Holocene environmental history of East Greenland - evidence from lake sediments

Seesedimente als Archive der Holozänen Umweltgeschichte Ostgrönlands

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„Research into the climates of the past shows that climatic change is, not surprisingly, never ending.“

E.C. Pielou, 1991

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Abstract

Sediment sequences from five different lakes in three different regions along the coast of East Greenland were studied for chronolgy, lithology, fossils, palynology, and biogeochemistry. The lakes Basaltsø and B1 are located on Geographical Society Ø at the outer coast of East Greenland, the lakes Noa Sø and N1 on Ymer Ø in the more interior region, and Raffles Sø on Raffles Ø in front of Jameson Land north of the Scoresby Sund mouth. The multi-disciplinary studies on the lake sediments were complemented by radiocarbon dating of marine bivalves from the Noa Sø surrounding.

The occurrence of glacial sediments at the base of three sediment sequences from Geographical Society Ø, Ymer Ø and Raffles Ø reflects the last glaciation of the investigated areas, probably during the Milne Land stade between 11,300 and 11,150 cal. yr B.P. Deglaciation was associated with a high accumulation of coarse grained terrigenous matter in all sediment sequences. The onset of more biogenic accumulation in the overlaying sediments was recorded by ages of terrestrial plant remains and marine bivalves, AMS radiocarbon dated to about 10,000 cal. yr B.P. Whilst the terrestrial plant remains and the larger amount of autochthonous organic matter in the sediments point to the establishment of bioproductivity in the lakes and their surroundings, the bivalves indicate a postglacial marine transgression that inundated the lower coastal areas.

The combination of the different investigated parameters and the comparison between the records from the individual sites indicate an early Holocene climatic optimum that lasted from ca. 9000 to 6500 cal. yr B.P. in the study area. The marine transgression reached at least 60 m a.s.l. on western Ymer Ø during the first half of this period, followed by a slow and constant regression. From ca. 7500 cal. yr B.P. the East Greenland Current weakened and thus caused at least two months of partly open-water conditions offshore during summer. This is shown by a remarkable nutrient enrichment in the sediments of Raffles Sø, interpreted as a result of settling seabird colonies in the catchment. Increasing snow accumulation in the outer coastal areas from ca. 6500 cal. yr B.P. reflects the onset of climatic deterioration that was supported by decreasing temperatures between ca. 5000 and 3000 cal. yr B.P. Cold and dry conditions lasted in East Greenland from 3000 cal. yr B.P. until 1000 cal. yr B.P. This period coincided with a strengthening of the East Greenland Current, which led to widely perennial sea-ice cover off East Greenland and thus to a decline of the seabird breeding population on Raffles Ø. A short warming in East Greenland between 1000 and 500 cal. yr B.P. was followed by the Little Ice Age from 500 to 100 cal. yr B.P., mirrored in lowest contents of organic matter in the sediments, and also in the vegetation assemblages in the lake surroundings.

Kurzfassung

Sedimentsequenzen von fünf verschiedenen Seen aus drei verschiedenen Regionen der Küste Ostgrönlands wurden auf ihre Chronologie, Lithologie, Fossilien, Palynologie und Biogeochemie hin untersucht. Die Seen Basaltsø und B1 liegen auf der Geographical Society Ø an der äußeren Küste Ostgrönlands, die Seen Noa Sø und N1 auf Ymer Ø in der inneren Region und Raffles Sø auf Raffles Ø vor dem Jameson Land nördlich der Scoresby Sund Mündung. Die multi-disziplinären Untersuchungen an den Seesedimenten wurden durch Radiokohlenstoffdatierungen an marinen Muscheln aus der Umgebung des Noa Sø ergänzt.

Das Vorkommen von glazialen Sedimenten an der Basis dreier Sedimentsequenzen von Geographical Society Ø, Ymer Ø und Raffles Ø spiegelt die letzte Vereisung der untersuchten Gebiete wider, wahrscheinlich während des Milne Land stade zwischen 11 300 und 11 150 cal. yr B.P. Der Eisrückzug war mit einer hohen Akkumulation von grobkörnigem terrigenem Material in allen untersuchten Sedimentsequenzen verbunden. Der Beginn erhöhter Bioakkumulation in den darüber liegenden Sedimenten ist durch die Alter von terrestrischen Pflanzenresten und marinen Muscheln bezeugt, die mittels AMS ^{14}C-Methode auf ca. 10 000 cal. yr B.P. datiert wurden. Während die terrestrischen Pflanzenreste und die höheren Anteile an autochthonem organischem Material auf einen Anstieg der Bioproduktivität in den Seen und ihrer Umgebung deuten, weisen die Muscheln auf eine postglaziale marine Transgression, die die niederen Küstenbereiche erfaßte.

Die Kombination aus den unterschiedlichen Untersuchungsparametern und der Vergleich der Rekorde zwischen den einzelnen Regionen zeigt ein frühholozänes Klimaoptimum zwischen 9000 und 6500 cal. yr B.P. im Untersuchungsgebiet an. Die marine Transgression erreichte mindestens 60 müNN auf der westlichen Ymer Ø während der ersten Hälfte dieser Periode, gefolgt von einer langsamen und konstanten Regression. Seit ca. 7500 cal. yr B.P. schwächte sich der Ostgrönlandstrom ab und verursachte so mindestens zwei Monate partiell offene Wasserbedingungen auf dem Meer im Sommer. Dies wird durch eine bemerkenswerte Nährstoffanreicherung in den Raffles Sø Sedimenten angezeigt, die als Ergebnis einer Ansiedlung von Seevögelbrutkolonien im Einzugsgebiet interpretiert wird. Ein Anstieg der Schneeakkumulation in der äußeren Küstenregion seit ca. 6500 cal. yr B.P. weist auf den Beginn einer Klimaverschlechterung, die durch sinkende Temperaturen zwischen ca. 5000 und 3000 cal. yr B.P. verstärkt wurde. Kühl-trockene Bedingungen herrschten in Ostgrönland von etwa 3000 bis 1000 cal. yr B.P. vor. Diese Periode fällt mit einer Verstärkung des Ostgrönlandstromes zusammen, die zu einer fast ganzjährigen Eisdecke vor Ostgrönland führte und somit zu einem Rückgang der Brutkolonien auf Raffles Ø. Einer kurzen Erwär-

mung zwischen 1000 und 500 cal. yr B.P. folgte die Kleine Eiszeit zwischen 500 und 100 cal. yr B.P., widergespiegelt sowohl in sehr geringen Gehalten an organischem Material in den Sedimenten, als auch in der Vegetationszusammensetzung in der Seeumgebung.

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1. Introduction

The study of ice cores from Greenland during the last three decades revealed information about climatic changes in the Northern Hemisphere during the past 250,000 years (e.g. Reeh 1989, Alley et al. 1993, Dansgaard et al. 1993, GRIP Members 1993, Meese et al. 1994, Johnsen et al. 1995). The most prominent and deepest ice cores are the Greenland Ice-Core Project (GRIP) and Greenland Ice Sheet Project II (GISP2) ice cores from the central ice sheet, Dye 3 from southern Greenland and Camp Century from the northwestern part of Greenland (Fig.1.1). An important contribution to the reconstruction of the Holocene glacial and temperature history in East Greenland comes from an ice core drilled on the Renland ice cap, interior Scoresby Sund region (Johnsen et al. 1992). In contrast to the more stable isotopic $\delta^{18}O$ records from the central Greenland ice sheet during Holocene times, the Renland record shows a broad maximum in its record during the early Holocene and thus indicates longitudinal differences.

Numerous onshore studies, for instance the PONAM project (Möller et al. 1994), enabled the reconstruction of the extensions and the periods of glacial advances and recessions in East Greenland (Hjort 1979, Möller et al. 1991, Funder et al. 1994, Funder and Hansen 1996). Study objects were mainly geomorphological structures, the occurrence of fossils, and lake sediments. They discovered a succession of glacial advances, recessions and interspersed marine transgressions during Pleistocene and Holocene times. Last important events at the transition to the Holocene epoch were the glaciation of the outer fjord region during the Milne Land stade at 11,300-11,150 cal. yr B.P. and a subsequent marine transgression during times of glacier recession (Hjort 1979, Funder and Hansen 1996, Björck et al. 1997). The palaeoenvironmental information in the terrestrial archives for the following period has a poor time resolution. However, the occurrence of animals during the early Holocene, which are not common in East Greenland today, gave a valuable hint on a warmer period (Bennike et al. 1994, Böcher and Bennike 1996, Bennike 1997, Bennike and Funder 1997). This is confirmed by a rapid glacier recession of the outlet glaciers into the inner fjord region after the Milne Land stade (Hjort 1979).

Marine studies on the extension of glaciers in the fjords and on the shelf (Fig. 1.1) are commonly based on seismic measurements (Dowdeswell et al. 1994, Hubberten 1995, Hubberten et al. 1995). Complementary information obtained by the amounts of ice-raftet debris indicate glacial oscillations (Marienfeld 1991, Nam 1997, Funder et al. 1998). The fossil occurrence of bivalves, diatoms, and the isotopic signals from foraminifera were correlated to climatic changes (Hjort and Funder 1974, Koç et al.

