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Holocene environmental changes and climate development in Greenland

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1 Introduction

1.1 Aims and framework

The primary aim of this report is to give an overview of the Holocene environmental and climatic changes in Greenland and to describe the development of the periglacial environment during the Holocene. Special emphasis is given to the influence of the ice sheet on its surroundings, both in terms of time (with respect to the response of the biosphere to deglaciation or ice sheet proximity) and in space (through the influence of the ice sheet on the regional climate, more specifically on temperature and aridity). Published records are reviewed, and regional trends are summarized in the chapters below. A range of different natural archives is available for such studies, including ice-core data, marine records, and continental sources of information, including peat profiles and lacustrine records. Because of the high number of lakes in all ice-free areas of Greenland, the lacustrine records offer the opportunity to get a spatial overview of past changes in environment and climate as well. This report focuses on (palaeo-) ecological studies, as it is intended to assemble basic information for future studies on adaptation of the biosphere to changes in climate. There is a bias towards pollen- and macro-remain-based reconstructions of past changes, as these dominate performed palaeoecological studies in Greenland; unfortunately, only a limited number of studies exist that include more modern proxies such as diatoms or chironomids (climate-indicators), but where available in the literature, these have been included.

The report starts with an introduction where the current climatic and biological zonation of Greenland is discussed together with an overview of the geology of Greenland (on the full geological timescale) in order to put the following sections in perspective.

Chapter 2 discusses the ice sheet history of Greenland from the Last Glacial Maximum (LGM) onward where special emphasis is given to the spatial variability of deglaciation at the onset of the Holocene. To enhance the readability of this chapter, we decided to divide Greenland in 4 geographical regions, which is also done in Chapter 3.

Chapter 3 forms the major part of this report, and provides summaries of reconstructions of temperature, precipitation, vegetation and other climatic and environmental parameters. These climate reconstructions are most often based on different climate-indicators, both biological (i.e. organisms, e.g. diatoms) and geochemical (e.g. oxygen isotopes or Pb-pollution). Unfortunately, major parts of Greenland have not been investigated in detail with respect to Holocene environmental development, inevitably leading to gaps in our overview. Also, a lot of information was available from e.g. the Greenland ice-core records, but we only summarize the most relevant papers in this report.

Chapter 4 continues with an introduction to the so-called "transfer function approach", where the development of training sets and mathematical techniques are discussed, followed by an overview of the work that has been done in Greenland.

Chapter 5 summarizes the period right after deglaciation as reconstructed from many lake-records, and discusses delayed responses of parts of the vegetation.

1.2 Present-day climatical and biogeographical trends in Greenland

Present-day climate

Greenland encompasses more than 22 degrees of latitudes (~60°–83° N) resulting in large differences in climate conditions. There is a strong north/south gradient in mean winter temperatures with average January temperatures around –32°C in the north and of –5°C in the south (Cappelen et al. 2001 cited in Nielsen 2008). Owing to the stabilizing influence of the Greenland ice sheet, summer temperatures are rather similar from north to south averaging between 3 and 7°C. However, the number of months with mean temperatures above freezing increases from 1 to 7 going from north to south (Funder 1989). Additionally, strong local gradients in climate exist as maritime conditions change to a more continental climate with distance from the coastline.

A strong north/south trend also exists in the amount of annual precipitation with increasing precipitation towards the south. Precipitation can reach a maximum of 2,500 mm/y in the southernmost parts, whereas the northern parts of Greenland only have precipitation levels of 120–190 mm/y (Cappelen et al. 2001 cited in Nielsen 2008). There are arid conditions in several other parts of the country as well. For instance, the Kangerlussuaq region in Southern West Greenland is situated in the rainshadow of the Sukertoppen icecap, and precipitation here averages only 150 mm/y.

Climatologically, Greenland is often divided into an eastern and a western zone. The driving mechanism for the differences in climate between the two zones can be found in the thermohaline circulation that influences the regional climate to a large extent. Anderson et al. (2004) indicate five main oceanic currents that are important in the area above 60°N (after Cronin 1999, Ruddiman 2001). As can be seen in Figure 1-1, the East Greenland Current (EGC) transports cold, low-salinity water southward along the eastern coast. The EGC carries both icebergs and multi-year pack ice from the Arctic Ocean, as well as sea ice from the fjords of East Greenland (Andresen et al. 2004). Along the southeast coast of Greenland, this EGC mixes with a branch of the warm and saline Atlantic Current called the Irminger Current. The EGC continues to flow around Cape Farewell and partly even up the westcoast as a mixed water body, with the ice quickly melting (Andresen et al. 2004). The wide transitional zone in the Denmark Strait between the polar waters of the EGC and the Atlantic waters of the Irminger Current is defined as the oceanographic Polar Front (Dietrich et al. 1980 cited in Andresen et al. 2004). This Polar

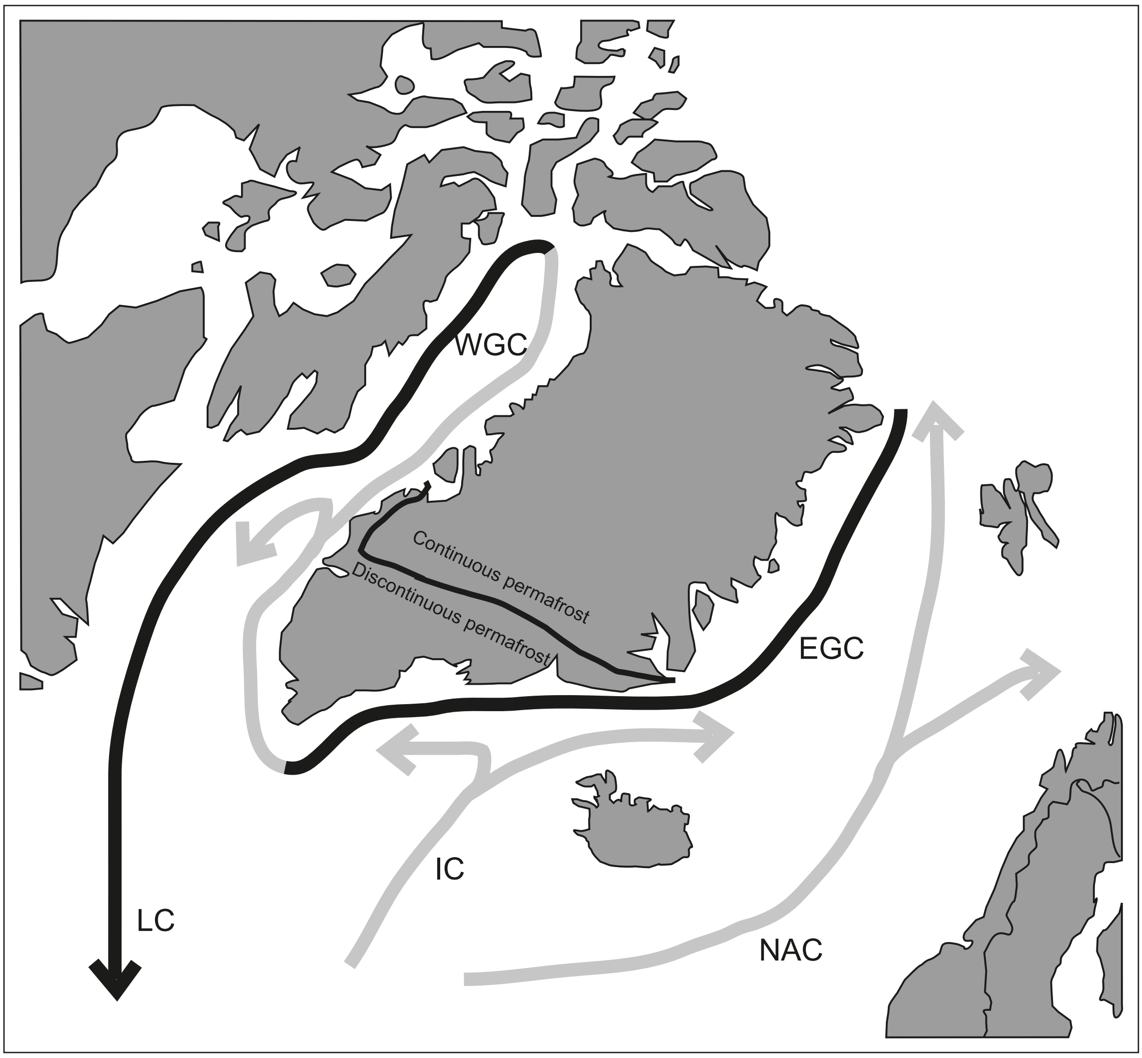

Figure 1-1. Oceanic currents around Greenland and the North Atlantic Islands. Abbreviations: NAC: North Atlantic current (warm); IC: Irminger Current (warm), ECG: East Greenland Current (cold), WGC: West Greenland Current (warm), LC: Labrador Current (cold). The distribution of discontinuous/continuous permafrost is also shown. After Anderson et al. (1999).

Front has a strong influence on the atmospheric circulation in the region, influencing annual, interannual and even longer-timescale weather and climate regimes. The western part of Greenland is influenced by the warm West Greenland Current that flows northward along the coast. As a result, the western side of Greenland in general is warmer than its eastern counterpart, for instance resulting in a more northward position of the discontinuous/continuous permafrost boundary (see Figure 1-1). More information on climatic trends and specific data for the Kangerlussuaq region can be found in Nielsen (2008).

Present-day biology

Biogeographically, there is also an important gradient going from the high arctic in north Greenland to the subarctic in south Greenland. North of 67°N, most of Greenland is characterized by continuous permafrost, and vegetation is dominated by dwarf shrub tundra. North of 80°N (and on higher elevations) vegetation is sparse and consists of herbs, mosses and lichens. Washington Land is an example of low biodiversity, with only 60-odd vascular plant species, and 4 species of land mammals (Bennike 2002).

The shrub and dwarf-shrub vegetation of East and West Greenland is characterized by willow (*Salix*) and dwarf birch (*Betula nana*) shrubs and heath plants (Ericaceae). Along the outer coast crowberry (*Empetrum*) is the most common plant (Feilberg et al. 1984), with blueberry (*Vaccinium*), mosses and sedges (Cyperaceae) as other major constituents of the vegetation. More inland, plants of heath grow to be slightly taller than in coastal areas. Blueberry (*Vaccinium*) is more common there, and is accompanied by plants such as alpine cinquefoil (*Potentilla alpestris*), mosses and common harebell (*Campanula rotundifolia*; Feilberg et al. 1984). The typical plant community on slopes is dominated by herbs, with up to 50 different species.

Because of the influence of the thermohaline circulation, the vegetation in West Greenland is more maritime than in East Greenland, and it is characterized by *Salix glauca*, *Betula nana* and several Ericaceae shrubs. The areas around the head of Kangerlussuaq fjord consists of well-drained sandy soils which are characterized by a steppe-like plant community. The patchy vegetation is dominated by grasses and grass-like plants, with a few scattered herbs in between (Feilberg et al. 1984).

An exception to the dwarf-shrub and shrub vegetation of the major areas of Greenland is found in the southernmost tip of the country. In the interior between Nanortalik and Ivigtut in South Greenland, valleys are covered by a low forest of birch (*Betula pubescens*) and willow (*Salix*) trees. More information on biogeographical trends and specific data for the Kangerlussuaq-region can be found in Nielsen (2008).

1.3 Geology of Greenland

Precambrian to Pliocene

A major part of Greenland is made up of Precambrian crystalline rocks, probably including the areas under the current ice sheet (Funder 1989). The oldest areas form a basement shield composed of folded gneisses. The basement shield is surrounded by sedimentary basins that developed in three periods: the Proterozoic, the Cambrian-Silurian and the Devonian-Neogene, in which sandstones and carbonates have been deposited (Henriksen 2008). In the late-Mesozoic, the present shape of the country was formed through the formation of the Baffin Bay/Labrador Sea in the west and the Greenland/Norwegian seas in the east, as the result of active ocean floor spreading. During this time, vast amounts of terrigenous detritus were deposited in the West Greenland shelf area and in a newly formed sedimentary basin that extended along the entire East Greenland coast (Funder 1989). As a result of the ocean floor spreading during the Tertiary around ~54 Ma and ~ 36 Ma ago, large parts of Greenland were uplifted.

However, the present landscape of Greenland is to a large degree formed by ice. The Quaternary is a period that has been characterized by multiple glaciation cycles, and the first major glaciation of Greenland probably occurred as early as 2.4 Ma ago near the base of the Quaternary, as indicated by an erosional disconformity in the marine deposits of East Greenland (Funder 1989). New evidence from drill cores from the shelf indicates that there might have been glacial activity as long as 7 Ma ago (Henriksen 2008).

