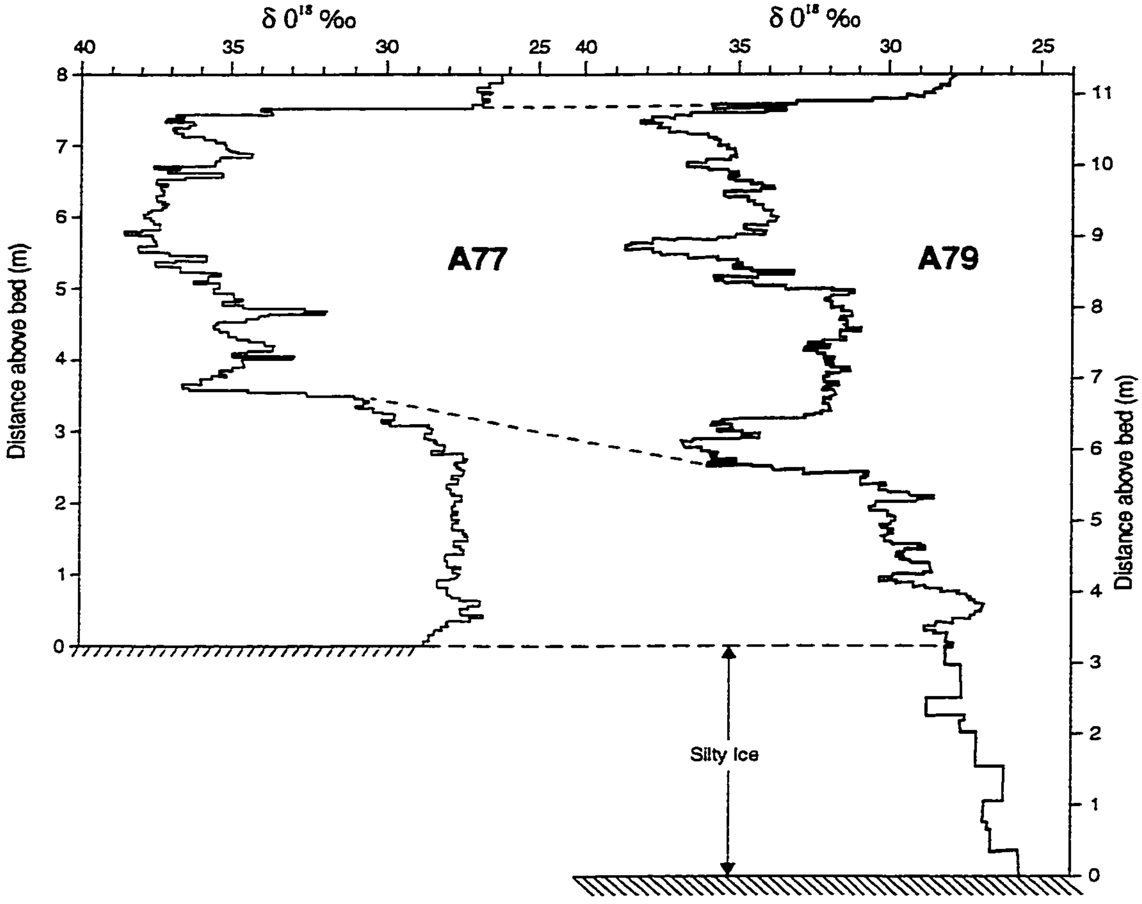

Figure 3. Oxygen isotope ratio ($\delta^{18}O$) for the lowermost section of cores A77 and A79. The late Pleistocene- Holocene transition is at 7.5 m (A77) and 10.8 m (A79) above the bed.