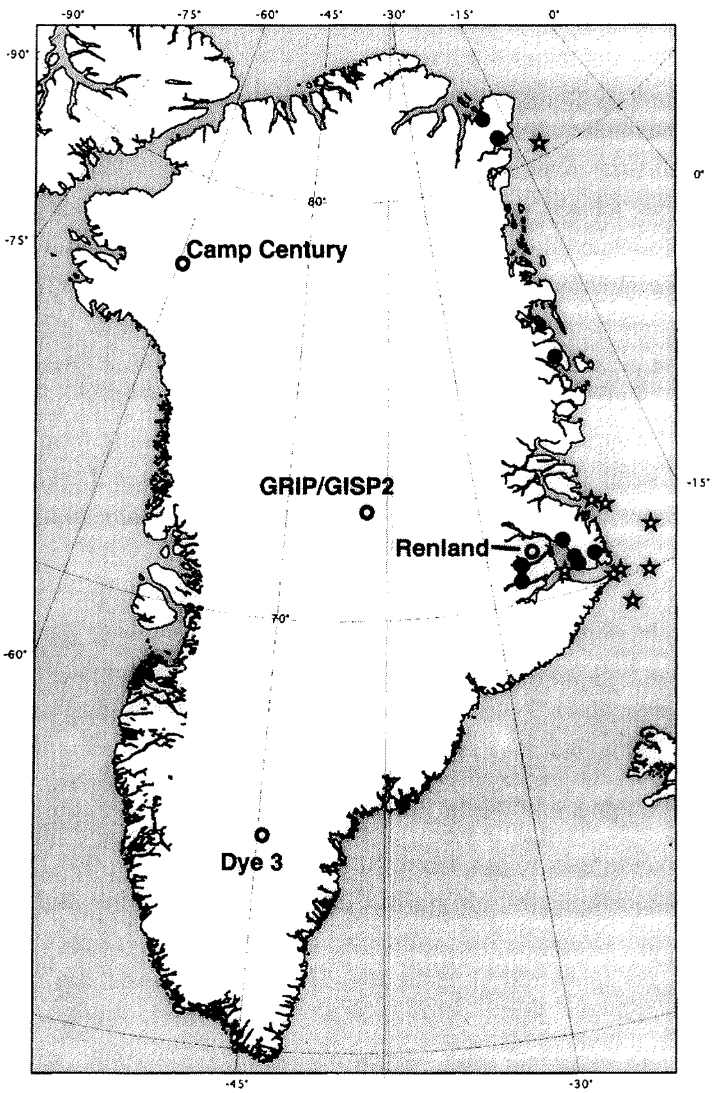


Fig. 1.1: Location of the most prominent Greenland ice cores with their names (open circles), marine studies carried out in the fjords and on the shelf of East Greenland (asterisks), and lacustrine sediment records from East Greenland (black dots). References are given in the text.

1993, Nam, 1997, Notholt 1998). However, due to the low sedimentation rates, the most marine records generally lack with high time-resolution for the Holocene.

Lacustrine sediment records have a much higher resolution for the Holocene period, similar to those of isotopic records from ice cores (Willemse and Törnqvist 1999). In eastern and northeastern Greenland, most existing lacustrine sediment records are based on palynological information, partly complemented with information obtained from biogeochemical measurements or the occurrence of fossils (Funder 1978, Björck and Persson 1981, Funder and Abrahamsen 1988, Björck et al. 1991, Fredskild 1992, Björck et al. 1994a, 1994b, Fredskild 1995, Bay and Fredskild 1997, Bennike and Funder 1997; Fig. 1.1). The results of these studies reveal a succession of warmer and colder periods during the Holocene, and distinct differences between the outer and the inner coastal region. However, almost all of these lakes cored have maximum water depths of less than 10 m, and the major part of the coring locations have water depths of less than 5 m. Therefore, the studied sediments could have been affected distinctly by waves, a thick lake-ice cover, or lake level fluctuations. In addition, recent studies on some of these lakes indicated too old ages obtained at the former studies, probably due to bulk sediment dating, reservoir effects, or the presence of coal in the catchments (Funder 1978, Björck and Persson 1981, Funder and Abrahamsen 1988, Björck et al. 1991, Björck et al. 1994b).

The aim of the new studies on lacustrine sediments is a detailed reconstruction of the Holocene environmental history of East Greenland, based on radiocarbon dating of terrestrial plant remains and other hand-picked fossils. Five lakes were selected according to sufficiently deep water to prevent sediment disturbation due to an influence of waves, lake-ice cover, or lake level fluctuations at the coring locations. Different distances of the lakes to the ice sheet and the outer coast enable valuable information about the glacier recession and the coinident marine transgression after the last glaciation. The Holocene temperature and precipitation changes indicated in the lacustrine sediment records may form a link between the information derived from the ice core records and marine records. Whilst the ice core records are apparently most influenced by atmospheric changes, the marine records are strongly dependent on changes in the oceanic circulation pattern. The multi-disciplinary approach, using sedimentological, biogeochemical, palynological data, and fossils, allows the evaluation of the need of individual parameters for palaeoenvironmental reconstructions. Between-site differences may enlighten local-scale effects, whilst similar variations in records from different sites indicate environmental changes at the regional scale.

2. Study area

2.1. Late Pleistocene history and Holocene transition

The study area for the reconstruction of the Holocene environmental history is a coastal region at East Greenland, which is bordered to the north by Hochstetter Forland and Shannon Ø, and to the south by the Scoresby Sund. The central Greenland ice sheet forms the western boundary and the Greenland Sea the eastern one (Fig. 2.1).

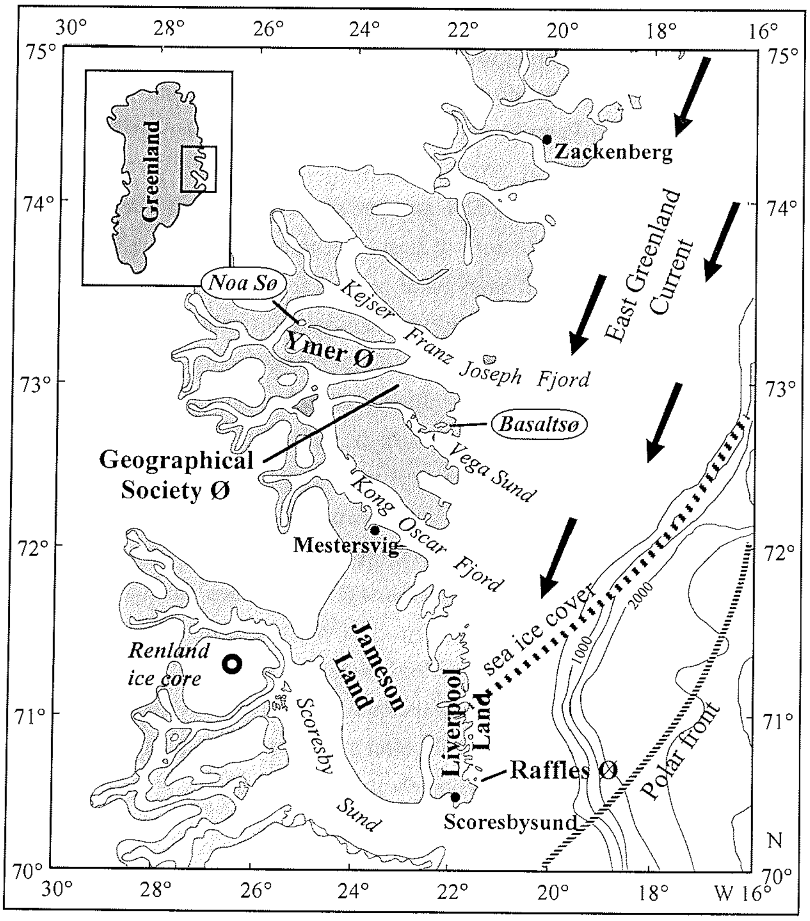


Fig. 2.1: Map of East Greenland with approximated boundaries of the Greenland ice sheet and local ice caps. Settlements are marked by black dots, and the open dot indicates the coring location of Renland ice core. The position of the Polar front is modified after Henrich (1998), that from the summer sea ice cover after Koç et al. (1993).

East Greenland is characterized by numerous fjords. Major fjords are the Scoresby Sund and the Kong Oscar Fjord, which both penetrate the inland up to a distance of 300 km (Fig. 2.1). Thus, they belong to the largest fjords in the northern hemisphere. The narrow fjords in East Greenland have steep slopes and are deep incised into the gently undulated high mountain plateaus at 1500-2500 m elevation, which are commonly covered by small ice caps in the more interior region (Funder 1989). Towards the east the landscape is gently sloping downwards and forms partly extensive lowlands at the coast.

The general bedrock geology of East Greenland is characterized by a belt shaped succession of older Lower and Middle Proterozoic (Caledonian) metamorphic/ plutonic rocks in the west via Devonian, Carboniferous and Permian sedimentary rocks towards younger Tertiary igneous rocks in the east (Harpøth et al. 1986, Funder 1990). The major part of the ice-free area is exposed as glacially eroded bedrock with scattered eratic boulders. In minor parts, in particular in the flat areas along the coast, the bedrock is drapped by Quaternary sediments. These are commonly formed by thin and discontinous ground moraines from the last glaciation during Late Weichselian and Holocene times. Exceptions are tills on the high mountain plateaus of Jameson Land, which document older than Late Weichselian glaciation of East Greenland, from the so called Scoresby Sund glaciation during Saalian age (Möller et al. 1994, Funder et al. 1998). The present exposure of older than Late Weichselian tills gives valuable hints on the intensity of the Late Weichselian glaciation.