Differences in bedrock between the northern and southern parts of East Greenland have resulted in different dimensions of the fjords (Funder 1989). South of Scoresby Sund (for location see Figure 2-1), where major part of the bedrock is of Precambrian age, fjords are narrow, unbranched, closely spaced and less than 50 km long. To the north (with bedrocks of Caledonian age), they are branched and penetrate the inland much deeper. The northern area has a topography dominated by gently undulated high mountain plateaus at 1,500–2,500 m.a.s.l and extensive areas of lowland near the outer coasts. Between 68 and 70°N the landscape is dominated by arêtes and pinnacles formed in the thick Tertiary basalt plateau. The valleys are filled with outlet glaciers from the ice sheet, and smaller ice caps are present, fused to the inland ice sheet on the western side. To the south of this region, the gneiss-dominated bedrock forms alpine landscapes with decreasing elevation to the south (the highest peak is 3,700 m.a.s.l).

Quaternary Geology of West Greenland

The floors of all major valleys in West Greenland are covered by fluvial and glaciofluvial sediments, which occur as outwash plains and fluvial terraces deposited by braided river systems. These flood-plains also form a source area for dust storms, and small-scale aeolian features can be found at the margins of the valleys. Along the valley sides, local kame terraces occur. Eskers are rare in this region. As a result of isostatic rebound and sea-level changes, marine deposits are found in the coastal regions up to a height of 120 m.a.s.l, and mollusks found in these deposits have been used to establish a geochronology for the Quaternary in West Greenland (Funder 1989, Section 2.2).

Continuous permafrost occurs north of 67°N in West Greenland (see Section 1.2), and periglacial features such as pingos, ice-wedges and rock glaciers are mostly found above this geographical limit. Palsas are most common in the border area between continuous and discontinuous permafrost.

Because of erosion by the major Sisimiut glaciations (correlated with the Last Glacial Maximum (LGM) or Marine Isotope Stage (MIS) 2) little evidence of earlier glaciations is preserved, and is only known from scattered localities. The deglaciation following the LGM will be discussed separately in Section 2. The uphill areas in West Greenland are dominated by a large number of lakes, as well as wet meadows, peatlands and thin aeolian deposits. The Holocene development of the landscape will be discussed in detail in Chapter 3. Three subperiods are often recognized within the Holocene records from Greenland: an Early Holocene period during which flora and fauna colonized recently deglaciated terrain and which is characterized by generally cold and abrupt centennial-scale oscillations; a Middle Holocene with a thermal optimum; and a Late Holocene that is characterized by decreasing temperatures. The ages for these different subperiods are given in Table 1-1, but will differ slightly from region to region.

Quaternary Geology of East Greenland

In general, the ice sheet along the east coast has a lower mass balance and lower ice movement velocity compared to the western part of Greenland, resulting in a less erosive regime (Funder 1989). Because of the less-erosive nature of the ice sheet in East Greenland, more till beds have been preserved than in other regions in Greenland, where the LGM-glaciation redistributed virtually all underlying sediments. Weathering of the till-deposits suggests a series of glaciations rather than one or two single events. Moraines are found along all the fjords, commonly occurring at topographical breaks. Again, fluvial deposits are found in extensive floodplains in fjords and valleys, and marine deposits can be found up to 110 m.a.s.l.

Table 1-1. Division of the Holocene in subperiods. Ages follow those reported by Johnsen et al. (2001) and Dahl-Jensen et al. (1998). Note that in different regions of Greenland, changes in for instance the warmer climate of the Middle Holocene to the cooling trend during the Late Holocene have occurred during different times, and regional transitions from one period to another will therefore slightly differ from those reported here.

Period	Age (a BP)	Age (Gregorian)
Late Holocene	4000 BP–present	2000 BC–2008 AD
Middle Holocene	8000–4000 BP	6000–2000 BC
Early Holocene	11,500–8000 BP	9500–6000 BC

2 Late Pleistocene and Early Holocene deglaciation in Greenland

The earliest glaciation of Greenland took place already at 2.4 Ma BP, as described in Section 1.3. It is very likely that there have been numerous glaciation/deglaciation cycles since then, but due to the erosive nature of especially the last two glaciations, little evidence to prior glaciations remains in situ on the Greenland mainland. Below, we summarize the deglaciation after the last full glacial, with a special focus on the spatial variability in the deglaciation (Sections 2.1–2.3). Section 2.4 provides an overview of the Holocene ice sheet variability in different locations on Greenland.

2.1 Deglaciation in East Greenland

Following the LGM, the Greenland ice sheet started to retreat between 13,000–11,000 BC in East Greenland (Funder et al. 1998). The Younger Dryas period is assumed to have been too dry and too cold to enable ice sheet or glacier readvances (Funder et al. 1998). So the ice retreat was only interrupted during the Milne Land stade (correlated to the Preboreal Oscillation (PBO) at 9300–9150 BC) when glaciers covered the outer fjord region in East Greenland (Hjort 1979 cited in Wagner and Melles 2002). Renewed glacier recession commenced after the PBO between 9200 and 8000 BC (Björck et al. 1994a, b), and present-day positions were reached within a 1,000 years (Wagner et al. 2000). The annual rate of recession was dependent on local factors such as topography, and periods of standstills and small readvances are likely to have occurred (Wagner et al. 2000). Below, we summarize the review by Cremer et al. (2008), who have reconstructed the deglaciation of East Greenland between 70°N and 82°N. The authors collected all published radiocarbon dates from basal parts of lacustrine sediment records, and present several previously unpublished results. A total of 24 sites over this zone were studied (see Figure 2-1).

Eastern North Greenland (north of 77°N)

Only 4 sediment sequences have been adequately dated in this part of Greenland, and the oldest basal date is 8785 yr BC. All other basal dates are clearly younger, ranging between 5300 and 5700 BC. This deglaciation date is late compared to other outer-coast sites of Greenland and may be the result of an unusually broad shelf (Bennike and Björck 2002).

Northern East Greenland (75–77°N)

There are lake records available for 2 different regions. Basal dates from Store Koldewey (SK in Figure 2-1) yielded minimum ages of 7700 and 7200 BC. The oldest dates from Hochstetter Forland are surprising, resulting in ages of 13,300 and 14,700 BC. Because the latter samples were bulkdates performed on samples with a loss-on-ignition of only a few percent, Cremer et al. (2008) considered these dates to be contaminated. The oldest reliable basal dates from this region lie between 4800 and 6400 BC, indicating a deglaciation late in the Early Holocene.

Central East Greenland (70–75°N)

A relatively large amount of information is available for this region, and minimum deglaciation ages are available for 15 lakes. The region around Zackenberg deglaciated approximately 7000 a BC. Further south, some regions were already ice-free at 8300 BC. The oldest available date reaches 13,600 BC, but this was again measured on a bulk organic carbon sample, and was regarded with caution owing to possible contamination. The mean deglaciation dates for Scoreby Sund range between 5100 and 9700 BC, suggesting that some locations were ice-free at a relatively early stage, predating 9000 BC. Analyses on marine shells indicate a similar pattern of deglaciation (Bennike and Björck 2002), with an earlier retreat of the ice sheet margin in central East Greenland.

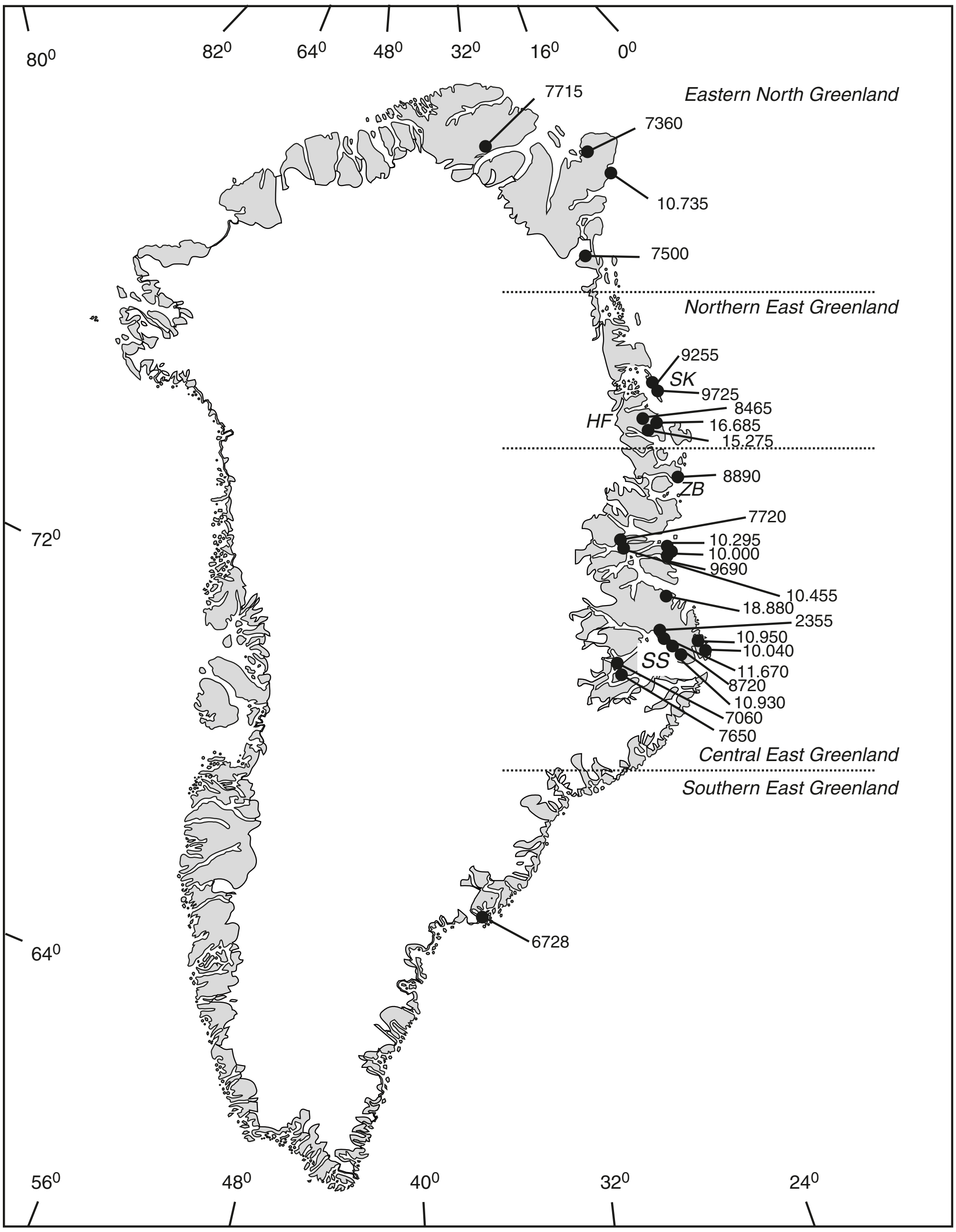

Figure 2-1. *Map showing the basal dates of lacustrine and marine sites in eastern Greenland discussed in the text, providing age estimates of deglaciation for the respective regions (after Cremer et al. 2008). Abbreviations: SK = Store Koldewey; HF = Hochstetter Forland; ZB = Zackenberg; SS = Scoreby Sund, ages in calibrated years before present.*

Southern East Greenland (south of 70°N)

No basal dates of lacustrine deposits have been published for this region, and only one radiocarbon date derived from a peat monolith was published, giving a basal date of 4700 BC. Foraminifera-based dates off southeast Greenland point to an extensive ice-coverage of the shelf prior to 15,000 BC.

In general, all dates have a certain degree of unreliability. The sediments are low in organic content and dates are often performed on bulk material, on different parallel cores and (especially for the sediments that seem to be deposited before the onset of the Holocene) the chances of contamination by marine carbon or Mesozoic coal-particles are high.

The variability in the reconstructed timing of deglaciation might be the result of the relative position of the lakes to the inland ice-margin, where inland areas are likely to have experienced deglaciation later. The low number of available sites precludes any reliable deductions of temporal differences in deglaciation over different regions, although it seems that central East Greenland was ice-free earlier than the southern and northern provinces, possibly due to topographical differences and the width of the shelf area.

Concluding, all the oldest reliable dates of plant remains from lacustrine sediments in East Greenland indicate that deglaciation did not start prior to the Early Holocene (here defined as 9500 BC).

2.2 Deglaciation in West Greenland

West Greenland might be divided into three subregions (as indicated in Figure 2-2): a southwestern zone, including the Disko-Bugt region and the large ice-free region to the south (including the Kangerlussuaq-area); a middle part including the region between the Disko Bugt in the south and the Melville Bugt in the North, which shows only a very narrow ice-free area; and northwestern Greenland, defined here as the region between Thule and the Naires Strait.

Northern West Greenland

Both samples from Greenland and from Canada indicate early ages for deglaciation of the northern part of the Nares Strait, the oldest dates being 9200 and 9300 BC. Deglaciation of the central part of Nares Strait is dated to 6700 BC (Canadian side) and 5900 BC (Greenland side) (Bennike and Björck 2002), which means that this region was one of the last coastal regions to be deglaciated, well into the Holocene (with 2,000 to 4,000 years; Bennike 2002). Raised marine deposits suggest that the northern part of Washington Land became ice-free earlier than the southern part (Bennike 2002).