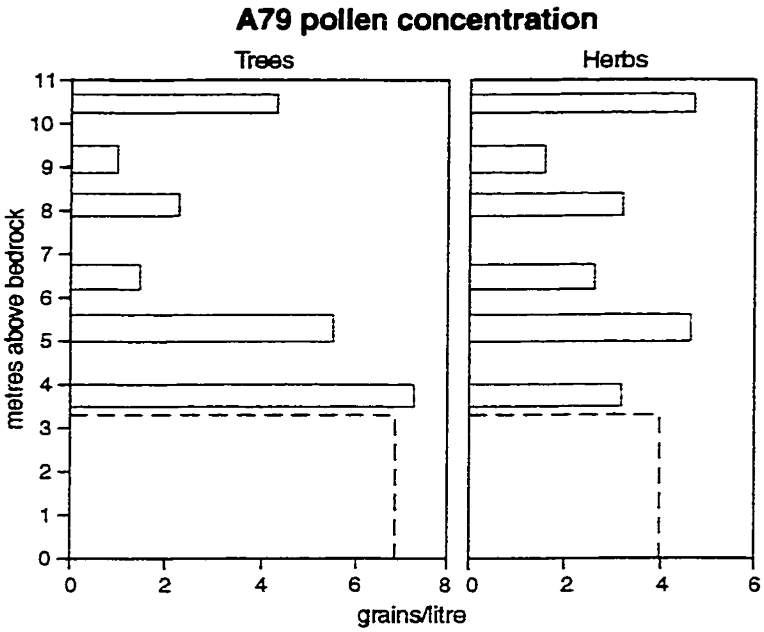

Figure 4. Summary pollen concentration diagram for the pre-Holocene section of core A79. Below the late Pleistocene-Holocene transition (top of diagram), the samples are not dated and only their approximate location above the bed is presented. The thickness of the bars indicates the thickness of the samples. The bottom sample (dotted line) is from the meltwater tank of the thermal drill and the pollen concentration is probably too low (see text).