The Late Weichselian glaciation culminated between ca. 22 and 15 ka B.P. This entire period corresponds to the Last Glacial Maximum (LGM) or to the Greenland Stadial 2 event (GS-2), dated to 21.1-14.7 GRIP ka B.P. with an intervening less cold period from 19.5-16.9 GRIP ka B.P. (Björck et al. 1998, Walker et al. 1999). During the LGM the temperature was about 20°C below present temperatures (Funder et al. 1998, Dahl-Jensen et al. 1998) and precipitation was ca. 20% of the present values. This extreme arid climate is probably the reason for a only scattered glaciation of the outer lowlands in East Greenland which had little erosive effects on the landscape. The fjords were filled by thin glacier tongues that occasionally formed end moraines at the mouths of the larger fjords. For instance, the Kap Brewster moraine in the Scoresby Sund originates from the LGM, and another unnamed one at the Kong Oscar Fjord further to the north (Hubberten et al. 1995, Funder et al. 1998). Despite the cold conditions at this time, a relatively high amount of ice rafted debris (IRD) in offshore records indicates at least partly open-water conditions that enabled the drift of icebergs (Nam 1997).

The glacier recession after the LGM took place in two major steps. The first step began with the Greenland Interstadial 1 (GI-1) at ca. 15 ka B.P. and cleared the shelf and the major inlets of marine based ice, in particular in the more southern parts of

East Greenland (Funder and Hansen 1996, Björck et al. 1998, Walker et al. 1999). The reasons for the glacier recession were an increase of the northern Hemisphere insolation that culminated at the Pleistocene/Holocene transition, and, more important, calving due to a global rising sea level (Berger 1978, Fairbanks 1989, Hald and Aspeli 1997, Haflidason et al. 1998). The Younger Dryas period (Greenland Stadial GS-1, 12.65-11.5 GRIP ka B.P.; Björck et al. 1998, Walker et al. 1999) was probably too cold and too dry to afford a new readvance of the glaciers, since there are no evidences in East Greenland. However, the first step of glacier recession ended at least at the cooling of the Milne Land stade, a period that corresponds to the Preboreal oscillation dated to 11,300-11,150 cal. yr B.P. (Björck et al. 1997, Funder et al. 1998).

During the Milne Land stade many glaciers readvanced in East Greenland, as a result of increased precipitation and temperatures at the establishment of the present day pattern of atmospheric circulation (Funder and Hjort 1973). Distinct geomorphological features close to the present fjord mouths and the lack of marine fossils in the fjords indicate maximum positions of glaciers during this period (Hjort 1979). In some areas these positions remained until ca. 9.5 ka B.P., which corresponds to ca. 10,050 cal. yr B.P., before glacier recession commenced (Hjort 1979).

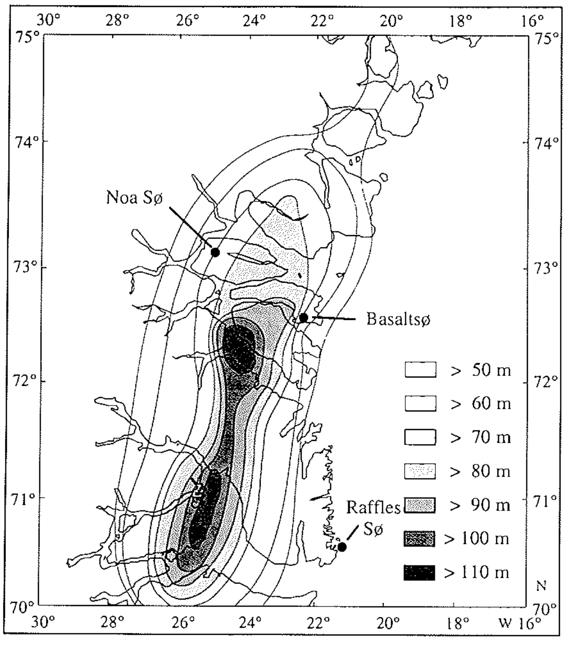


Fig. 2.2: Isobases for the maximum marine limit of the transgression after the Milne Land stade reconstructed from bivalve fossils and marine onshore features along the fjords (after Hjort 1979). Maximum altitudes were attained in areas of the Milne Land stade ice margin (Funder and Hansen 1996).

The glacier recession after the Milne Land stade was asynchronous in East Greenland. Stillstands or even small readvances during the recession were likely, either induced by short-term climatic shifts, or due to different geomorphological structures in the fjords. In contrast to GI-1, the driving factor of glacier retreat after the Milne Land stade was probably the northern Hemisphere insolation maximum, supported by a distinct change in the North Atlantic circulation pattern and still rising sea level (Berger 1978, Fairbanks 1989, Koç et al. 1993, Henrich et al. 1995, Henrich 1998, Haflidason et al. 1998). One of these reasons, the inflow of warm water masses into the North Atlantic is still controversial (Sarntheim et al. 1992, Sarntheim and Altenbach 1995, Hebbeln et al. 1998), but led to a distinct temperature rise of about 5.5°C in the Norwegian Sea during the early Holocene (Hald and Aspeli 1997).

Due to the glacier recession after the Milne Land stade, the fjords became progressively ice free and thus gave way for an extensive marine transgression coinciding with an isostatic rebound (Fig. 2.2). In addition, the glacier recession led to a formation of numerous lakes and ponds onshore. They were left in front of retreating glaciers by the meltwater infill of basins or were formed by melting of dead ice after the glacier recession.

2.2. Present climate and vegetation

The present temperatures in East Greenland are characterized by a latitudinal decrease northwards due to reducing insolation energy, and by a longitudinal decrease from the interior to the coastal region (Tab. 2.1). The major reason for this longitudinal temperature trend is the cold East Greenland Current (< 0°C) that transports low-salinity water masses southwards along the coast of East Greenland (Swift and Aagaard 1981, Andrews et al. 1998). Both the low salinity and the cold surface waters lead to an almost perennial sea-ice cover off East Greenland (Fig. 2.1), thus causing temperature inversions and a frequent fog formation in particular in the flat and low coastal regions up to about 300 m a.s.l. (Fredskild et al. 1986, Bradley et al. 1996, Funder et al. 1998, Fredskild 1998). The fog hampers insolation, and leads to a strong temperature decline from the interior to the outer coastal region of East Greenland.

The amount of precipitation along the eastern coast of Greenland varies due to the atmospheric circulation pattern (Tab. 2.1). The main direction of cyclone tracks follows the coast in SW-NE direction. Thus, a relatively high amount of precipitation is supplied particular in the south and in the coastal areas, often deposited as snow also during the summer. Towards the west the stable anticyclone over the Greenland ice sheet blocks the cyclones, and leads temporary to strong catabatic western or northern

Tab. 2.1: Mean July temperatures and annual mean precipitation of stations along the eastern coast of Greenland (different sources: Funder 1979, Harpøth et al. 1986, Ohmura 1987, Ohmura and Reeh 1991, Bay 1992, Funder et al. 1998).

Station	**Latitude**	**Longitude**	**Temperature** (°C)	**Precipitation** (mm)
Station Nord	81°36' N	16°40' W	2	200
Danmarkshavn	76°46' N	18°46' W	3	150
Daneborg	74°18' N	20°13' W	4	200
Myggbukta	73°29' N	21°34' W	4	300
Mestersvig	72°15' N	23°54' W	8	350
Scoresbysund	70°25' N	21°58' W	3	550
Kap Tobin	70°25' N	21°58' W	3	450
Aputitek	67°47' N	32°18' W	3	800
Ammassalik	65°36' N	37°34' W	7	950
Timmiarmiut	62°32' N	42°08' W	5	1500
Prins Christiansund	60°02' N	43°07' W	7	2500

winds. These winds cause a transfer of snow masses towards the south and east facing lee sides of the hills (Meltofte and Rasch 1998).

The vegetational development in the study area is hampered by the harsh climatic conditions, and by the slow soil formation on the continuous permafrost (Funder 1989, Fredskild 1998). The permafrost leads to unstable soils, distinct changes in the annual water balance, and a very poor aeration of soils, conditions which strongly reduce the vegetation assemblage (Pielou 1991, Dierßen 1996). By that, the vegetation is only patchy dispersed in large areas of East Greenland, but shows a general trend from a denser covered continental interior region to a sparser covered maritime coastal region.

In the northern part of the study area the vegetation covers only 2-3% in the interior fjords and even less in the outer coastal region, where the snow cover is thicker in spring and lasts longer during summer (Bay 1992). Most common species are *Salix*, *Cassiope*, *Oxyria*, *Polygonum*, *Saxifraga*, *Vaccinium* and *Empetrum* forming a dwarf shrub heath. In particular *Salix* and *Cassiope* are adapted to a short vegetation period and dominate also the vegetation assemblage in the southern coastal study area (Funder 1979, Elander and Blomqvist 1986, Fredskild et al. 1986). In the more continental interior regions of the southern study area *Betula nana* is abundant due to the general higher temperatures. It grows together with *Dryas* preferably on the southfacing slopes (Funder 1978, Fredskild 1991, Fredskild 1998). Trees are today completely absent from this area.