Central West Greenland

This region is characterized by the extremely limited extent of the ice-free land mass, and only 2 dates have been found in the literature for the region that indicate the youngest deglaciation age. The reported ages are 7500 and 8500 BC which is slightly younger than the dates that are reported from the Disko Bugt area to the south (Bennike and Björck 2002). A recently obtained date on a whale bone derived from Melville Bugt yielded an age of 7300–7000 BC, suggesting that a position of the ice margin similar to the present position was reached earlier in central West Greenland than in other regions of Greenland, possibly due to the topographical settings (narrow ice-free land, deep offshore waters).

Southern West Greenland

The oldest date from the mouth of the Disko Bugt bay is 8300 BC, whereas the oldest dates from the inner part of the bay reach 7600 (shell) and 7700 (bulk lake sediments) BC (Bennike and Björck 2002). These dates are approximately 2000 years younger than the oldest dates from eastern Greenland, and also younger than those derived from central West Greenland (see above). A possible explanation for the late deglaciation at Disko Bugt is given by Weidick (1996), who argues that a group of islands might have functioned as a pinning point or threshold for the Disko Bugt glacier.

The area south of Disko Bugt (including the Kangerlussuaq region) is the most extensive ice-free part of Greenland, and fjords are generally narrow in this zone. It is therefore argued that the recession of the ice front was caused mainly through ablation, and not through calving (Bennike and Björck 2002). This would also explain the younger dates for deglaciation that are found in the inner parts of this region (Bennike and Björck 2002). The oldest reported dates from the inner part of the Kangerlussuaq fjord is 6000 BC, and a lake situated close to the middle part of the fjord provided a date on basal gyttja of 7100 BC (Willemse 2000). Along the outer coast, the oldest dates have ages of about 8300 BC. Therefore, a clear trend of later deglaciation can be seen going from the coast towards the current ice margin.

Again, all dates have several sources of errors, but nevertheless certain conclusions can be drawn. First, in southern West Greenland there is a clear signal of earlier deglaciation in the coastal zone, and later deglaciation in the inland regions. Second, there is a lot of spatial variation in the deglaciation history of West Greenland, probably as a result of differences in shelf width/water depth and also of topography of the land itself.

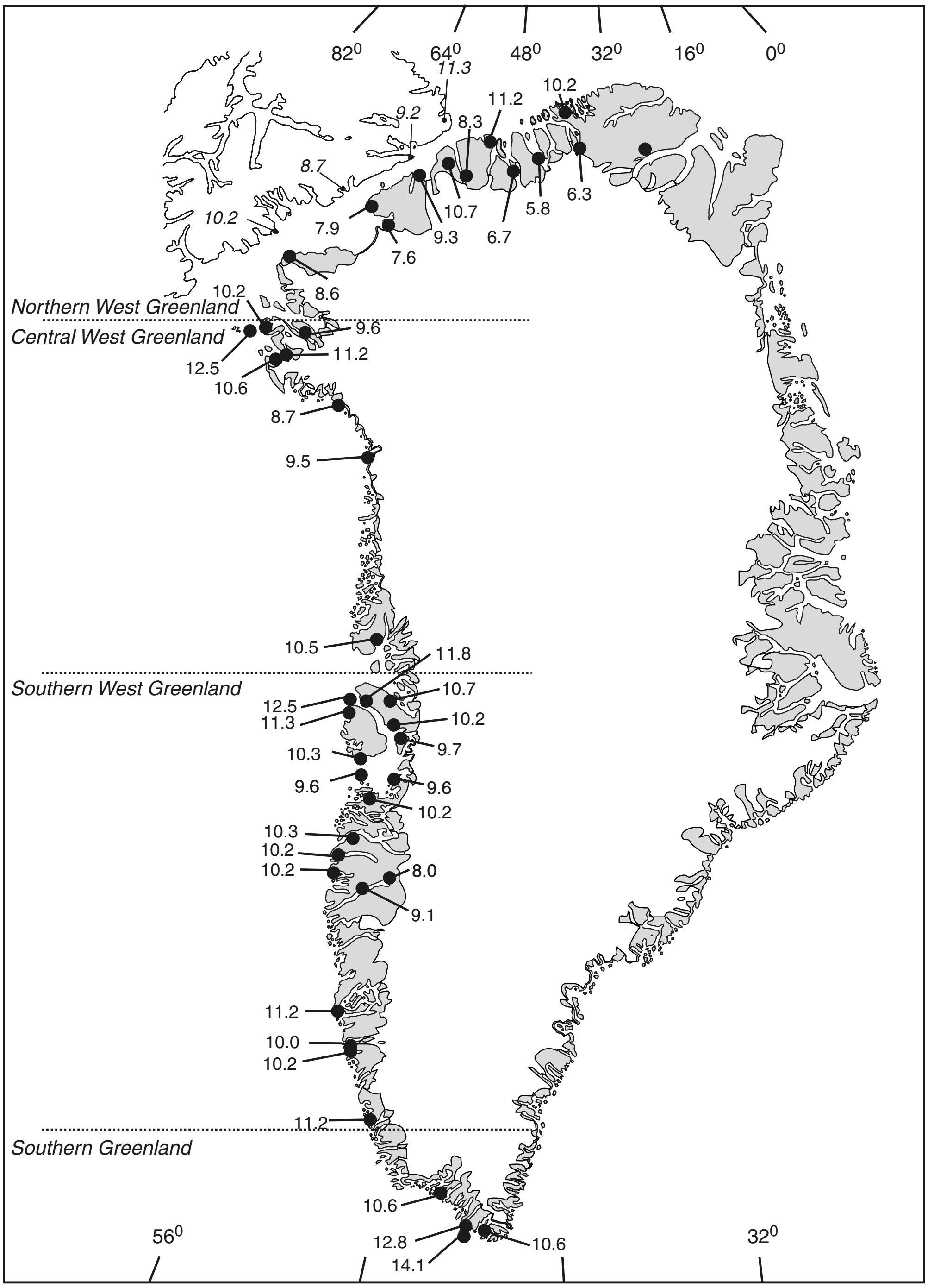

Figure 2-2. *Map showing the basal dates of lacustrine and marine sites in western Greenland discussed in the text, providing age estimates of deglaciation for the respective regions (after Bennike and Björck 2002) ages in calibrated years before present.*

2.3 Deglaciation in South Greenland

The southern tip of Greenland is situated at 60°N, which means that it is strongly influenced by the oceanic currents that are also associated with the marked warming of Greenland Interstadial 1 (e.g. Bennike et al. 2002; see Section 3). Together with the narrow shelf, an early deglaciation is expected at Kap Farvel (Bennike and Björck 2002).

It is likely that the Greenland ice sheet extended out to the shelf margin during the LGM, and that the ice thickness must have been at least 1,500 m over the outer coast (Bennike et al. 2002). This means that all of South Greenland, including the highest mountains in the region, must have been covered by ice. A number of lakes cored by Bennike and Björck (2000) have shown that the sediments contain a late-glacial signature, and the oldest sediments date back to 10,100 BC (Bennike and Björck 2002). The other records provide dates that are of Younger Dryas age. Bennike et al. (2002) sampled several other lakes, and conclude that the major part of the recession of the ice sheet must have occurred relatively late after the LGM and quickly, maybe from 12,000 to 10,000 BC.

Going north towards the region identified in this report as West Greenland (Section 2.2), dates are all centered at around the onset of the Holocene or the Early Holocene (9500–6000 BC; Table 1-1) at the latest, and do not indicate late-glacial ice-free conditions. Southeastern Greenland shows a large absence of dated sediments, possibly due to the small ice-free fringe of land that is present here, together with the inaccessibility of the region.

Concluding, only the southernmost tip of Greenland shows reliable evidence of ice-free land prior to the Holocene. Other regions only became ice-free during different stages of the Early Holocene.

2.4 Holocene ice sheet variability

It is expected that during the Holocene Thermal Maximum (between ~6000 and 2000 BC; more information in Chapter 3), the ice margin retreated towards more inland positions over large parts of Greenland. There is little hard evidence for the extent of the ice marginal retreat, but van Tatenhove et al. (1996) provide a tentative reconstruction. van Tatenhove et al. (1996) have worked on Holocene ice sheet variability in the Kangerlussuaq area, and dated two moraine systems: the Umivit/Keglen Moraine System, situated near the harbor of Kangerlussuaq, and the Ørkendalen Moraine System that is situated 1–2 km from the present ice margin. The Umivit/Keglen Moraine System is dated to 7.3 ^{14}C ka BP (~6100 BC). The Ørkendalen Moraine System is dated to 5.6–6.0 ^{14}C ka BP (~5000 BC). After the Ørkendalen period, the ice margin retreated behind its present margin during the Mid-Holocene climatic optimum. The pre-Ørkendalen-deglaciation-rate amounts to about 40 km per thousand years, and there was at least 800 years (but possibly much more) available for retreat after the formation of the Ørkendalen system. Extrapolating the 40 km/1,000 years retreat rate, and even including a slowdown of the retreat to zero, van Tatenhove et al. (1996) have reconstructed an inland retreat of at least 16 km beyond the current ice-margin. The authors do emphasize that this is a tentative suggestion.

Following the Holocene Thermal Maximum the Greenland ice sheet began to expand during the so-called Neoglacial in the Late Holocene. For the major part of Greenland, the readvance reached its maximum extension during the Little Ice age (LIA) between 100 and 200 years ago (or 1800–1900 AD). In South Greenland the Narssarssuaq stade re-advance exceeded the historical (Neoglacial) advance (Weidick et al. 2004). Cedar Lake is situated inside the two outer moraines attributed to the Narssarssuaq stade re-advance, and its oldest sediments date back to 750 AD, clearly predating the Neoglacial/LIA maximum of the Greenland ice sheet. The glacier probably left the Narssarssuaq moraine not more than 2,000 years ago. A second example is a moraine system encountered northeast of Disko Bugt, where ages of around 1000 BC are reported for the Drygalski Moraines (Kelly 1980 cited in Bennike and Sparrenbom 2007). During the readvance, a lot of sediments were reworked, and historically documented moraines were formed. Hereafter, recession took place (e.g. Bennike and Sparrenbom 2007).

In general, there is not a lot known about spatial variations of the Greenland ice sheet advances during the Holocene, although a general accepted summary seems to be that the deglaciation was quick in many locations, starting in the Early Holocene, and not lasting more than 1,000 years. During the Early to Middle Holocene a slower retreat of the ice sheet took place, reaching minimal positions between 2500–1000 BC. Hereafter, ice sheets started to expand again up till the Little Ice Age ca 200 years ago, after which a final retreat took place.

3 Holocene climate variability and vegetation development in Greenland

3.1 Terrestrial records from East Greenland

Climate – temperature and precipitation

Wagner et al. (2000) and Wagner and Melles (2002) have reconstructed climate in central East Greenland using multiple lake records, the most important ones from Geographical Society Ø and close-by islands (Figure 3-1a). They used sedimentary characteristics such as Total Organic Carbon (TOC), Total Nitrate (TN) and opal (biogenic silica) content as well as pollen to infer past changes in climate.

The region was glaciated during the Milne Stade (which is correlated to the Preboreal Oscillation at approximately 9200 BC), and after deglaciation it probably took almost 1,000 years before the limnic system got into equilibrium. Opal content of the sediments is initially low (4%) and TOC values are stable around 1.5–3%.The major explanation that is offered for this delayed response is the unavailability of nutrients through delayed soil formation and other environmental processes in the catchment (Wagner et al. 2000).

Reconstructed temperatures are high for the period between 7000 and 3000 BC, and summer temperatures are estimated to have been 1–2°C higher than present-day temperatures (Bennike and Funder 1997; see Figure 3-1b). This temperature evolution is indicated by the immigration of *Betula* and sharp increases of TOC (to 2–8%) and opal content (to 4–12%), suggesting an abrupt climate warming. The reconstructed vegetation suggests that the climate must initially have been relatively dry. This reconstructed climate development shows a strong resemblance to the Renland ice core record of Holocene temperatures (see Section 3.5). The onset of the warm period might have been earlier than 7000 BC, however, as the immigration of *Betula* might have been delayed as the result of a migrational lag (see below, Section 3.1 – vegetation development).

The period between 3000 and 1000 BC shows a gradual decrease in temperatures combined with high amounts of precipitation. Cooling is indicated by steadily decreasing values for most biogeochemical parameters (e.g. a decrease in opal content from 12 to 5%), an increase in grain size of the sediments (percentage of grains >20 µm increases from ~5 to 40%) and a decrease of *Betula* pollen in favor of *Salix*. The pollen assemblages suggest wet soils and moist conditions during this time interval, which is supported by findings of *Botryococcus* and *Pediastrum,* taxa that indicate a greater influx of nutrients due to larger snowmelt. Wagner et al. (2000) conclude that the climate deterioration, that started around 4500 BC, was primarily induced by a rise in snow accumulation rather than by a temperature decline. A possible reason for the changes in climate is the retreat of Atlantic water masses to the southern central part of Greenland (Koç et al. 1993).