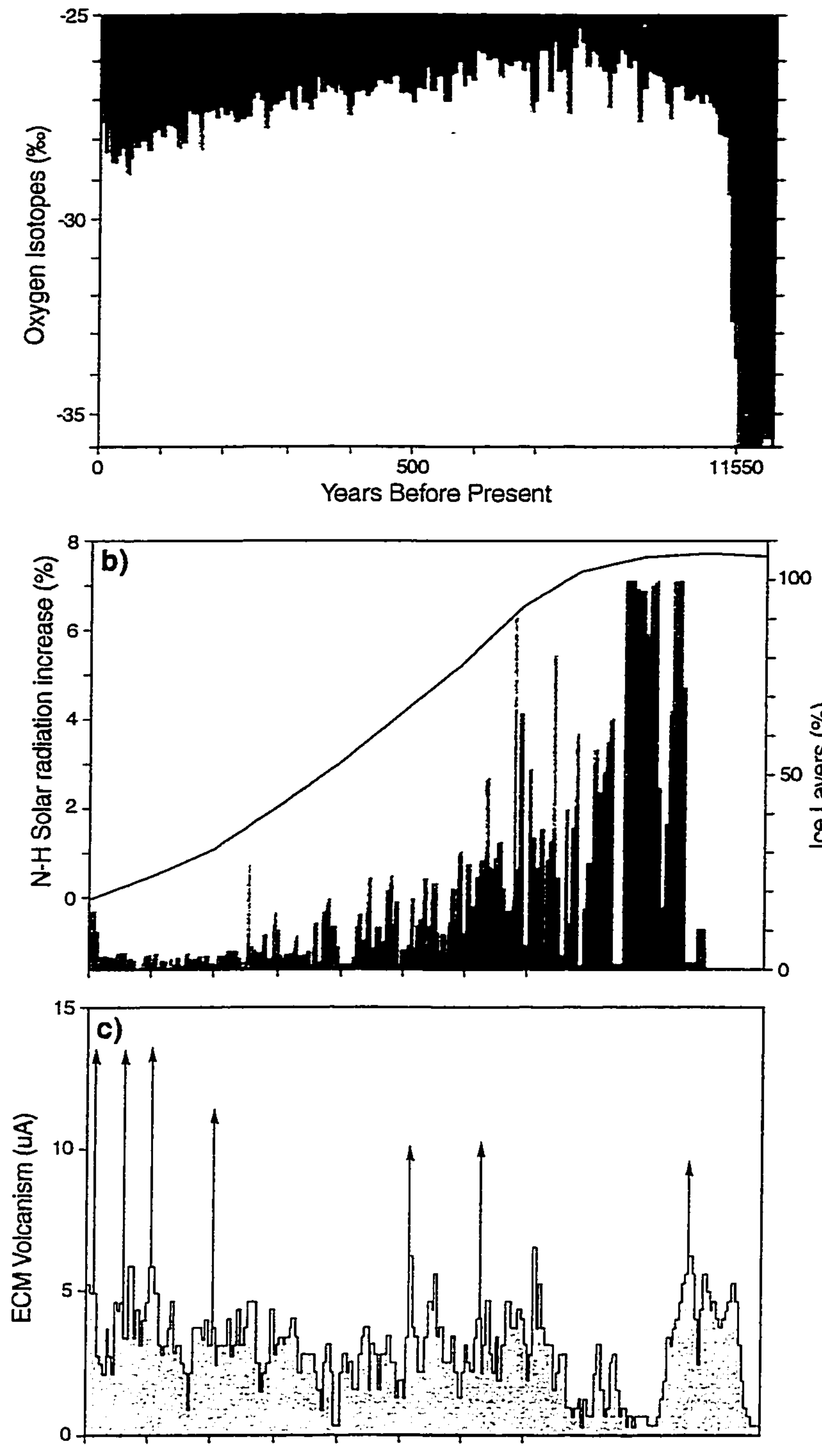

Figure 5. a) Oxygen isotopes ($\delta^{18}O$) for the Holocene section of core A84, b) melt layer concentration, and c) volcanic activity. For a comparison of the four Agassiz Ice Cap ice core records see Fisher & Koerner (1994) and Fisher *et al.* (1995).

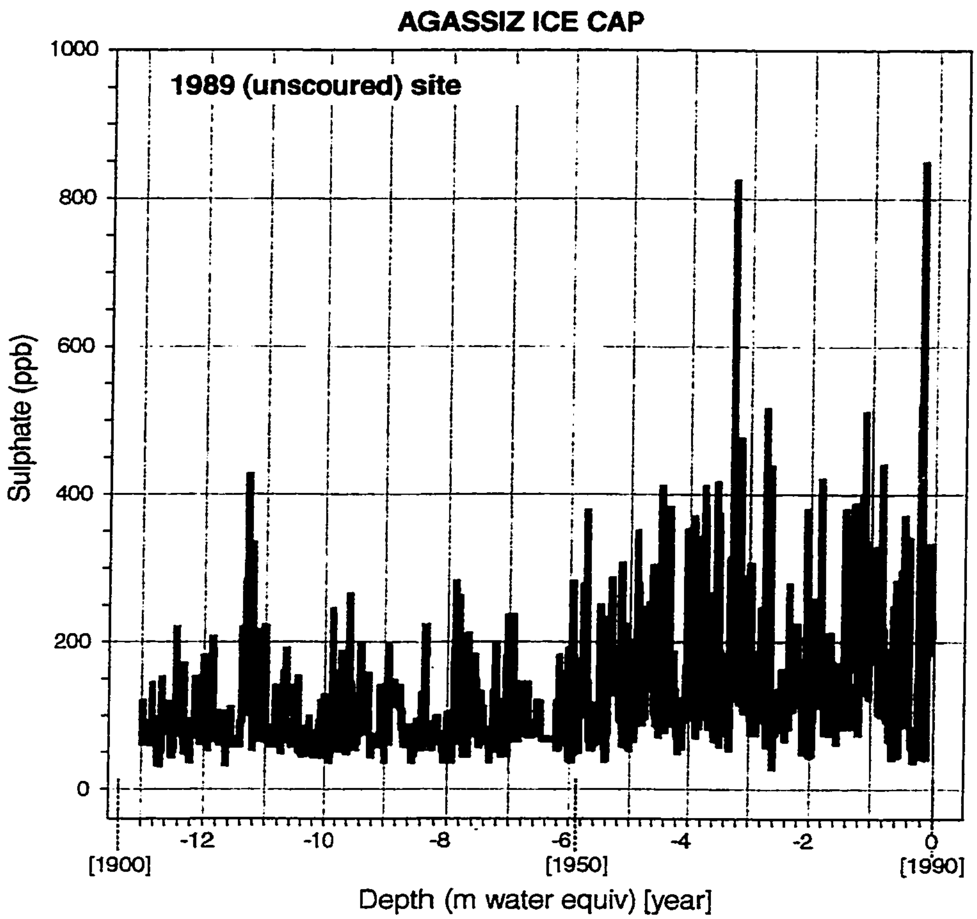

Figure 6. Concentration of sulphate (in parts per billion (ppb)) in a shallow core recovered from the unscoured zone of the Agassiz Ice Cap. A high sulphate, at approximately 11 m depth, is associated with the Katmai volcanic eruption, dated at 1912. The marked increase in concentration in the middle of this century (approx. 6 m depth), is attributed to anthropogenic pollutants.