2.3. Studied lakes

2.3.1. Basaltsø

Basaltsø is situated on the southeastern Geographical Society Ø (Figs. 2.1 and 2.3) at an altitude of 110 m a.s.l. The lake has an elongated shape with a length of ca. 2.3 km and a maximum width of ca. 1 km. The maximum water depth is 22 m in the eastern basin. The catchment covers an area of 15-20 km^2 with steep slopes to the north of the lake up to 700 m a.s.l. and a ridge of slightly more than 200 m a.s.l. to the south of the lake. Today, the catchment does not comprise local ice caps. The main inflow enters the lake at its eastern shore, additionally, numerous brooks drain the slope to the north of the lake during the melting period in summer. Thus, a few small deltas are formed at the northern lake shore, where these brooks enter into Basaltsø. The outflow at the western lake shore drains into the Vega Sund.

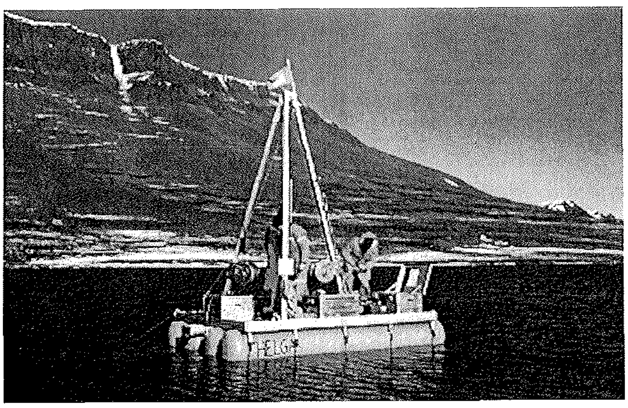


Fig. 2.3: Photograph of Basaltsø with view from the south towards the steep slopes north of the lake. Sediment cores were taken with gravity and piston corers from the floating platform.

2.3.2. Lake B1

Lake B1 is situated ca. 1 km south of Basaltsø at an altitude of 140 m a.s.l. Its circular shape has a diameter of ca. 300 m (Fig. 2.4), and the maximum water depth is 9 m in the center. The catchment covers an area of less than 1 km^2 on the southern ridge, from where the main inflow enters the lake. The outflow is situated on the opposite shore and drains into the western basin of Basaltsø.

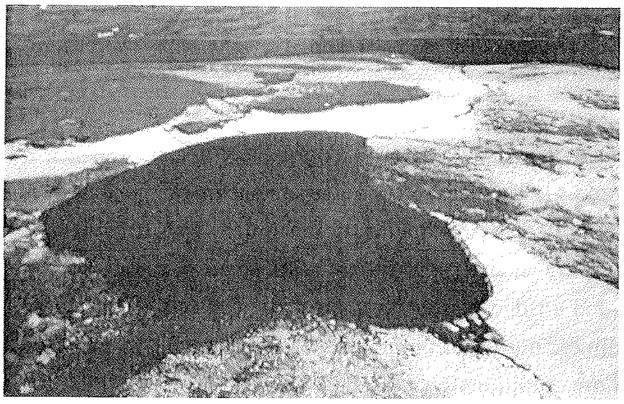


Fig. 2.4: Photograph of lake B1 with view from the southern ridge towards Basaltsø in the background.

2.3.3. Noa Sø

Noa Sø is situated at an altitude of 32 m a.s.l. on an anticline that separates Ymer Ø in a northern and a southern part (Figs. 2.1 and 2.5). The lake has an irregular oval shape with a length of 3.8 km in W-E direction and a width of ca. 2.7 km in N-S direction. The maximum water depth is more than 120 m. The main inflow enters the lake at the southeastern corner of the lake forming a large delta. The inflow is mainly fed by an ice cap to the south of Noa Sø, which culminates at an altitude of more than 1650 m a.s.l. The outflow at the eastern shore drains into the Dusens Fjord.

Fig. 2.5: Photograph of Noa Sø with view from the east. The main inflow forms the large delta to the left, and a ridge at the opposite lake shore separates the lake from Kejser Franz Josephs Fjord. The arrow marks the position of lake N1.

2.3.4. Lake N1

Lake N1 is situated 1.5 km to the northwest of Noa Sø. Its basin spans in W-E direction at an altitude of 120 m a.s.l. with a length of 1 km, a width of 0.5 km, and a maximum water depth of 27 m in the central part (Fig. 2.6). The catchment area of 3.5 km^2 is strongly elongated due to a large ridge to the north of the anticline and a smaller one to the south that divides the catchment areas of lake N1 and Noa Sø. The main inflow enters at the estern lake shore, whereas the outflow is opposite.

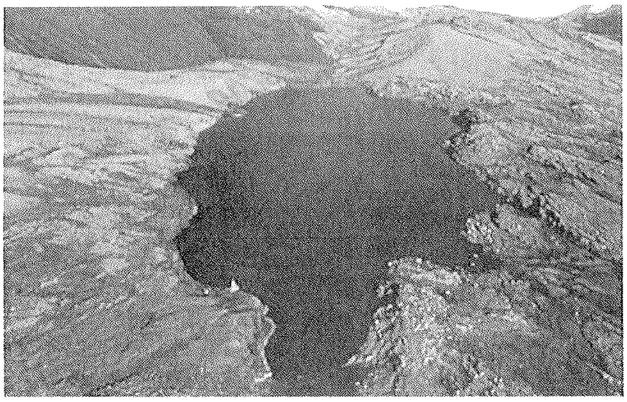


Fig. 2.6: Photograph of lake N1 taken from helicopter with view from the west. The inflow is in the background, the outflow in the front.

2.3.5. Raffles Sø

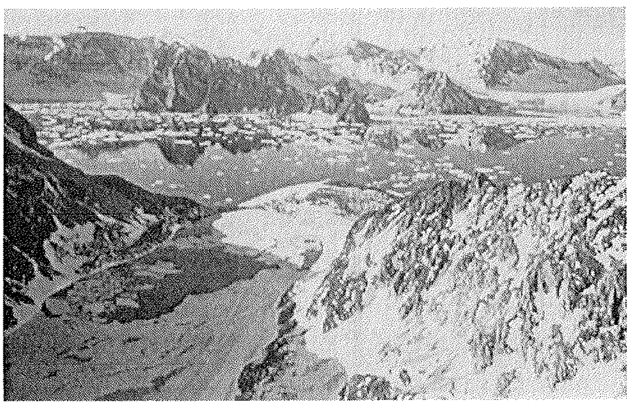


Fig. 2.7: Photograph of the ice-covered Raffles Sø (in front) from the steep slopes to the northeast. Liverpool Land is in the background.

Raffles Sø is situated 30 m a.s.l. on Raffles Ø, which is ca. 3 km in front of Liverpool Land (Figs. 2.1 and 2.7). The 1 km long and 0.3 km wide lake originates from a cirque and has a maximum water depth of ca. 60 m, steep surrounding slopes up to 550 m a.s.l., and a small catchment area of less than 1.6 km^2. The inflow is fed by perennial snowfields at the northeastern lake shore, opposite to the outflow at the southwestern corner.

3. Methods

Bathymetrical and shallow-seismic measurements using a surface echosounder (Furuno) and a sediment echosounder (GeoChirp 6100A, Geoacoustics Corp.) were carried out to obtain information about the basin morphologies of the lakes and their sediment structures prior to coring. A detailed description of the Chirp technique is given by Niessen and Melles (1995). Sediment cores of different lengths were taken using a gravity corer and a piston corer. Melles et al. (1994) describe in detail the coring technique. The most sediment sequences reached a glacial facies or bedrock at the base. This was at the coring site in Basaltsø at a depth of 10 m, in Noa Sø at 2 m, and in Raffles Sø at 2.7 m. In the lake B1 a sediment sequence 2.6 m was recovered, and in lake N1 one of 7.4 m, both with a glaciolimnic facies at their base.

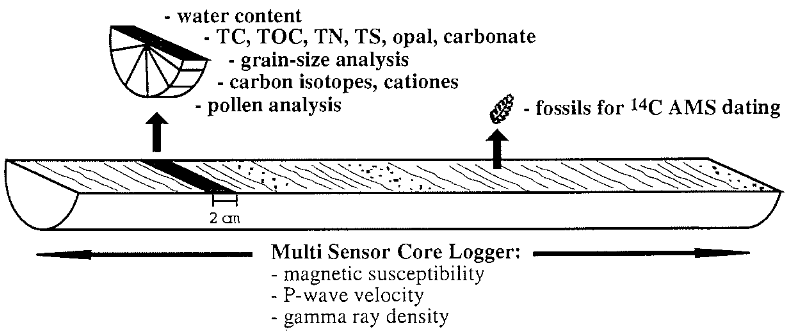


Fig. 3.1: Core half with the methods used for investigation. The physical properties measured with the Multi Sensor Core Logger were attained prior to core opening.

The physical properties magnetic susceptibility, gamma ray density and P-wave velocity of the sediments were measured with a Multi Sensor Core Logger (MSCL, Geotek Corp.) in intervals of 1 cm at the whole-core segments (Fig. 3.1). Technical details are given in Weber et al. (1997). After core opening and subsampling in intervals of 2 cm, the water content (in % of the wet bulk sediment) of the subsamples was calculated from the mass differences between wet and freeze-dried samples.

Grain-size analyses were conducted by a laser particle analyser GALAI CIS-1 (LOT Corp.). For a higher resolution of the fine-grained fraction, the maximum diameter of grain sizes measured with the GALAI CIS-1 was limited to 150 μm.