The period from 1000 BC onwards is characterized by a further decrease in organic production only interrupted between 1450–1750 AD. A combination of biogeochemical proxies and palynological data suggests very cold and dry conditions between 1000 BC–1150 AD (Wagner et al. 2000, Figure 3-1b). A short period of increased temperatures is reconstructed during the time period 1050–1450 AD from the pollen-record, which slightly lags the well-known Medieval Warm Period (vaguely recorded in the Renberg core at 950 AD). A cold episode is reconstructed between 950–1050 AD, although the changes witnessed in the records might have been an effect of increased niveo-aeolian activity as well as of cooling. A recent warming is indicated by the increase in shrub pollen (mostly *Salix*) in the top-samples of the lake sediment record of Basaltsø.

In northeast Greenland, a similar climate evolution is witnessed in the grain-size record of Hjort Sø, Store Koldewey (Wagner et al. 2008, Figure 3-1a). A Holocene thermal maximum is reconstructed between 7000 and 3000 BC, followed by a cooling that probably starts at 4500 BC but becomes distinct from 3000 BC onwards. The Late Holocene shows minor fluctuations that resemble the medieval warming (~500–950 AD) and the Little Ice Age (up to 1800 AD). No latitudinal differences are discerned between this record and those from central East Greenland.

Figure 3-1a. *Location map of the sites discussed in Section 3.1.*

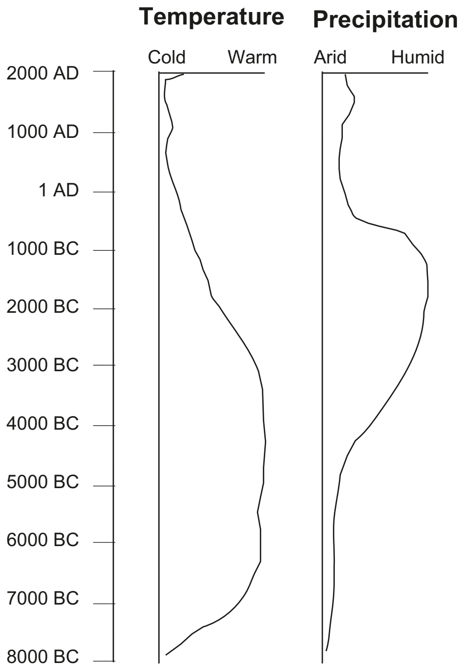

Figure 3-1b. *Changes in relative temperatures and effective precipitation during the Holocene, reconstructed for central East Greenland (after Wagner et al. 2000).*

Sedimentation

Most of the lake sediment sequences that have been analysed record a similar development of infilling processes. Often, an initial homogenous inorganic sediment is deposited directly after deglaciation, possibly in a glaciofluvial or glaciolacustrine environment. After this initial phase, eustatic sea-level rise resulted in marine inundations of many low-lying, recently formed lakes. Shells of marine bivalves and foraminifera, marine deltas and washed out boulders encountered in these records are the result of this marine inundation (Funder 1989, Björck et al. 1994b). The transition from a freshwater system to a brackish/marine system was abrupt. Still in the Early Holocene, the sites were subsequently raised to higher elevations as a result of isostatic rebound of the region, and many where positioned outside of the reach of marine incursions. After a transitional phase, the marine sediments therefore grade into lacustrine deposits. Depending on where the lake is situated (with respect to the ice sheet), the lacustrine sediments can be rather inorganic and laminated (typical for proglacial lakes) or richer in organic components. Marine deposits as described here can be found up to an elevation of ~70–110 m in central East Greenland, lakes situated above this threshold will not contain a marine section in their sediment record.

The Holocene lake sediment sequences analysed by Wagner et al. (2000) have thicknesses of 9.85 m for Basalt Sø and 2.7 m for the sediment record from Lake B1. Sedimentation was rather constant throughout the records, indicating sedimentation rates between 0.2–1 mm/y. The sediments in the uppermost part of the cores consist of organic gyttja, with an opal content of 4–8% and a TOC of 2–5%.

A sediment record from Loon Lake, central East Greenland (Figure 3-1a), has recently been published by Wagner et al. (2007). Intercalated in the ~5 m thick marine layer at Loon Lake, deposited between ~6600–5500 BC, the authors found a 0.7 m thick sandy sediment sequence with fining-up structures and an erosive basis. According to eight radiocarbon dates, the sands were deposited around 6400 BC. Although this date slightly predates the Storrega tsunami deposits found in Europe and several other places in the Atlantic, the authors suggest that the sandy sediments at Loon Lake were formed by the Storrega tsunami. Possible explanations for the slightly too-old ages are the shells that were used for dating and possible errors in the reservoir effect correction.

A sedimentological study was performed on a fluvial deltaic system encountered in the Zackenberg area (see Figure 3-1a). The raised delta system accumulated mainly between 7500 and 4300 BC. A Holocene thermal optimum was reconstructed for this period, based on pollen evidence of a rich herb flora combined with the macroscopic findings of remains of *Empetrum nigrum* and *Salix herbacea* (Christiansen et al. 2002). Podzol formation started during the Holocene thermal maximum and continued on well drained surfaces up to 950 BC–50 AD. Podzol formation probably ended due to decreased temperatures as evidenced by the presence of a less developed heath vegetation, nutrient-poor soil horizons and increased nivation activity.

Vegetation development in Eastern Greenland

The sediments that are deposited in central East Greenland directly after deglaciation are characterised by a pollen assemblage reflecting a fell field vegetation. Pollen assemblages show high values of Poaceae, Caryophyllaceae, Saxifragaceae and Rosaceae (Funder 1978, Wagner et al. 2000, Wagner and Melles 2002). Fell field vegetation is an open vegetation adapted to unstable soils, a high amount of soil erosion, and problems concerning the water balance of soils (as a result of underlying permafrost; Funder and Fredskild 1989; see also Section 5.1).

Around 6900 BC, *Betula* settles in the coastal areas of East Greenland (Fredskild 1991). Pollen values of *Betula* increase to values over 70%, coinciding with a sharp decrease in percentages-abundances of pollen of herbs and of spores. The *Betula* pollen are all attributed to the shrub *Betula nana*, which still grows in the region. Its sudden high abundances in the records are interpreted as a delayed response rather than as a climatic shift (although the species is often indicative of temperatures higher than present). This interpretation follows from the observance that in some lakes an increase is seen in organic components of the sediment before the shift in pollen assemblages occurs (for instance an abrupt increase in opal content of the sediments from 5 to 10% before the abrupt increase in pollen percentages of *Betula*; Wagner et al. 2000). Since interior land areas experienced deglaciation at a later stage than coastal regions, there has been less time available for soil development, possibly delaying the inland migration of *Betula* (Wagner et al. 2000, Wagner and Melles 2002). The extensive occurrence of *Betula* and Ericaceae (as interpreted from the pollen records) is associated with a climate warmer than today in East Greenland (Funder 1978, Fredskild 1991, Wagner and Melles 2002), a period that lasted from ~8600–3000 BC (see below).

A gradual decrease in *Betula* pollen abundance can be observed around 3000 BC, coinciding with an increase in *Salix*, Cyperaceae and *Dryas*. This transition is observed over large parts in East Greenland (Wagner and Melles 2002) and is interpreted as the transition towards a cooler environment, the trend culminating in the Little Ice Age (Wagner et al. 2000, Wagner and Melles 2002). However, *Betula* remains common in the inland regions of East Greenland, where a more continental climate exists compared to the moister and cooler coastal regions (Wagner and Melles 2002).

The period between 1000–600 BC is characterized by a sharp decrease in *Salix* pollen percentages and low *Betula* values. Hereafter, plants typical of fell field vegetation develop again (e.g. Poaceae, Caryophylaceae), with the exception that *Salix* values are still relatively high. A short period between of 1050–1450 AD shows a weak return of birch and a sharper increase in willow (Wagner et al. 2000). The current landscape at the study sites is characterized by dwarf shrub heath dominated by *Salix glauca* and herbs like *Dryas*. At Basalt Sø, the vegetation is patchy, scattered between boulders and rock outcrops (Wagner et al. 2000).

3.2 Terrestrial records from West Greenland

Western Greenland, and more specifically, the area around Kangerlussuaq, comprises the largest ice-free part of Greenland. Due to the presence of the Kangerlussuaq airport and a road to the ice sheet, logistics are readily available in this region, and as a result of the easy accessibility a relatively high number of terrestrial records are available for this region (Figure 3-2a). As stated in Section 2.2, the coastal areas of West Greenland were ice-free several millennia earlier than the inland regions. This also results in differences between coastal records and inland records registering changes in temperature, productivity or aridity during the Holocene.

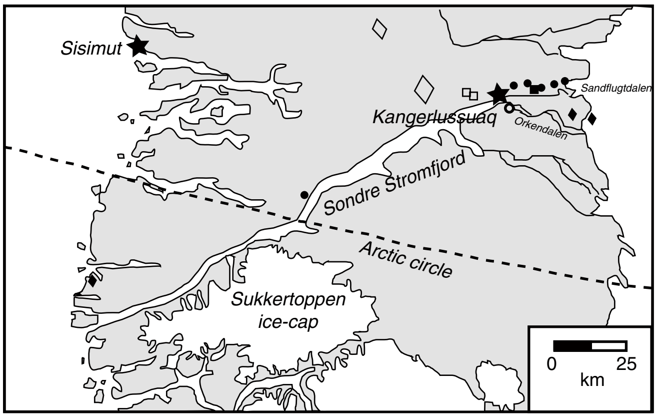

Figure 3-2a. ***Figure 3-2a.*** *Map of the Søndre Strømfjord, in the text referred to as the "Kangerlussuaq area", indicating the sites of lake sediment studies in this regio. Solid black circles indicate the locations of the sites studied by Willemse and Törnquist (1999); Store Saltsø (e.g. Bennike 2000) is indicated by an open circle; the open rectangles show the study-sites of Anderson and Leng (2004); the solid rectangle shows Lake31 from the Eisner et al. (1995) study; solid diamonds indicate the three lakes studied by Bindler et al. (2001a); open diamonds indicate the regions with the 7 lakes studied by Malmquist et al. (2003).*

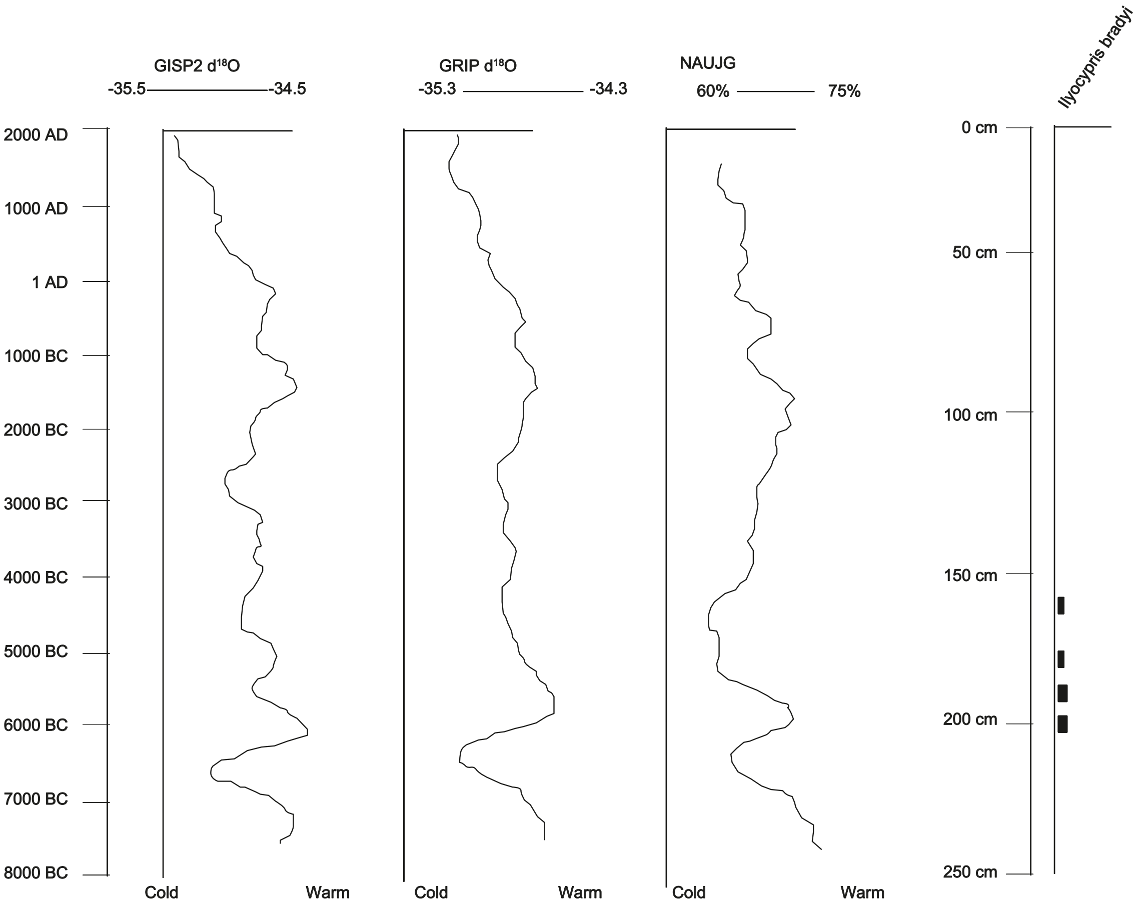

Figure 3-2b. *Smoothed oxygen isotope curves for the GISP2 and GRIP cores (100 yr bandwidth) in ‰ and one of the lake records of Willemse and Törnquist (1999; NAUJG) showing residue-on-ignition (%). All after Willemse and Törnquist (1999).*

Climate – air temperature

Willemse and Törnquist (1999) used Loss On Ignition (LOI) as an indicator for past changes in air temperature, reasoning that the extremely consistent results between their 6 study lakes can only be explained through an overriding influence of climate on in-lake processes. High temperatures are reconstructed between 8000 and 5500 BC, with a very distinct cold temperature anomaly visible around 6200 BC (or 8.2 cal ka BP; Figure 3-2b). This anomaly is also known from the Greenland ice-core records (Figure 3-2b and Section 3.5) and is commonly referred to as the 8.2-event (e.g. Alley et al. 1997).