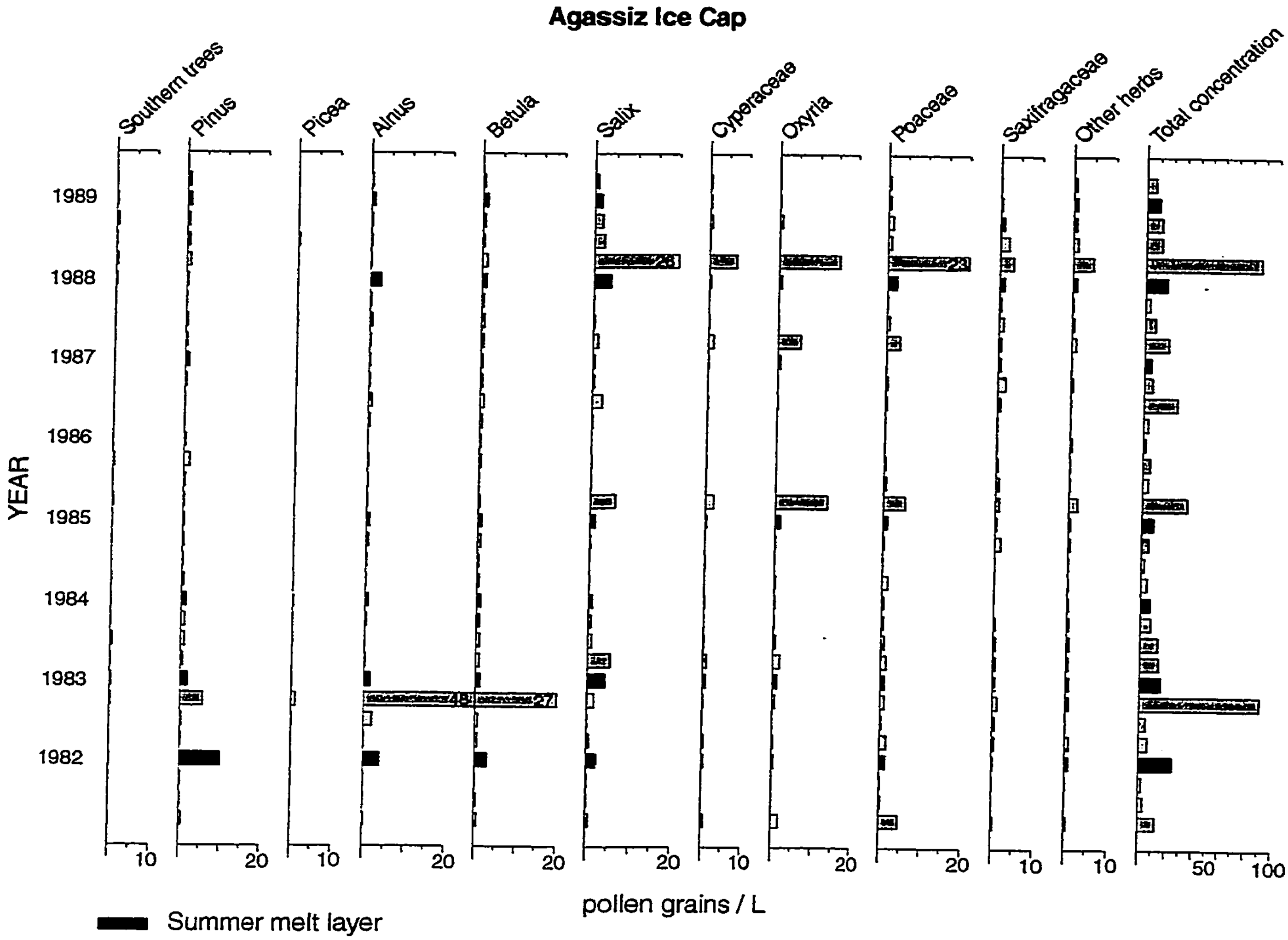

Figure 7. Seasonal pollen variations (number of grains/L) in the snow layers of the Agassiz Ice Cap during an eight-year period.