The preparation of the samples selected for palynological analyses, the determination and counting of pollen grains was carried out by J. Hahne according to a method described in detail in Hahne and Melles (1999). *Lycopodium* spores (tablet with 12,548 spores) for the calculation of the absolute pollen values in each sample were added at core PG1214.

Biogeochemical analyses were carried out after sediment grounding to < 63 μm for homogenization of the sediment within the subsamples. The contents of total carbon (TC), total nitrogen (TN) and total sulphur (TS) were measured with a CHNS-932 determinator (LECO Corp.). Total organic carbon (TOC) was analyzed with a Metalyt-CS-1000-S (ELTRA Corp.) in corresponding samples, which had been treated with HCl (10 %) at a temperature of 80°C to remove carbonate. The carbonate contents were calculated from the contents of carbonaceous carbon, derived from the difference between TC and TOC, and the atomic weights of the elements. Biogenic silica (opal) contents were measured according to the wet chemical method described by Müller and Schneider (1993).

Isotopic measurements of $\delta^{13}C$ on carbonates were carried out using a Finnigan MAT Delta S mass-spectrometer. An Inductively Coupled Plasma Optical Spectrometer (ICP-OES, Perkin-Elmer) enabled the determination of the carbonate cations.

For radiocarbon dating samples of terrestrial plant remains, bivalves, foraminifera and a sticklebar of a fish were separated from bulk sediment using tweezers and a needle under a binocular microscope. The samples were dated by Accelerator Mass Spectrometry (AMS) at the Research Laboratory for Archaeology and the History of Art at the University of Oxford, and at the Van de Graff Laboratory, University of Utrecht. In addition, four samples of bivalves were dated at the Alfred Wegener Institute, Research Unit Potsdam, by conventional ^{14}C dating. All ^{14}C ages were calibrated into calendar years before present (cal. yr B.P.) using the calibration programme CALIB 4.0 (Stuiver and Reimer 1993, Stuiver et al. 1998). Their means and uncertainties are calculated from the upper and lower boundaries of the probability distribution at the 2σ level.

4. Holocene climate history of Geographical Society Ø, East Greenland - evidence from lake sediments

Bernd Wagner, Martin Melles, Jürgen Hahne, Frank Niessen und Hans-W. Hubberten

(Palaeogeography, Palaeoclimatology, Palaeoecology, in press)

4.1. Abstract

Sediment cores from two lakes in the outer coastal region of East Greenland were investigated for chronology, lithology, palynology, and biogeochemistry. A 10 m long sequence recovered in Basaltsø comprises the entire lake history following the last glaciation of the area, probably during the Preboreal oscillation. This is indicated by a succession from glacial via glaciolimnic to limnic sediments. Deglaciation of the area was associated with a high sedimentation rate, mirrored also in the basal part of a 2.6 m long core from a smaller lake (B1) about 1 km south of Basaltsø. Limnic sedimentation without glacial influence commenced about 10,000 cal. yr B.P., according to radiocarbon-dated terrestrial plant remains. Biogeochemical and palynological data indicate an early Holocene climatic optimum from 9000 to 6500 cal. yr B.P. A climatic deterioration began at 6500 cal. yr B.P with an increase of snow accumulation, documented by a change in the pollen assemblage and a coinciding change in the grain-size distribution. At least since 5000 cal. yr B.P, a decrease in the biogeochemical parameters in both lake sediment successions indicates a temperature decline. This deterioration culminated at about 3000 to 1000 cal. yr B.P., when the climate was cold and dry. A slight warming is indicated in the pollen assemblage between ca. 1000 and 800 cal. yr B.P. Following a subsequent rise in precipitation, cooling during the Little Ice Age is mirrored in lowest dwarf shrub pollen percentages and in low contents of organic components.

4.2. Introduction

During the past decades, detailed information about Late Quaternary climate history was obtained by investigations of the Greenland ice cores (e.g. Dansgaard et al. 1993, GRIP Members 1993, Meese et al. 1994, Dahl-Jensen et al. 1998). Most detected climatic trends are validated by other records from across the Northern Hemisphere. An important contribution to the glacial and temperature history comes from an ice core drilled on the Renland ice cap in the Scoresby Sund fjord system in East Greenland (Johnsen et al. 1992, Funder et al. 1998). Although not recovering ages as old as those from the Greenland ice sheet, the Renland ice core provides a high resolution record of the Holocene.

In addition, important information about the glacial history of East Greenland was obtained by the PONAM (Polar North Atlantic Margins; Late Cenozoic Evolution) program, which allowed the reconstruction of glacial and marine transgression periods during Saalian, Eemian, Weichselian and Holocene time in Jameson Land, East Greenland (Funder et al. 1991). Hjort (1979) documented the last glacial advance in East Greenland during the Milne Land stade, which corresponds to the Preboreal oscillation 11,300 to 11,150 cal. yr B.P. (Björck et al. 1997, Funder et al. 1998). The following deglaciation led to ice-free conditions in all outer parts of the fjords within about 1000 years.

In marine records, the climate history of East Greenland is documented by the occurrence of bivalves and diatoms, and by the isotopic signals from foraminifera (Hjort and Funder 1974, Marienfeld 1991, Koç et al. 1993, Nam 1997, Notholt 1998). Most sedimentary records indicate an early Holocene climatic optimum in East Greenland between 8000 and 5000 ^{14}C yr B.P., corresponding to about 8900 to 5800 cal. yr B.P., that is followed by a cooling. These climate changes correspond with changes in the North Atlantic thermohaline circulation (Taylor and Stephens 1980, Duplessy et al. 1992, Veum et al. 1992, Koç et al. 1993, Stein et al. 1994, Sarntheim and Altenbach 1995, Ingólfsson et al. 1997, Hebbeln et al. 1998). Compared to limnic records, the time resolution in marine records is generally poor because of the low sedimentation rates.

Lake sediment records of East Greenland form an important link between ice core records and marine records. They provide high-resolution data archives and yield information about the development of temperature and precipitation. Most existing lacustrine records of East Greenland are based on palynological data (Funder 1978, Björck et al. 1991, Björck et al. 1994a, Fredskild 1995). However, the vegetational succession is often delayed compared to the climatic development due to the state of soil formation, the spreading form of species or the distance from their closest relict areal (Pielou 1991). In contrast, lake environments have a much more rapid reaction to climatic shifts due to the reduced lifetime of freshwater organisms, often combined with a high reproduction rate.

Most age models of the existing records are based on conventional radiocarbon dating of bulk sediment samples (Funder 1978, Björck et al. 1991, Björck et al. 1994a). They have high dating uncertainties due to reservoir effects in hard-water lakes, lake-level fluctuations, and the presence of old carbon in the sediments (Björck et al. 1994b, Noe-Nygaard 1995).

The aim of our lacustrine sediment studies is a detailed reconstruction of the Holocene deglaciation and climate history of the island Geographical Society Ø using a multi-disciplinary dataset, derived from combining sedimentological and biogeochemical parameters from two neighbouring lakes with a palynological record. The radiocarbon dating of terrestrial plant remains by accelerator mass spectrometry (AMS) provides a secure chronology for the interpretation of the dataset. The small distance between the lakes and the Renland ice cap with its Holocene isotopic record provides a unique opportunity to evaluate the significance of single sediment parameters for palaeoclimate reconstruction. Thus, the timespan needed for the adjustment of a lake equilibrum with the climate and the surrounding vegetation may be indicated, and local effects may be distinguished from regional effects.

4.3. Geographical setting

Sediment cores investigated in this study were recovered from two lakes on southeastern Geographical Society Ø, East Greenland, namely the Basaltsø and a small, un-named lake in its catchment, referred to here as lake B1 (Figs. 4.1 and 4.2).

The Basaltsø is an elongated, east-west orientated lake, 2.3 km long and 0.5 to 1.0 km wide. It is located at an altitude of 110 m a.s.l., about 30-40 m over the limit of the postglacial marine transgression in this region (Hjort 1979). Echosounding along 18 profile lines crossing Basaltsø indicates a maximum water depth of about 22 m in the eastern lake (Fig. 4.2). The catchment area of Basaltsø is 15-20 km^2. To the north, the landscape ascends rather steeply to about 700 m a.s.l. To the south, a gently inclined ridge of about 200 m a.s.l. separates Basaltsø from Vega Sund. Lake B1 is located on the northern flank of this ridge, at an altitude of about 140 m a.s.l. The roughly circular lake B1 has a diameter of about 300 m and only a small catchment area (Fig. 4.2). Echosounding in lake B1 revealed its uniform shape, with maximum water depths of about 9 m in the lake centre.