After a sharp decline in reconstructed temperatures around 5500 BC, temperatures gradually increase, reaching a maximum shortly after 2000 BC. After 1500 BC, a downward trend is recognized in the LOI-curves (and thus in inferred relative temperatures) which is also visible in both the Greenland Ice Core Project (GRIP) and the Greenland Ice Sheet Project 2 (GISP2) ice-core records (Figure 3-2b). The climate fluctuations during the Late-Holocene as known from other locations in Greenland (i.e. the Medieval Warm Period (~500–1000 AD) and the Little Ice Age (~1500–1800 AD)) are absent from their record, although there is century-scaled variability present in the records of Willemse and Törnquist (1999).

A macro-remain based study on the lake sediments of the Store Saltsø shows that between 5000 and 4500 BC the ostracod *Ilyocypris bradyi* occurred in the region, whereas this thermophilous species is absent from Greenland at present (Bennike 2000). Therefore, Bennike (2000) concluded that this period must have been the warmest of the Holocene, with temperatures reaching values higher than the present temperature of the region.

Using another lacustrine record from the Kangerlussuaq region, Anderson et al. (1999) date the onset of the Neoglaciation (or the start of the decrease in temperatures) to 2000 BC, based on lithological features of the core.

Climate – precipitation/aridity

Anderson and Leng (2004) produced stable isotope records (^{18}O and ^{13}C) from two oligotrophic and saline lakes in the Kangerlussuaq region. Both lake records and both isotopes show strong variations, reflecting the changing precipitation/evaporation (P/E)-balance throughout the Holocene. It is concluded that the area was more arid than today, and climate was quite dynamic in the Early to Middle Holocene, more specifically between 5100 and 2300 BC. On the same lake records, McGowan et al. (2003) used diatoms to reconstruct conductivity, and they found a zone barren of diatoms that they link to rapid lake level lowering as the result of changes in the precipitation/evaporation-balance to a negative effective precipitation. The lake level lowering resulted in resuspension of glaciolacutrine clays and subsequent low productivity and poor preservation of diatoms in the sediments during the Middle Holocene.

Anderson and Leng (2004) proposed that the relationship between the Greenland high pressure area and the North Atlantic low (as described in Barlow et al. 1997) fluctuated more substantially prior to ~ 2300 BC than it does today, which resulted in alternating wet and dry periods in the Kangerlussuaq region.

Furthermore, van Tatenhove et al. (1996) suggest that the margin of the Greenland ice sheet was positioned 20 km further to the east (i.e. inland) around 4500 BC compared to today (see Section 2.4), being the minimum value for the Holocene period. This would have meant that the winds blowing from the ice sheet in the Kangerlussuaq region may have led to stronger cyclonic conditions, and the associated depression (the Baffin trough, which today parallels the outer coast (Anderson and Leng 2004)) created stronger ESE winds, and hence drier conditions at the head of the fjord. This would have resulted in lower precipitation in the region. The Early Holocene also had higher summer insolation values and increased winds, which could have led to higher evaporation. Together, this would have led to a negative P/E ratio, and lowering of the lake levels during the Early to Middle Holocene. Eisner et al. (1995) also reconstruct higher values of aeolian influx in their lake record between 4000 (the lowest sample of their record) and 2500 BC, supporting the theory of increased aridity and variability of climate during the Early to Middle-Holocene as proposed by Anderson and Leng (2004) and McGowan et al. (2002).

McGowan et al. (2003) proposed that their diatom-inferred increase in ionic concentrations is restricted to the inland area of Kangerlussaq, and that the general trend in lakes from West Greenland was one towards oligotrophication as indicated by Fredskild (1983, 1992). The area around Kangerlussuaq is currently still characterized by a negative P/E-balance, and many lakes are saline as a result of the prolonged concentration of ions in the lake-water.

Climate – eolian activity

Willemse et al. (2003) analysed the aeolian landforms that are encountered in the two floodplain systems near Kangerlussuaq: Sandflugtdalen and Ørkendalen (Figure 3-2a). The authors studied exposures that were available as a result of undercutting by the river and carried out ~400 hand corings, and thus reconstructed the history of aeolian activity in the region. Generally, the sedimentological sequences were shallow (up to a couple of meters depth) and started with a melt-out till, were then followed by poorly sorted sediment, and then by the first wide-spread aeolian deposit (I). Above this aeolian deposit, a peaty silt (II) was encountered that could also be followed throughout the region. Finally, a return to an aeolian facies was observed (III).

Using radiocarbon dates, a chronology was established. The chronology indicates that a long Middle to Late Holocene interval (2900–1750 BC) with high silt influx was present (=I), followed by an interval of subdued dust deposition (=II) bracketed by a series of distinct aeolian depositional phases at 1300–800 BC and 200 BC–150 AD. A renewed increase of silt deposition was reconstructed for the period between 350 and 1430 AD (=III), with ~250 yr-long recurring phases of increased silt deposition. Finally, a strong reduction in silt deposition was reconstructed after 1430 AD. The braided-river systems still form a source-area for wind-transported sediments, and less-frequently flooded parts of the floodplain show small-scale active dune fields.

The silt influx data demonstrate strong changes in intensity of aeolian activity during the last 5.000 years, which are tentatively linked to periods of changing winter aridity (Willemse et al. 2003).

Vegetation

Eisner et al. (1995) were the first to reconstruct vegetation and associated climate from pollen and macro-remains taken from a lacustrine core from the Kangerlussuaq area[1]. Their record starts at 3800 BC, when there already was a well established vegetation, dominated by *Betula* and Cyperaceae. This phase also shows high concentrations of *Chara* and *Tolypella* oospores in the macro-remain record, as well as maximum values for the aquatic plants *Myriophyllum spicatum* and *Potamogeton filiformis*.

Between 3000 and 1700 BC, Eisner et al. (1995) reconstruct a phase with higher pollen accumulation rates and higher *Salix*, Gramineae and Ericales abundances. A peak occurrence of *Alnus* pollen occurs at 2700 BC, which is interpreted as long-distance transported pollen of *Alnus*, which migrated to southwest Greenland between 4–3.5 ^{14}C ka BP (~4000 BC). The organic sediments that were deposited during this time-period are completely barren of Characeae oospores, but do contain macro-remains of several *Potamogeton* species and of chironomids, *Spongilla* and *Daphnia pulex*. The period between 3000–1700 BC is interpreted as a climatic optimum. Several of the *Potamogeton* species, the genus *Tolypella* and the encountered sponge (*Spongilla lacustris*) are currently recorded in locations south of Kangerlussuaq.

After 1700 BC, Eisner et al. (1995) show decreasing pollen percentages of *Betula* in favour of *Salix* and Graminae, and the total pollen accumulation rates steadily decrease. The uppermost 10 centimeters of sediment are barren of macro-remains of macrophytes except for oospores of *Nitella*.

[1] This is probably 'Two-tent lake' of the 2008 SKB field work campaign.

Pollutants (organic and metals) and sedimentation rates

The Arctic is considered to be a major focus for atmospheric pollutants that travel north from the mid-latitudinal industrial regions. Bindler et al. (2001a) studied the spatial variability in Pb-pollution over the last 200 years in the Kangerlussuaq region. Lead is an ideal indicator of pollution, as it is immobile in the sediments, has four different isotopes, is emitted from various sources (offering the potential to determine the major source regions) and is easy to detect and analyse. The Pb that accumulates in lake sediments is derived from anthropogenic sources as well as from natural sources in a lakes' catchment. Long records were derived from three lakes: one near the coast, one near the head of the fjord, and one on a nunatak on the ice sheet (Figure 3-2a). The upper parts of the cores were dated using ^{210}Pb-dating, and the results indicate that sedimentation was uniform over the last 130 years for the two coastal and inland lakes, with a mean accumulation rate of 0.053 cm/y (Bindler et al. 2001a). The Nunatak-lake had slower sedimentation rates, only reaching 0.020 cm/y (Bindler et al. 2001a). Surface sediment samples (the upper 5 mm of a gravity core) of an additional 14 lakes on a E-W transect were also used to get an overview of the spatial trend of pollution.

Evidence of pollution is shown in all three lake records, although they also all show that Pb-influx has declined in recent decades, proposedly due to improved emission controls. Lakes near the coast have received more Pb, which was expected given the higher rate of precipitation in this region (500–1,000 mm at the coast vs <150 mm at Kangerlussuaq). Unexpectedly, the ratios between the different Pb-isotopes suggest Eurasian sources (both European and Russian) as the major contributors of this pollution, instead of the American sources, which are situated closer to West Greenland.

Bindler et al. (2001b) studied the fluctuations in Hg concentrations in the same lake sequences and surface lake samples. All three sediment cores show a similar trend of low mercury concentrations in the lower parts with increasing values towards the sediment surface. The lower sediments show accumulation rates of Hg of about 1–3 ug/m2/y. The first significant increase is seen at Nunatak Lake at ~1700 AD (to values of 5 ug/m2/y), whereas the first increase at the other lakes is seen not until the mid 1,800 s, when the Hg concentrations increase 2- to 3-fold, reaching peak accumulation rates of 5–10 ug/ms/y (Bindler et al. 2001b). In two lakes, the uppermost sediments show a decrease in Hg concentrations, which corresponds to trends observed elsewhere in the Arctic. The current and the historical Hg accumulation rates are similar in Greenland and values reported from the Canadian Arctic (Bindler et al. 2001b).

Lindeberg et al. (2006) studied fluctuations of lead and mercury focusing on longer timescales in three lakes: Lake B (situated 50 km northwest of Kangerlussuaq, and providing a record back to a maximum age of 900 AD), Lake G (20 km northwest of Kangerlussuaq; 2800 BC) and Lake SS16 (10 km southeast of Kangerlussuaq; 6000 BC). The carbon content of these lakes fluctuates between 6–14%. The mean sedimentation rate in Lake B was calculated at 0.008 ± 0.002 g/cm^2/y. The records show that both Hg and Pb concentrations fluctuated during the whole time interval that was studied. Hg concentrations fluctuated between 40 and 110 ng/g (Lake B), 10 and 70 ng/g (Lake G) and 10 and 120 ng/g (Lake SS16). In the most recent parts (i.e. the last 200 years) increasing concentrations can be seen in all three cores. The range of the increase exceeds the natural variability in two of the cores, and only in the sediments from Lake G the abrupt increase in Hg concentrations falls within the range of variations observed in the deeper layers (Lindeberg et al. 2006). Pb concentrations fall between 0.5 and 1.9 ug/g for all three lakes in sediments predating the 19th century. In the recent sediments, the Pb concentrations increase 2- to 5-fold, reaching values between 2.9 and 8.7 ug/g. These values are lower than Pb concentrations reported from other high-arctic regions such as Canada, which is attributed to the low precipitation in this region.

Persistent organic pollutants (POP's) in the Arctic have been a cause for concern over the last two decades (Malmquist et al. 2003). Malmquist et al. (2003) used 10 1-cm thick samples from gravity cores from 7 lakes to establish the temporal trend in POP-deposition in the area around Kangerlussuaq (Figure 3-2a). Observed concentrations of POP's are low when compared to sediment records from more populated areas, and lower-than to comparable-to those of other Arctic records (Malmquist et al. 2003). Time-trends follow those of other records, with a highest influx around the 1950s (to 1980s), and there seems to be some delay in the POP-concentrations in Greenland when compared to more populated areas (Malmquist et al. 2003).

3.3 Terrestrial records from South Greenland

Climate – temperature and precipitation

The climate of southern Greenland is influenced by both cold air masses associated with the proximity of the ice sheet, as well as by warmer air masses related to the regional oceanographic circulation pattern (Kaplan et al. 2002). The southwestern coast of Greenland has the highest precipitation values of entire Greenland, (e.g. 1,300 mm at Lake Qipisarqo, Kaplan et al. 2002) as the result of the presence of the warm West Greenland Current.

South Greenland comprises the only known lacustrine records from Greenland covering the Late Glacial time-interval. Björck (pers. comm.) studied multiple lakes that include Late Glacial lacustrine sediments in the southernmost tip of Greenland, including a part of the Allerød and the Younger Dryas. Björck et al. (2002) employed a multi-proxy approach in order to reconstruct past climatic parameters from sediments of lake N14, which provides the only registration of Late Glacial climate conditions in Greenland that we are currently aware of.