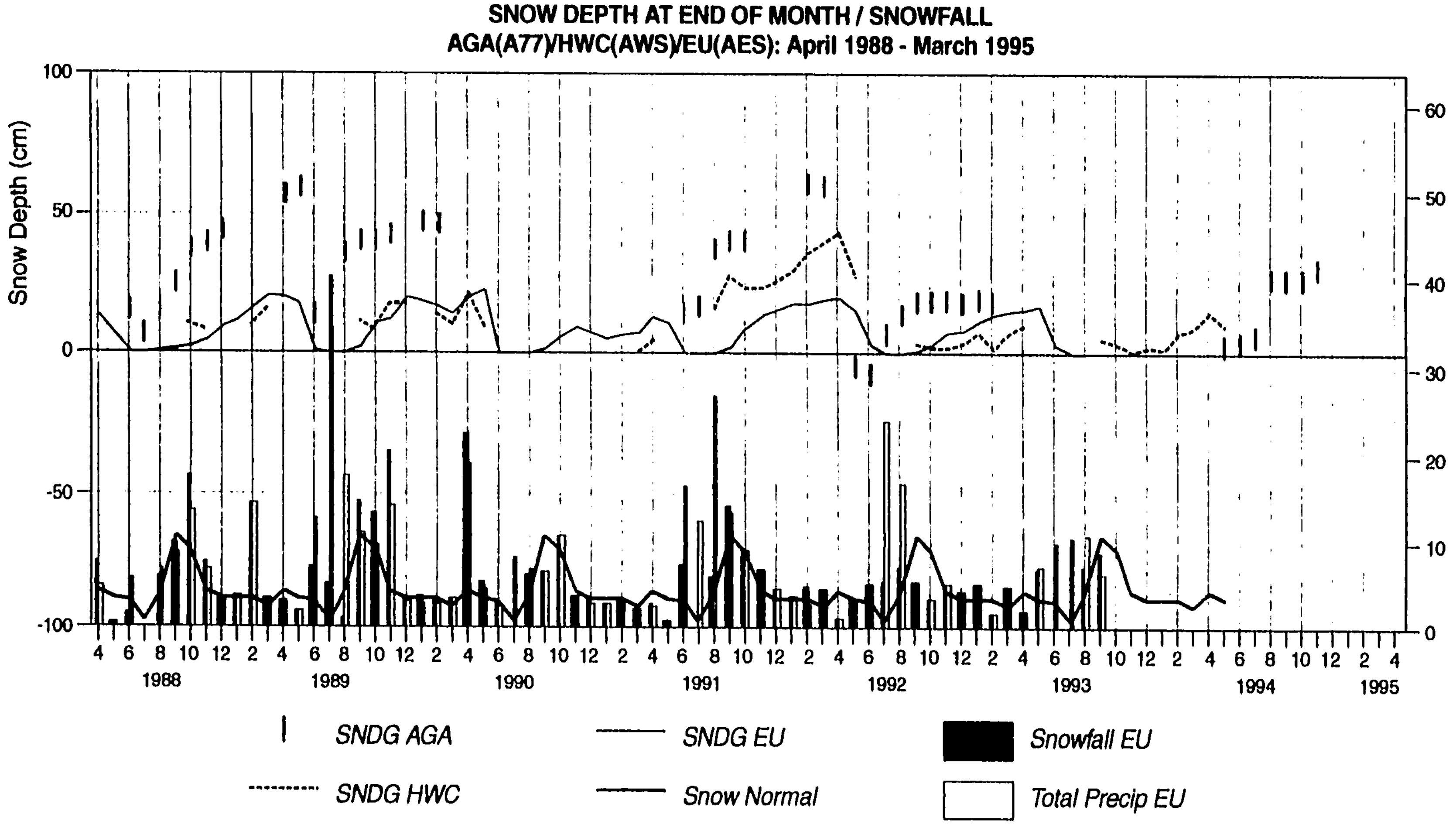

Figure 8. Top: Snow depth (in centimetres) at the end of each month measured using an autostation snow depth sounder at A77 on the Agassiz Ice Cap (horizontal bars) and at Hot Weather Creek (dashed line) and a ruler at Eureka (solid line). Bottom: Total monthly precipitation (in millimetres; open bars), total monthly snowfall (in centimetres; solid bars), and 30 year normal monthly snowfall for Eureka for comparison. SNDG = snow on the ground, AGA = Agassiz Ice Cap, HWC = Hot Weather Creek, EU = Eureka, Snow normal = 30 year normal snow fall.