The climate in East Greenland is influenced by latitudinal trends, and also by the oceanic and atmospheric circulation. A cooling trend towards the north due to decreasing radiation energy and increasing influence of the cold (< 0°C) East Greenland Current is shown by temperature records at meteorological stations along the coast (Fig. 4.1). Whilst mean values of ca. 9°C during the warmest month are recorded at Mestersvig, Kong Oscar Fjord (Harpøth et al. 1986), those at Zackenberg (Meltofte and Rasch 1998), 250 km further north and at a comparable distance from the shore, are lower than 6°C. The cold sea surface, covered by ice for most of the year, causes temperature inversions with the frequent formation of coastal fog. The fog hampers

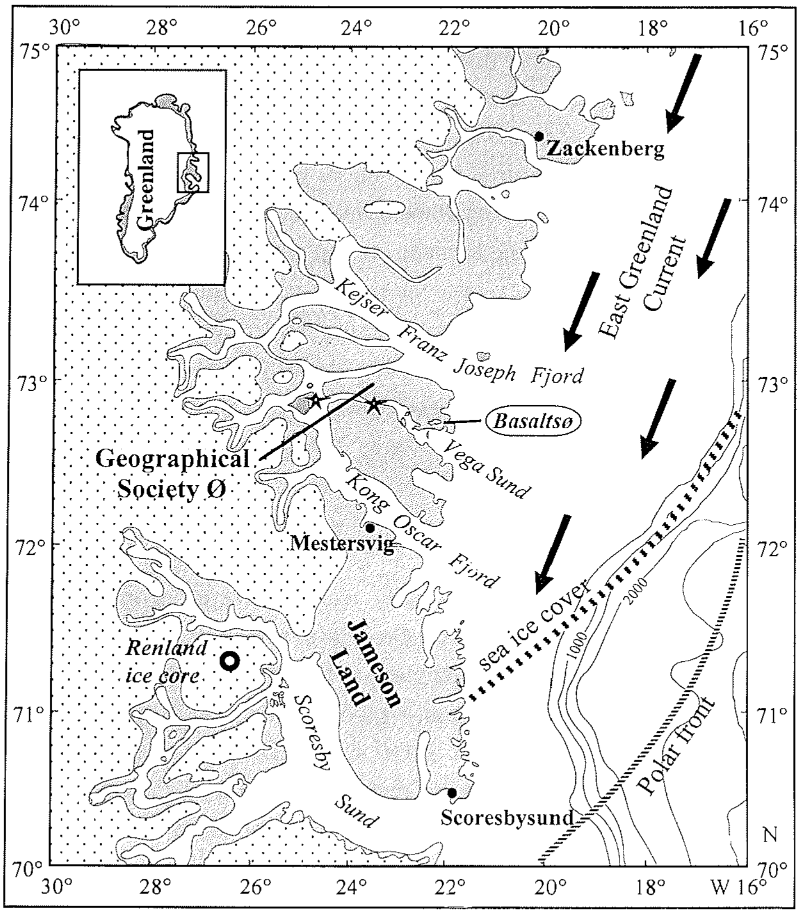


Fig. 4.1: Map of East Greenland showing the Basaltsø on southeastern Geographical Society Ø and geographical names mentioned in the text. Stars indicate the locations of radiocarbon-dated bivalve shells (Håkansson 1973). Settlements are marked by black dots, and the open dot indicates the coring location of Renland ice core. The position of the Polar front is modified after Henrich (1998), that of the summer sea ice cover after Koç et al. (1993).

insolation and thus leads to a temperature decrease from the inner to the outer coast (Fredskild 1998), intensified by the stable high pressure over the ice sheet. This is shown by a mean temperature of 3°C during the warmest month at the Scoresbysund settlement (Funder 1989), being distinctly lower than that at Mestersvig further inland and further north.

Precipitation in East Greenland is mainly supplied due to maritime air masses that follow the cyclone tracks along the coastline from southwest to northeast (Funder 1989). As a consequence, a gradient from wetter conditions in southern Greenland and at the outer coast to dryer conditions northwards and to the inner regions is observed

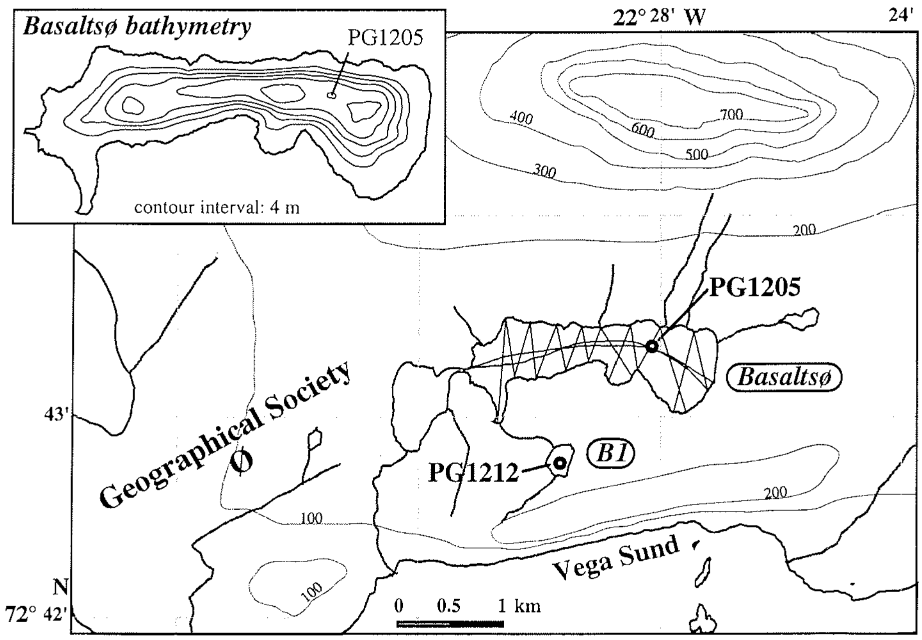


Fig. 4.2: Map of the Basaltsø region. The steep slope to the north of Basaltsø and the small ridge to the south of lake B1 are indicated by approximated contour lines (in metres). The coring sites are shown by heavy dots, the tracks of echosounding and subbottom profiles by small lines. The inset shows the bathymetry of Basaltsø with the coring site.

(Reeh 1989). The mean annual precipitation decreases from about 550 mm/yr at Scoresbysund (Funder 1989) via 350 mm/yr at Mestersvig (Harpøth et al. 1986) and 300 mm/yr in the Basaltsø region (Reeh 1989) to 220 mm/yr at the Zackenberg station (Meltofte and Rasch 1998; Fig. 4.1).

The present vegetation at the outer coast of East Greenland is characterized by a poor dwarf shrub heath with dominating *Salix arctica*, *Dryas* and *Cassiope tetragona* (Funder 1978, Fredskild and Mogensen 1998). In the surrounding area of Basaltsø, the vegetation is scattered between exposures of boulder debris and bedrock. Taller shrubs and trees are completely lacking.

The bedrock geology in the catchment area of both lakes is characterized by Cretaceous sedimentary rocks of terrestrial and marine origin. They are interspersed by dolerite sills that come from a period of effusive volcanism during Late Cretaceous and Early Tertiary times and precede the North Atlantic ocean floor spreading (Funder 1989). To the west of the present catchment area a succession of belts with increasing ages from Upper Carboniferous to Proterozoic occurs, comprising metamorphic, volcanic and sedimentary rocks.

4.4. Material and methods

4.4.1. Subbottom profiling, coring and physical properties

Prior to sediment coring, initial information about the bathymetry and, in Basaltsø, the sediment thickness and internal structures were obtained by acoustic profiling with a surface echosounder (Furuno) and a sediment echosounder (GeoChirp 6100A, Geoacoustics Corp.). A detailed description of the Chirp technique is given by Niessen and Melles (1995). In Lake B1, the coring location was situated in the deepest part of the lake, and in Basaltsø in an area where the unconsolidated sediment fill reaches maximum thickness (ca. 10 m) and is characterized by undisturbed and widely horizontal bedding (Figs. 4.2 and 4.3).

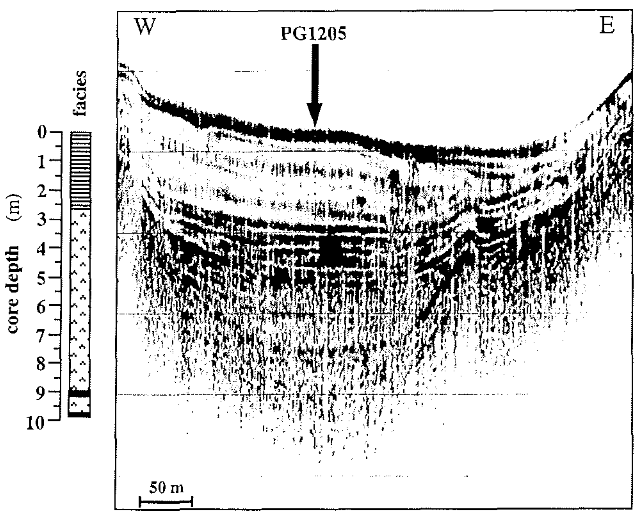


Fig. 4.3: Subbottom profile crossing coring site PG1205 (modified after Niessen and Melles 1995). The geometry and reflection strength are characteristic for the glacial, glaciolimnic and limnic lithofacies cored at site PG1205 (for legend, see Fig. 4.4).

Coring was carried out from a floating platform by gravity corer for obtaining undisturbed near-surface sediments and by piston corer (UWITEC Corp.) for longer sediment sequences, both equipped with PVC tubes of 6 cm diameter. The maximum recovery with every employment of the piston corer is limited by the tube length to 3 m.

Longer records can be sampled by coring of several overlapping depths and subsequent correlation of the core segments. A more detailed description of the coring technique is given by Melles et al. (1994).