The multi-proxy data of Björck et al. (2002) suggests that the period before the Younger Dryas was rather unstable and humid. In many European Late Glacial lake records, there is a distinct lithological change at the Allerød-Younger Dryas transition. Surprisingly, such a change is absent in the lake records from Southern Greenland. Instead, the sedimentary changes are subtle (Björck et al. 2002).

The pre-Holocene productivity of the lake was rather constant, possibly with higher in-lake productivity and less surface run-off in the catchment during the Younger Dryas (after 11,000 BC). The pH of the lake water increased during the Younger Dryas, which might have been the result of increased nutrient concentrations during the growing seasons. The high insolation during the Younger Dryas might have caused lake-levels to drop, thus increasing nutrient concentrations in the lake. Winters were probably cold and dry, which is deduced from the stable conditions in the lake record and the lack of vegetation development. Therefore, the Younger Dryas in southernmost Greenland was surprisingly characterized by anomalously mild summer temperatures and arid conditions (Björck et al. 2002).

After the Younger Dryas, again more unstable and humid conditions (similar to those preceding the Younger Dryas) are recorded at lake N14 (Björck et al. 2002). Small fluctuations in the records between 9100 and 8800 BC might point to a cooling corresponding to the Preboreal Oscillation.

Andresen et al. (2004) studied the Holocene part of the same lake record in a similar fashion, and the authors attempt to reconstruct past changes in precipitation. Biogenic silica in the sediments of lake N14 is used as the most important indicator of climatic changes for the period between 7300 BC and 1450 AD, and interpreted as such: increased precipitation led to an increase in surface run-off; this increased run-off carried additional nutrients to the lake, which increased diatom and chrysophyte production and thus biogenic silica content (Andresen et al. 2004). However, the increased production could also be the result of an increase in temperature; the problem with the latter explanation is that there are several episodes with increased abundances of moss in the record, and the changes in the abundance of these mosses can be explained by changes in light penetration (through algal blooms) but are less likely to have resulted from changes in temperature (Andresen et al. 2004).

Andresen et al. (2004) show increasing diatom production (from ~0 to 0.2 mg/cm^2/y) and organic-matter production (from ~5 to ~8 mg/cm^2/y) during the earliest Holocene (Figure 3-3), suggesting a transition toward warmer conditions with decreased summer run-off between 8700 and 8600 BC. Between 8600 and 7600 BC, increased warming and more humid conditions are reconstructed, which could have been the result of a northward shift of the Icelandic low to its present position (Andresen et al. 2004). To the northwest, Kaplan et al. (2002) studied a lake that was isolated from the sea around 7100 BC, and for the first 3 millennia this lake shows a progressive warming. In general, the Early Holocene data shows a transition from a dry and cool climate to a more humid and mild climate, which is also suggested by the Greenland ice-core records (Johnsen et al. 2001).

High levels of biogenic silica of up to 25 % of the sediment-weight are encountered in the sediments dated to 6000–3000 BC at lake N14 (Andresen et al. 2004). Maximum values are reached between 6000–4500 BC, which is therefore interpreted to be the warmest and wettest period in the record (corresponding to the Holocene climatic optimum; Andresen et al. 2004). Kaplan et al. (2002) reconstruct increasing temperatures during 7000–4000 BC, and stable warm conditions between 4000–1000 BC.

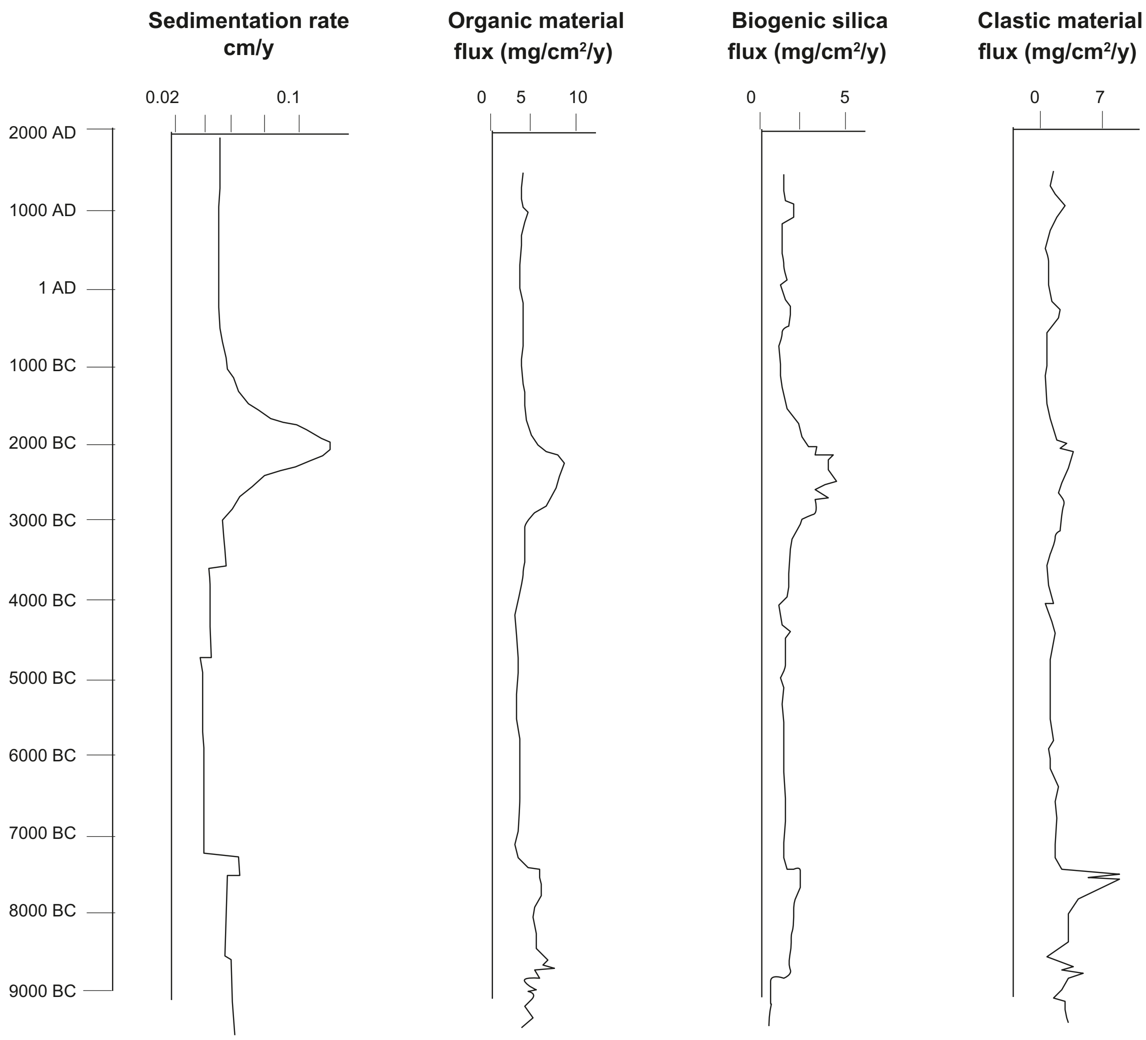

Figure 3-3. *Changes in sedimentation rate, and flux rates of organic material, biogenic silica and clastic material, reconstructed based on the Holocene lacustrine record of Lake N14. After Andresen et al. (2004).*

Between 2700 and 2000 BC, there are peaks in sedimentation rates of organic matter (up to 10 mg/cm^2/y), biogenic silica (up to 5 mg/cm^2/y) and clastic material (up to 6 mg/cm^2/y), and of sulphur, in the record of lake N14. Concordantly, moss-index (a qualitative measure of the amount of moss in the sediment) shows highly fluctuating values. Altogether, this is interpreted to indicate a late Middle Holocene deterioration of climate conditions (colder conditions), also known as the beginning of the Neoglaciation. The onset of the Neoglaciation in lake Qipisarqo, located 60 km west of Cape Farewell (South Greenland), is dated to ca 1000 BC, where a marked cooling is inferred from that point onward (Kaplan et al. 2002).

After 1700 BC, the climate changes towards a generally drier state, and particularly the periods between 1700–400 BC and 150–1350 AD are characterized by dry conditions in the record of lake N14 (Andresen et al. 2004). Another record from South Greenland, situated several 100 km northwest of Lake N14, also shows dynamic and unstable conditions in the period between 1000 BC–present with an increasing instability in the recent past. The general trend in the latter record is toward cooling, which culminated in the Little Ice Age. However, two reversals are witnessed at 650–1050 AD and 1450–1670 AD, of which the earlier one corresponds to the Medieval Warm Period (Kaplan et al. 2002). The Neoglaciation cooling in the interval between 1000–50 BC is pronounced in both records, as well as in Baffin Island records and east Greenland, showing a teleconnection involving decreased ocean heat transport (Kaplan et al. 2002). In the Qipisarqo record, an increased warming is reconstructed over the last 45 years (Kaplan et al. 2002).

Vegetation (Late Glacial)

The period between 10,900 and 9500 BC is characterized by pollen assemblages that include Poaceae, Caryophyllaceae (including *Sagina*-type), Chenopodiaceae and *Saxifraga caespitosa*-type and the trend is towards a gradual establishment of higher plants (Björck et al. 2002). *S. caespitosa* might indicate arid conditions, which is in contrast with the more wet-demanding *S. stellaris* that is found before 10,800 BC and after 9500 BC (Björck et al. 2002). At 11,200 BC, more than 50% of the pollen assemblage is made up of the Caryophyllaceae *Sagina*-type, indicating its local presence. At 9500 BC, pollen of *Empetrum nigrum* and *Vaccinium* are present in the record, indicating their presence on Greenland (Björck et al. 2002). The diatom flora is dominated by a few pioneer species during the Late Glacial interval, but increasing values of Achnantes-taxa might point to increasing pH (Björck et al. 2002).

Human Settlement

The south of Greenland was colonized by the Norse under the wing of Eric the Red at 985 AD. This is close to the peak of Medieval Warm Period as reconstructed in the GISP2 ice-core (Stuiver et al. 1995) and in lacustrine records of the South Greenland region (see above). This period probably had favorable sailing conditions (open water on the oceans; Lassen et al. 2004). Although it is not entirely clear when the last Norse settlements in South Greenland were abandoned (the so-called Eastern Settlement), it is very likely that their disappearance coincides with the Little Ice Age climate deterioration (Lassen et al. 2004).

3.4 Terrestrial records from North Greenland

Climate and vegetation

Although rare, there are a few lacustrine records from the high arctic biozone of northern Greenland. Bennike and Weidick (2001) published one record of macro-remains, organic carbon and carbonates. The organic content of the lake sediments fluctuates between 5–20%, and the carbonate contents between 5–40%, as determined by loss-on-ignition. The total core length is 120 cm, and the lacustrine sediment sequence starts at 6800 BC.

One of the problems that Bennike and Weidick (2001) encountered is the fact that the samples only have very few remains in them or are completely devoid of macro-remains. They interpret these samples as representing stages where the lake was completely frozen all year round, as is also the case for many lakes in the region at present. The samples that do contain macro-remains reflect a desert-like landscape, with varying amounts of chironomidae, chydorus etc., which might reflect changes in ice-free periods. Samples covering the period from 8000 to 6800 BC suggest summer temperatures higher than today. A Middle Holocene warm period is inferred for 7000–2000 BC, although there might have been a migratory lack, which means that the “real” warm period might have started earlier than the date reported by Bennike and Weidick (2001).

From the northwestern part of Greenland, hardly any information on Holocene climate evolution is available. Bennike (2002) dated shells encountered in lateral moraines of the Humboldt Gletscher on Washington Land. The dates indicate that the glacier was smaller than present during ca 5000–2000 BC, and the front might have been 25 km further from the coast compared to the present position.

3.5 Ice-core records

The Greenland ice-core records, and especially the oxygen isotope records, are amongst the most well-known natural archives of past climate change. By now, results of seven individual deep ice-cores have been published (Figure 3-4a), and the resolution and chronology of these records (especially those from central Greenland) surpass any other type of records that is available from the North-Atlantic region at the moment.

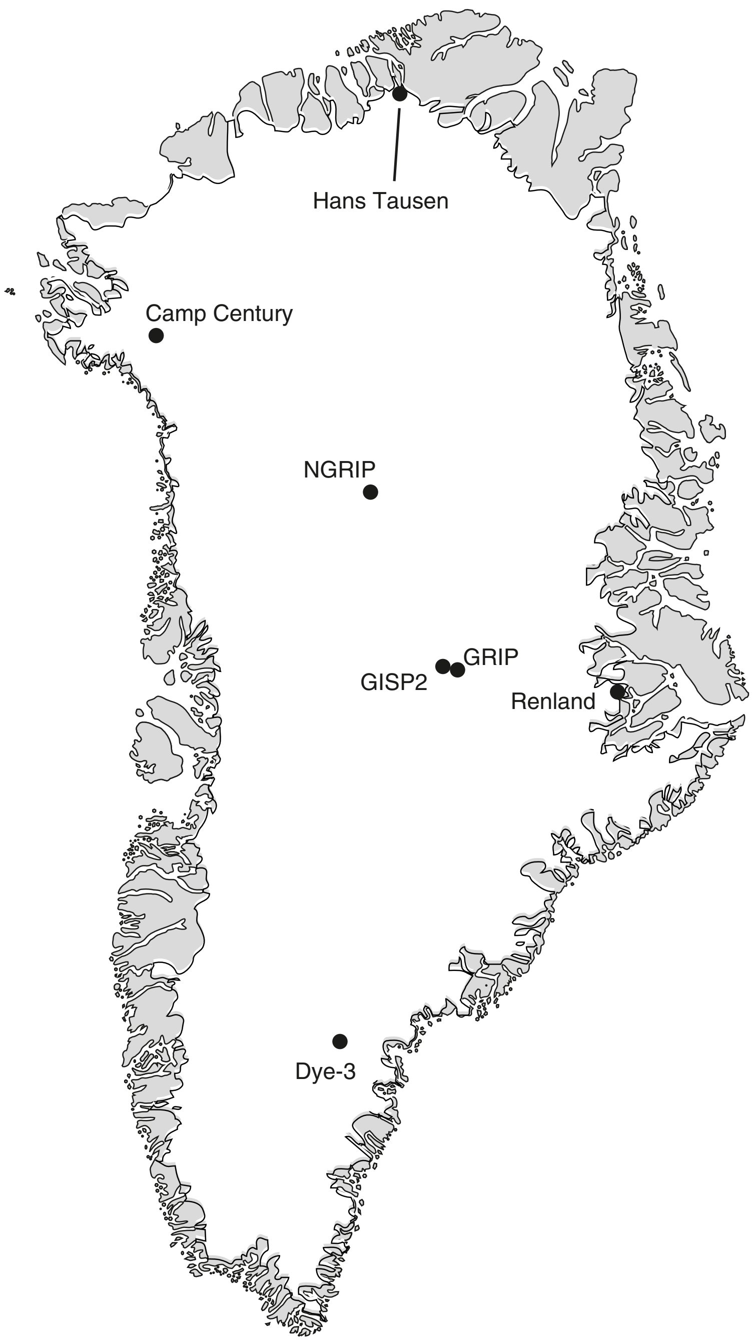

__Figure 3-4a.__ The 7 deep drilling sites in Greenland, after Johnsen et al. (2001).

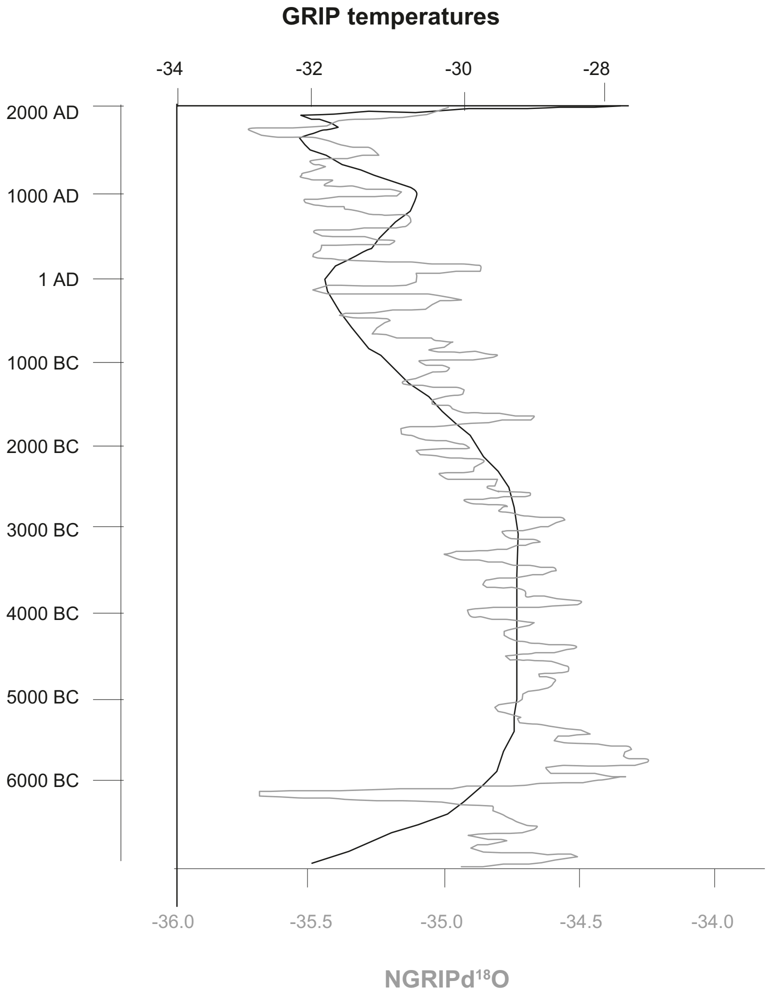

Figure 3-4b. *The NGRIP $\delta^{18}O$ record with a 200y Gaussian filter superimposed on the GRIP temperature profile calculated with the Monte Carlo inversion technique. After Johnsen et al. (2001).*

The oldest records start in the Eemian, although it is assumed that most of the Eemian ice has been subject to displacement. Using a method based on borehole temperatures and calibration against modern $\delta^{18}O$ values, Johnsen et al. (2001) calculated palaeotemperatures from the oxygen-isotope ratios of the GRIP core. It suggests that Eemian (annual) temperatures over central Greenland were approximately 5°C higher than present-day temperatures.

The Last Glacial Period (110–11.5 ka BP) is characterized by rapid climate oscillations (e.g. Dansgaard et al. 1982), which are now commonly referred to as the Dansgaard/Oeschger events. These abrupt and high amplitude climate oscillations typically start with a sharp increase in temperature of ~10–15°C. A period of gradually decreasing temperatures follows (in the order of several hundred-few thousand years) and the event is terminated by a sharp return to cold conditions. There seems to be a periodicity of ~1.5 ka to these events, and possible explanations or driving mechanisms include reorganizations of the ocean thermohaline circulation, possibly following periodic surging of northern hemisphere ice sheets. During the Last Glacial Maximum, temperatures were ~20°C colder than today, and GCM modeling suggests that hardly any precipitation fell over the central parts of the ice sheet during winter months (Werner et al. 2000 cited in Johnsen et al. 2001).

The Holocene has been studied in greater detail, and using only the oxygen-isotopes as temperature proxy, a general trend of decreasing temperatures towards the present (max difference is ~ 3°C) is detected (e.g. Figure 3-2b), with three cooling events at 9300 (the PBO), 7300 and 6200 (the 8.2-event) BC. However, the record based on Monte Carlo inverse modeling shows a temperature record that is much more dynamic (Figure 3-4b), and closely resembles the temperature records derived from lake sediments, especially those from South Greenland (Dahl-Jensen et al. 1998, Johnsen et al. 2001).

A thermal maximum is reconstructed between 6000–3000 BC (Dahl-Jensen et al. 1998), after which a decline in temperatures can be seen, culminating in a cold minimum around 50 BC. Hereafter, a Medieval Warm Period is reconstructed (around 1000 AD; Dahl-Jensen et al. 1998) with temperatures ~1°C warmer than present, followed by two cold-oscillations that correspond to the Little Ice age (around 1550 and 1850 AD) with temperatures 0.5–0.7°C below present-day values. This trend is reconstructed for both the GRIP (central Greenland) and Dye-3 (South Greenland) cores independently, and the temperature history for both records is very similar (Johnsen et al. 2001). The biggest difference is the amplitude in reconstructed changes in temperature, which is higher in South Greenland (Dahl-Jensen et al. 1998, Johnsen et al. 2001). During the last few decades, a decrease in temperatures is reconstructed (and measured). However, the reported temperatures are less reliable further back in time (Johnsen et al. 2001) and the variations of temperature, especially during the Late Holocene, are small compared to the expected errors arising from the conversion of oxygen isotope values to temperature estimates.

3.6 Records from the marine realm

South

Lassen et al. (2004) published a marine record from South Greenland that shows a similar Late Holocene climate signature compared to the lacustrine record. They used a core from a fjord where there are currently two water-masses overlying each other (a lower, warmer one belonging to the Irminger Current, and a colder upper one belonging to the East Greenland Current). They reconstruct increased mixing during the Medieval Warm Period between 885–1235 AD (as a consequence of increasing wind stress). The transition to the Little Ice Age is stepwise with a short climate amelioration at 1520 AD before the maximum cooling occurs. Lassen et al. (2004) suggest that solar activity in interplay with variations in the ocean thermohaline circulation might have played an important role as forcing factors for the climate variability in the Late Holocene.

East

In general, most records obtained from marine sediments in East Greenland show an Early-Middle Holocene climate optimum between 6900–3800 BC, that is followed by a cooling (e.g. Koç et al. 1993, Nam 1997). These climatic changes correspond to changes in the thermohaline circulation (Wagner et al. 2000 and references therein).

4 Training sets and the transfer-function approach

4.1 General

During the last two decades, new approaches have become available to reconstruct past changes in environment and climate from a range of different organisms. One of the most important developments in palaeolimnology has been the development of different transfer functions, mathematical multi-variate formulas that can be used to infer quantitative estimates of an environmental variable of interest from an assemblage of organisms (e.g. Birks 1998, Brooks 2006). These transfer functions rely on information of the modern relationship between the group of organisms and their environment. This information is derived from so-called training sets, which are basically a suite of lakes, peat cores or any other type of natural archives distributed over a gradient of the environmental variable of interest.

To illustrate this, let's look at an example: If one wants to reconstruct palaeoproductivity (as total phosphorous (TP) or total nitrate (TN)) of a lake, one can use fossil diatom assemblages as a proxy. To be able to use them in a quantitative approach, the modern relationship between diatom taxa and productivity needs to be established. To do this, a group of lakes with very low to very high TP and TN values is sampled, which means that TP and TN are measured at the sites (amongst other environmental variables) and that the surface lake sediments are cored, in order to established which species occur in the lake, and to determine their relative abundances. This way, two datasets are obtained: the first dataset containing information on the environmental data per lake, the second dataset containing information of the diatom-assemblages per lake. These two datasets are together called the training set. Using a transfer function (TF; the above mentioned multivariate formula) the relationship between the two datasets can be modeled. This TF essentially describes the relationship between the diatom assemblages of the modern lakes and the TP/TN. Applying this same TF to the fossil diatom assemblages, we can infer past values of TP and TN from them (Table 4-1).

Table 4-1. Table a) shows the two datasets retrieved when building a training set: the first containing environmental information for a suite of different lakes (in this case including total phosphorous (TP) and total nitrate (TN) of the lake water); the second containing information of the relative abundance of the selected proxy per lake (in this case diatoms). Both of the tables under a) contain known values, and using this information a transfer function (TF; indicated by the red arrow) can be developed. Table b) shows the application of the same transfer function, developed using the information in a) to a fossil set of samples. In this case, only the table on the right hand side contains known values, and the table to the left hand side is unknown.

Training set

	Lake1	Lake2	Lake3	Lake4	Etc			Lake1	Lake2	Lake3	Lake4	Etc
TP	…	…	…	…	…		Diatom1	…%	…%	…%	…%	…
TN	…	…	…	…	…	⟵	Diatom2	…%	…%	…%	…%	…
Env3	…	…	…	…	…	TF	Diatom3	…%	…%	…%	…%	…
Env4	…	…	…	…	…		Diatom4	…%	…%	…%	…%	…
Etc	…	…	…	…	…		Etc	…%	…%	…%	…%	…

Fossil dataset

	Sample 0–1 cm	Sample 1–2 cm	Sample 2–3 cm	Sample 3–4 cm	Sample Etc			Sample 0–1 cm	Sample 1–2 cm	Sample 2–3 cm	Sample 3–4 cm	Sample Etc
TP	???	???	???	???	???		Diatom1	…%	…%	…%	…%	…%
TN	???	???	???	???	???	⟵	Diatom2	…%	…%	…%	…%	…%
						TF	Diatom3	…%	…%	…%	…%	…%
							Diatom4	…%	…%	…%	…%	…%
							Etc	…%	…%	…%	…%	…%

Several aspects are important when assessing the applicability of a training set and/or a transfer function: 1) Training sets that are developed in order to reconstruct climatic variables should ideally cover a large gradient over the climatic variable of interest. For instance, if one wants to reconstruct summer temperatures in Greenland, an ideal training set would incorporate a high number of lakes that are equally spread over a region where the coldest summer temperatures lie around 3–5°C and the highest around 17°C. This way, the optimal temperatures of the different organisms are determined most reliably; 2) The number of species incorporated in the training set should be high. This way, the assemblages included in the training set are more likely to be suitable modern-analogues of fossil samples encountered in the lake sediment sequences. This is often reached by including a lot of sites in the training set; 3) The lakes in the training set should have different geological or biological settings, but the environmental variable of interest should be the overriding parameter in determining the flora or fauna at the site. This means that it is best, for instance, to have both alkaline and acidic lakes in a training set that is developed to reconstruct TP from diatoms. If alkalinity changes throughout your fossil record, your reconstruction will not be negatively influenced by such a change.