In Basaltsø, a complete sediment sequence down to a depth of 9.85 m was recovered by four, partly overlapping piston core segments (PG1205-2 to PG1205-5), and an undisturbed gravity core (PG1205-1). A first correlation of the overlapping depths was based on whole-core measurements of the physical properties P-wave velocity, gamma ray density (GRD), and magnetic susceptibility (Niessen and Melles 1995), carried out in one-centimetre intervals with a Multi Sensor Core Logger (MSCL, Geotek Corp.). This technique is described in detail by Weber et al. (1997). The early correlation was later slightly corrected based on additional information obtained from the core descriptions and biogeochemical analyses (Fig. 4.4). The much shorter, 2.60 m long sediment sequence from site PG1212 in lake B1 only required a correlation between one gravity core (PG1212-2) and one piston core (PG1212-3).

Description and photographic documentation of the cores were carried out immediately after their opening in the laboratory, and after cleaning the surface of one core half which was used for continuous subsampling in 2 cm intervals. The other core half was kept complete and stored as archive for possible future work. The 2 cm thick subsamples were freeze-dried, and their water contents (as a percentage of the wet bulk sediment) calculated from the mass differences between the wet and dry samples. Subsequent splitting of the subsamples into aliquots supplied sufficient material for the measurement of all sediment parameters in the same sample horizons.

4.4.2. Grain-size and biogeochemical analyses

Grain-size analyses of the bulk sediment were carried out in 4 cm (0-374 cm), 8 cm (374-500 cm) and 10 cm (500-985 cm) intervals on core PG1205 by a laser particle analyser GALAI CIS-1 (LOT Corp.). All samples were treated with NH_3 to diminish the surface tension, and put in an ultrasonic sifter for about 15 sec to disaggregate the sediment components before measurements started. A comparison with the pipette method, that is based on sedimentation velocities following Stoke´s Law, has shown lower absolute clay contents determined by the laser technique, but comparable fluctuations within sediment sequences (Konert and Vandenberghe 1997). The maximum grain diameter to be measured with the CIS-1 is limited to 150 μm, but an underestimation of the coarse silt and fine sand (> 20 μm) by incomplete transfer into the measuring cell cannot be excluded.

For measurements of total carbon (TC), total organic carbon (TOC), total nitrogen (TN), and total sulphur (TS), an aliquot of the sediment subsamples was ground (to < 63 μm) and homogenized in achate breakers by a planet mill. The analyses in cores PG1205 were conducted in the same intervals as the grain-size analyses (see above), and in core PG1212 in intervals of 4 cm (0-200 cm) and 8 cm (200-260 cm). The contents of TC, TN, and TS were analyzed with a CHNS-932 determinator (LECO Corp.). TOC was measured with a Metalyt-CS-1000-S (ELTRA Corp.) in corresponding samples that have been treated with HCl (10%) at a temperature of 80°C to remove the carbonate. The carbonate contents were calculated from the contents of carbonaceous carbon, derived from the difference between TC and TOC, and the atomic weights of the elements. On seven samples of core PG1212 (at depths of 120, 132, 144, 196, 216, 232 and 260 cm) the stable carbon isotope ratios ($\delta^{13}C$) of the carbonate were determined by a Finnigan MAT Delta S mass spectrometer; the identification of carbonate was conducted using an inductively coupled plasma optical emission spectrometer (ICP-OES, Perkin-Elmer).

Biogenic silica (opal) contents were measured according to the wet chemical method described by Müller and Schneider (1993), a rather precise method with a relative error of 4-10% in samples containing at least 2% SiO_2, and about 20% in samples with smaller opal concentrations (Müller and Schneider 1993, Conley 1998).

4.4.3. Palynological analyses and radiocarbon dating

For qualitative and quantitative palynological analyses on core PG1205 about 1 g of dry bulk sediment was treated with HF (70%). Pollen grains were further enriched by sieving (6x8 μm mesh). An exact description of the method is given in Hahne and Melles (1997). The amount of counted pollen grains was more than 300 in the uppermost 200 cm of core PG1205, and slightly lower down to 239 cm. Below that depth, pollen were too rare for statistically significant counting.

The palynological results are presented as a total pollen percentage diagram, i.e. the pollen sum (100%) includes all pollen, with the exclusion of aquatics, spores, algae (*Pediastrum* and *Botryococcus*) and redoposited/exotics (incl. *Pinus*, *Betula* sect. *Albae*-type and *Alnus*). Within the genus of *Betula*, the trees (*Betula* sect. *Albae*-type) are separated from the shrubs (*Betula nana*-type), since *Betula nana* is the only species in East Greenland and other species derive from a long distance transport (Fredskild 1991). Willows are grouped under the term *Salix*. Within the Ericales, the term *Cassiope*-type comprises the genus *Cassiope* and other Ericales, which cannot be differentiated by their sizes and morphologies. Despite this difficulty, this group was distinguished from the other Ericales due to its particular importance for the interpretation of the Holocene climatic history of East Greenland (Funder 1978). Equally, *Dryas* was separated from the other Rosaceae. *Oxyria* and *Rumex* were combined to a *Oxyria/Rumex*-type due to the very difficult distinction between the species, even if the majority of grains are believed to belong to *Oxyria*.

Tab. 4.1: Radiocarbon dates discussed in this study: (A) new dates from the Basaltsø and lake B1 sediment cores, and (B) published radiocarbon dates from the region. All dates were calibrated into calendar years before present (cal. yr B.P.) using the calibration program CALIB 4.0 with the terrestrial and marine calibration datasets (Stuiver and Reimer 1993, Stuiver et al. 1998). Means and uncertainties were calculated from the lowest and highest dates at the 2σ probability distribution. Prior to calibration, marine samples were corrected (corr. age B.P.) for a marine reservoir effect of 550 years (Tauber and Funder 1975) if no correction is given in the reference.

(A) core	depth (cm)	lake	material	sample no.	$\delta^{13}C$	^{14}C age B.P.	cal. yr B.P.
PG1205-2	32-34	Basaltsø	twigs	OxA-7253	-27.9	845 ± 40	790 ± 115
PG1205-2	40-42	Basaltsø	twigs	OxA-7286	-27.6	985 ± 50	870 ± 105
PG1205-2	88-90	Basaltsø	leaves, twigs	UtC-8453	-28.0	3050 ± 80	3200 ± 200
PG1205-2	124	Basaltsø	mosses	OxA-7254	-27.9	4175 ± 50	4700 ± 130
PG1205-2	148-150	Basaltsø	leaves, twigs	UtC-8222	-29.9	5433 ± 35	6230 ± 60
PG1205-2	180-182	Basaltsø	leaves, twigs	OxA-7287	-27.7	6455 ± 70	7370 ± 120
PG1205-2	240-242	Basaltsø	leaves, twigs	UtC-8454	-28.0	8960 ± 160	10010 ± 410
PG1212-3	4-6	B1	leaf	OxA-7265	-28.6	215 ± 50	160 ± 160
PG1212-3	36-38	B1	leaves, twigs	OxA-7266	-29.4	1590 ± 50	1460 ± 110
PG1212-3	68-70	B1	leaves, twigs	OxA-7267	-27.8	2970 ± 55	3120 ± 150
PG1212-3	90-92	B1	leaves, twigs	OxA-7268	-29.1	4025 ± 55	4580 ± 230
PG1212-3	118-120	B1	leaves, twigs	OxA-7559	-28.8	6005 ± 80	6840 ± 180
PG1212-3	125	B1	fibres of wood	OxA-7269	-23.8	6535 ± 65	7440 ± 130
PG1212-3	136-138	B1	leaves, twigs	OxA-7270	-27.7	6950 ± 65	7800 ± 130
PG1212-3	160-162	B1	leaves, twigs	OxA-7264	-28.8	8580 ± 160	9620 ± 360

(B) reference	source/ sample no.	^{14}C age B.P.	corr. age B.P.	cal. yr B.P.
Koç et al. (1993)	marine	-	9000	10060 ± 230
Koç et al. (1993)	marine	-	5000	5780 ± 50
Koç et al. (1993)	marine	-	3000	3290 ± 30
Hjort (1979)	marine	10300	9750	11210 ± 390
Hjort (1979)	marine	9500	8950	10050 ± 230
Håkansson (1973)	Lu-646	9740 ± 90	9190 ± 90	10330 ± 370
Håkansson (1973)	Lu-608	8090 ± 80	7540 ± 80	8400 ± 180
Funder 1978, Böcher and Bennike (1996)	terrestrial	8000	-	8890 ± 110
Funder (1978)	terrestrial	7500	-	8280 ± 80
Funder (1978), Funder (1990)	terrestrial	5000	-	5700 ± 40
Funder (1978), Funder (1990)	terrestrial	4000	-	4470 ± 40
Funder (1978), Funder (1990)	terrestrial	3000	-	3170 ± 90
Björck et al. (1994b)	terrestrial	7000	-	7840 ± 80
Björck et al. (1994b)	terrestrial	6500	-	7360 ± 20
Hjort (1997)	marine/terrestrial	-	5000	5780 ± 50

Radiocarbon dating was conducted by accelerator mass spectrometry (AMS) on handpicked terrestrial plant remains. Using tweezers and a needle, the material was separated under a binocular microscope, and mechanically cleaned from other sediment particles. By dating terrestrial plant remains, any influences of reservoir effects in the lake water body on the radiocarbon dates can be excluded. By cleaning the plant remains, any contamination by coal particles that would produce ^{14}C dates that are too old (Funder 1989, Björck et al. 1994a, 1994b) is avoided.