There is a range of other (minor) aspects that one has to bear in mind when determining the suitability of a training set or a transfer function, but luckily a couple of different numerical (statistical) tools exist to determine this objectively (Birks 1998).

4.2 Training set development in Greenland

Region

Unfortunately, the before mentioned large climatic gradients that one would like to include in a training set do not occur in Greenland within a limited space. The region around Kangerlussuaq, however, has some potential, as the distance from the outer coast to the western margin of the Greenland ice sheet is around 175 km (e.g. Bennike 2000), and a relatively strong climatic gradient can be seen. The climate at the outer coast is maritime, due to the proximity of the sea, with a mean July temperature of 6.3°C and an average annual precipitation of 380 mm. A more continental type of climate is found near the margin of the ice sheet, because of a high-pressure system that is normally present over the ice sheet, and the mean July air temperature is ~10.6°C and mean annual precipitation is ~150 mm, (Nielsen 2008). In terms of other interesting environmental parameters (such as conductivity of lake-waters, which is often used as a proxy for the precipitation: evaporation ratio), there are large gradients within small regions.

Between 1996 and 2000, over 100 lakes have been sampled by Anderson and co-workers in the Kangerlussuaq area, covering the land between the outer coast and the ice sheet (e.g. Ryves et al. 2002). Over 28 environmental parameters were measured for all these lakes. Using automatic data loggers, high-resolution temperature records were obtained for a subset of the lakes. 40 of the lakes were cored and analyzed for diatoms (Ryves et al. 2002); 47 of the lakes were analyzed for chironomids (Brodersen and Anderson 2002). Below, we discuss both training sets and their potentials and problems individually.

Diatoms

There is some diversity in the salinity of lakes in the Kangerlussuaq area, ranging from dilute (conductivity <30 uS/cm) to subsaline (conductivity ~4,000 uS/cm). Because this parameter is often interpreted as a proxy for precipitation/evaporation ratio, application of a transfer function describing the relationship of salinity and a group of organisms might provide a tool to infer past changes in precipitation, which is an important part of the climate system.

Ryves et al. (2002) analyzed the influence of conductivity and 27 other chemical and limnological parameters of a subset of 40 lakes on the diatom assemblages encountered in these lakes. The main variation in the diatom assemblages (12% of the species variance) can be explained by conductivity, even though the range of conductivity is (in absolute terms) not that high, and even the most saline lakes do not contain a truly saline diatom flora (Ryves et al. 2002).

Using a weighted-averaging partial least squares (WA-PLS) model (Birks 1995, 1998), the relationship between diatom assemblages and conductivity is described quantitatively, and the model shows performance statistics that are comparable to other developed inference models (Ryves et al. 2002). The model has been applied to a short core from Braya Sø, located 40 km west of the current ice-margin and just north of Søndre Strømflord, and although there were some dissolution problems, the results indicate fluctuating conductivity-levels throughout the sedimentation history, which are interpreted as changes in the P/E-ratio.

Chironomids

Because of the lack of meteorological stations throughout Greenland, it has been more difficult to establish the relationship between (summer) temperature and different proxies. Brodersen and Anderson (2002) used automatic loggers in 21 lakes, which continuously measured temperatures for these sites for a period of 3 months during the summer season. Using formulas that take into account height-differences and distance to the coast and the ice sheet, the results are interpolated for the other 26 lakes used in this study, resulting in a training set consisting of 47 lakes and including temperatures between 7.3 and 16.5°C (Brodersen and Anderson 2002).

A total of only 24 taxa were recorded in the training set, and using different exploratory techniques, a strong correlation between the trophic variables (TP and TN) and temperatures was shown. Both parameters explain an almost equal amount of variability in the chironomid data, but when TN was partialled out in a redundancy analysis, temperature lost its significant explanatory power. Therefore, a lake classification system was developed using a different technique (called two-way indicator species analysis or Twinspan), where temperature, the trophic variables, salinity and lake-morphometry were used as defining parameters. Using this method, 14 chironomid taxa showed significant percentage abundance difference for the different groups.

This indicates that there is potential for the application of chironomids as a palaeoenvironmental indicator in both short-term and long-term studies of lake ontogeny and palaeoclimatological conditions (Brodersen and Anderson 2002). However, because of the strong relationship between temperature and the trophic variables, a quantitative reconstruction of lake type and habitat type is recommended, in combination with direct reconstruction of single climatic variables such as temperature (Brodersen and Anderson 2002).

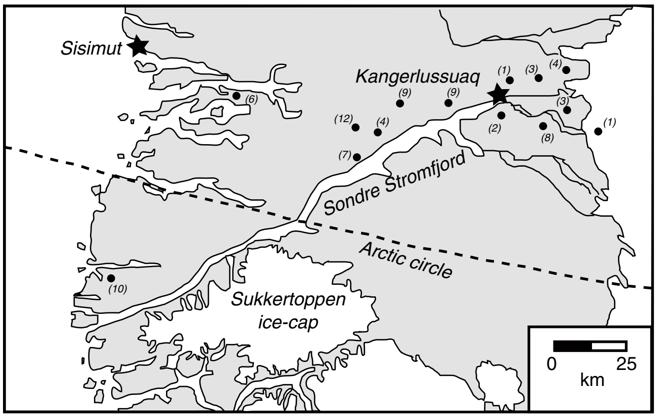

Figure 4-1. *Lakes for which surface samples have been collected (after e.g. Anderson et al. 1999). The number between bracket indicates the number of lakes sampled at that locality.*

5 The period directly after deglaciation

5.1 Terrestrial plants and animals

All over Greenland, the general terrestrial vegetation development after deglaciation starts with a similar face: the vegetation that spread over the newly ice free land was one of "fell field" (scattered herbs). Fell field vegetation is adapted to unstable soils, a high amount of soil erosion, and problems concerning the water balance of soils (as a result of underlying permafrost; Funder and Fredskild 1989). The vegetation cover is sparse and characterized by high abundances of grasses, Caryophyllaceae and Rosaceae.

Among these herbs were some thermophilous low arctic species, indicating summer temperatures slightly higher than today (Funder 1989). The species-richness in the fell field vegetation differed between the various regions of Greenland, but was generally considered rich, although exceptions are reported for e.g. northwest and central east Greenland. Because of the richness of the fell-field vegetation it is argued that vertebrates had already migrated to Greenland (they are considered to have been absent during the LGM) during the Early Holocene (Bennike 1997).

Bennike et al. (1999) show that only 13 vascular plant taxa were present in East Greenland during the first 500 years after deglaciation, (though probably several encompass multiple species), most of which have a current distribution with preferences for arctic or alpine environments. They are also characteristic of unstable soils (Bennike et al. 1999), which might be an important observation for other regions. It is still a question whether these species survived the coldest periods of the last glacial cycle in Greenland, or if they migrated to this region after the last deglaciation.

Remarkably, no shrubs or dwarf shrubs were yet present, which might have been the result of delayed migration, soil formation and hydrological constraints. It is argued (Funder and Fredskild 1989, Bennike 2000) that climate was not a limiting factor, as mean summer temperatures were high enough to ensure the presence of thermophilous plants. Woody plants arrived later in many of the records, and it is suggested that they all migrated to Greenland during the Holocene. *Empetrum nigrum* was the first dwarf shrub to arrive in east Greenland. This species is not cold-resistent and does not grow in northern Greenland at present (Bennike et al. 1999). The next plants to migrate to this region were *Salix herbacea* and *Dryas octopetala*. By 6000 BC, a major part of the Greenland vascular plant flora of today were already present.

In the Kangerlussuaq-region of Southwest Greenland, two lakes near the coast and two lakes inland have been analysed, and the results exemplify that the timing of deglaciation is important with respect to a possibly delayed response of the flora (Bennike 2000). Where the westernmost sites were deglaciated around 9500 BC, the pioneer phase (without any woody plants) lasted several millennia at these sites. The inland sites were deglaciated around 5000 BC, and here the pioneer phase only lasted several centuries, possibly to the fact that a number of dwarf shrubs had already established themselves well in Greenland by this time (Bennike 2000).

5.2 Aquatic plants and animals (lacustrine)

Bennike et al. (2002) state that few studies have been performed on the colonization of lakes by macrolimnophytes and invertebrates following isolation from the sea. In one lake (N14) from south Greenland, the cladoceran *Chydorus arcticus* and several chironomid taxa were amongst the first colonizers. This was later followed by cladoceran *Daphnia pulex* and aquatic bryophyte *Warnstorfia exannulata*. In another lake (N18), the diversity is much lower, and only a single freshwater organism (*D. pulex*) is present right after the isolation of the basin (Bennike et al. 2002). Other lakes in this region that got isolated from marine influence during the Early Holocene show a much higher diversity of both freshwater plants and animals, including a fair number of thermophilious species (Bennike et al. 2002). The first two lakes might have been colonized from lakes in northwestern Europe and North America, following transoceanic dispersal. The other lakes might later have been colonized by the already-established taxa in the region (Bennike et al. 2002).

Hjort Sø in northeast Greenland shows sediments deposited right after deglaciation that are devoid of any limnic organisms, which suggests that the water in the lake was too turbid because of suspended clays to allow limnophytes to live in the lake (Wagner et al. 2008). The only macro-remains that were present in these sediments were seeds of *Papaver radicatum* and moss-fragments of *Distichium* sp. which both have been interpreted as pioneer species, able to grow on unstable soils (Bennike et al. 1999, Wagner et al. 2008). After this period, an initial fell field vegetation is reconstructed (similar to other locations in Greenland) and early limnic animals include *Daphnia pulex* and several chironomid taxa (Wagner et al. 2008).

6 Summary and concluding remarks

This report discusses the Holocene climatic and environmental changes in Greenland, with particular emphasis on the ice sheet history. Although data is scarce for many regions and time intervals, some summarizing remarks can be made with regard to the ice sheet movements in the latter part of the Last Glacial period and in the Holocene:

- After the Last Glacial Maximum, the Greenland ice sheet started retreating between 13,000–11,000 BC. However, only the southernmost tip of Greenland was ice-free before the onset of the Holocene (at ~9500 BC).
- In East Greenland, it seems that the central part was deglaciated earlier than the northern and southern regions. However, this reconstructed variability in timing of deglaciation might be the result of the low number of sites that have been studied so far.
- In West Greenland, there is a clear signal of earlier deglaciation in the coastal regions as compared to the inland regions. There is a lot of spatial variation in the deglaciation history of West Greenland, which might be the result of differences in shelf width and local topography.
- A retreat of the ice sheet margin followed during the Early Holocene, and for many localities it is suggested that the ice sheet retreated to positions that are several tens of kilometers land inward from the current ice margin. After ~2500–2000 BC (though spatially variable), the ice sheet started to readvance again, a period referred to as the Neoglaciation. The period of ice build-up culminated in the Little Ice Age, at ~1850 AD. Although there is limited data available, it seems there is a lot of local and regional variation on this general picture with local advances or still stand phases.

Climate reconstructions are more abundantly available for many regions, as lake records have been analysed for a variety of proxies in many different regions. Although there is often some ambiguity in separating the effects of climate from migration patterns, the following can be concluded:

- Reconstructed temperature records over Greenland seem to be similar: the earlier parts of the Holocene (Table 1-1) show rising temperatures, culminating in a thermal maximum during the Middle Holocene. Hereafter, a more gradual decline in temperatures is reconstructed, culminating in a long cold period between 1000 BC and 1850 AD. In many records, a short-lived temperature increase is seen around 1000 AD.
- There are a few reconstructions of (effective) precipitation available, based on aeolian sediments encountered in lake sequences or on fossil pollen- and macro-remains. Alternating phases of wet and dry conditions are reconstructed for the Kangerlussuaq region for the period prior to ~2300 BC.
- Vegetation development shows strong spatial variance. Especially in West Greenland, there were big differences in the timing of immigration of higher vegetation: pioneer vegetation was present on the recently deglaciated terrain at coastal sites for several millennia, whereas at inland sites this phase was much shorter, in the order of several centuries. Vegetation trends in general follow the reconstructed temperature evolution (see above), although some circular reasoning cannot be avoided here (as changes in temperature are often deduced from pollen- or macro-remain evidence).
- A fell field vegetation established at all sites over Greenland after deglaciation. This vegetation type is characterized by relatively high abundances of grasses and herbs, and indicates a patchy vegetation on unstable soils. Some higher plants show clear migration lags: for instance, *Betula nana* is known to have arrived in East Greenland at 6900 BC, and its sudden high abundances in Holocene records are interpreted as a delayed response rather than as a climatic shift. In contrast, *Salix arctica* spread during the Middle Holocene, and this might actually have been a response to climate change.
- Above 80°N and on higher elevations, the vegetation consists of prostrate dwarf-shrubs, lichens and mosses. Washington Land is an example of a region with an extremely low biodiversity, with only 60-odd vascular plant species. Further south, the vegetation grades into a hemi-prostrate to low dwarf shrub vegetation community. The inland regions of west Greenland (including the Kangerlussuaq area) are currently characterized by a low-shrub community, and the southernmost tip of the continent has a subarctic vegetation-type (see Nielsen 2008).

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