Seven samples from Basaltsø core PG1205, and eight samples from B1 lake core PG1212 contained sufficient organic matter for AMS dating, conducted at the Research Laboratory for Archaeology and the History of Art at the University of Oxford, and at the Van de Graff Laboratory, University of Utrecht. All ^{14}C ages were calibrated into calendar years before present (cal. yr B.P.) using the terrestrial dataset from the calibration programme CALIB 4.0 (Stuiver and Reimer 1993, Stuiver et al. 1998). Their means and uncertainties are calculated from the upper and lower boundaries of the probability distribution at the 2σ level. Most published radiocarbon dates from the region are given in uncalibrated radiocarbon years before present (B.P.), but these dates were calibrated for comparison (Tab. 4.1).

4.5. Lithofacies classification and interpretation

Three sedimentary facies are distinguished on the basis of sediment colours, structures, physical properties, grain sizes, biogeochemistry, and palynology in cores PG1205 and PG1212. They are formed during (1) the last glaciation, (2) the deglaciation, and (3) an unglaciated setting.

The glacial facies occurs twice in core PG1205, at its base (9.75-9.85 m) and as a thin layer (9.08-8.98 m) in the lower part of the core (Fig. 4.4). Core PG1212 did not reach this facies (Fig. 4.5). The facies consists of a diamicton, characterized by a clear dominance of minerogenic sediment particles, a wide range of particle sizes from mud and sand to gravel, and thus a poor sorting. It is massive, is grey in colour, has a remarkably stiff consistency, and shows high gamma-ray densities (GRD), P-wave velocities, and susceptibilities, and low water contents. In the seismic profile crossing the coring location (Fig. 4.3), the upper diamicton may be associated with a weak subbottom reflector that is not further penetrated by the acoustic waves. The biogeochemical parameters opal, total organic carbon (TOC), total nitrogen (TN) and total sulphur (TS) have very low values in this facies, the C/N ratio ranges between 7 and 8, and the pollen concentration is too low for palynological analyses. Carbonate contents reach up to 2%.

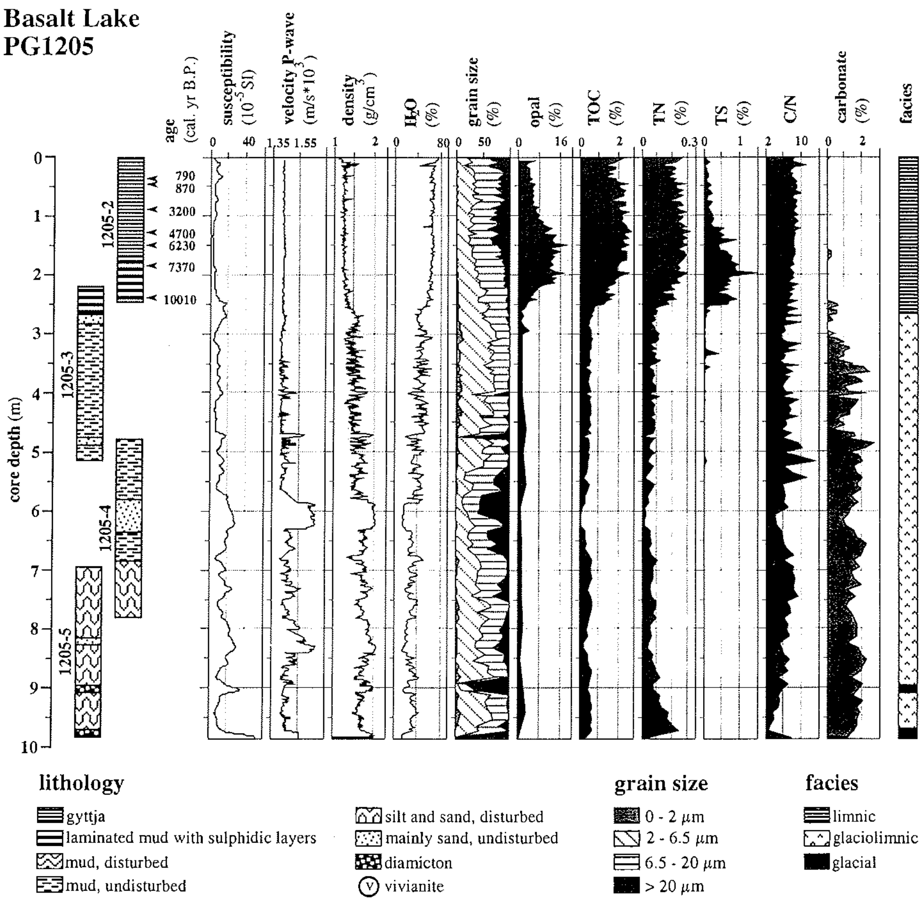


Fig. 4.4: Lithology, physical properties, grain-size distribution, biogeochemical composition, and facies classification of core PG1205 from Basaltsø, with legend for Figs. 4.3-4.6. Core PG1205 consists of four overlapping segments (PG1205-2 to PG1205-5). Black arrows mark the levels of radiocarbon-dated plant remains.

The stiff consistency of both diamicton layers, their acoustic non-transparency, and their physical properties indicate overconsolidation. The overconsolitation, together with the grain-size distribution, the low values of the biogeochemical parameters, and the lack of pollen grains can best be explained by a formation below grounded ice masses (lodgement till). However, the small thickness of the upper till layer, and the lack of an erosion discordance to the underlaying glaciolimnic facies, indicate its redeposition by gravitational sliding or transport by lake ice. Glacial transport of clastic matter fixed on icebergs or ice floes is described by Marienfeld (1991) for glaciomarine

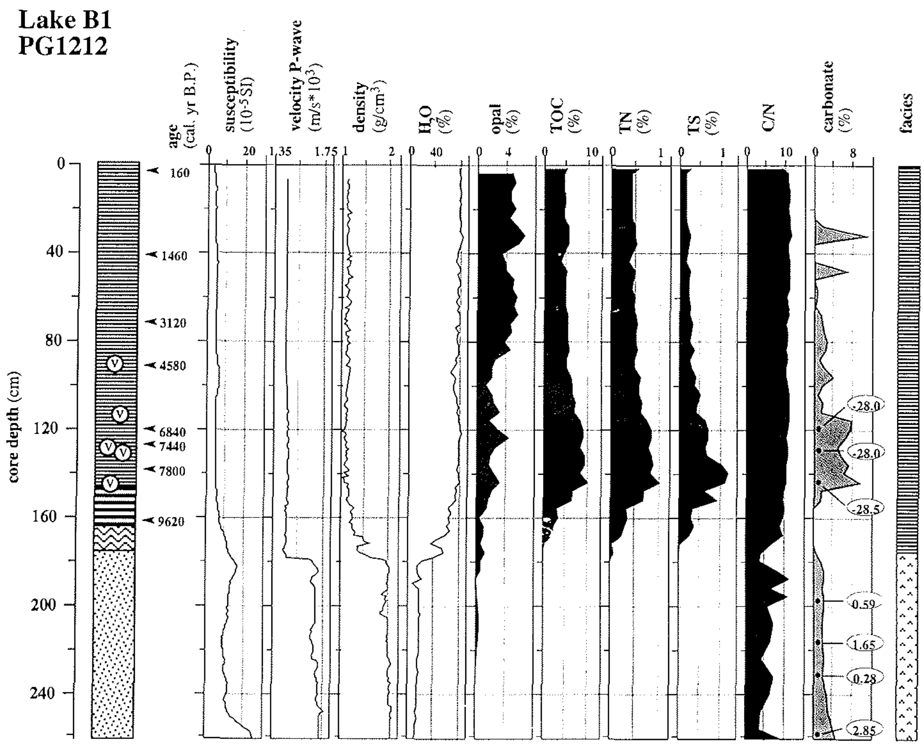


Fig. 4.5: Lithology, physical properties, biogeochemical composition, and facies classification of core PG1212 from lake B1 (for legend, see Fig. 4.4). Black arrows mark the levels of radiocarbon-dated plant remains; carbon isotope ratios in the carbonate of selected samples (in ‰ V-PDB) are given in the carbonate concentration column (encircled data).

sediments from East Greenland, but also characterizes glaciolimnic sediments (Lønne 1995). In any case, the deposition of both till layers is associated with the last ice advance into this region.

The glaciolimnic facies occurs between and above the two glacial horizons of core PG1205, with a sediment depth of up to about 2.7 m (Fig. 4.4), and also exists in the basal part of core PG1212, below 1.75 m (Fig. 4.5). This facies is characterized by an alternation of different lithological units. Muddy horizons with little internal bedding alternate irregularly with stratified sediments containing thin, ungraded sandy layers. These layers commonly exhibit folds, indicative of sediment deformation. In addition, more horizontally bedded layers of several centimetres to decimetres in thickness occur and are commonly graded from sand via silt to clay. Occasionally single gravel grains are embedded in the facies, particularly in its lower part. The sediment colour repeatedly changes with an overall change upwards from greyish to brownish colours.

The seismic profile crossing coring site PG1205 shows well-stratified sediments in the upper part of the facies, but more diffuse reflections with a still high back scatter in the underlaying sediments (Niessen and Melles 1995). The geometry of these diffuse reflectors indicates lenticular sediment bodies. Strong reflectors within this facies correspond with sediment layers of a coarse grain size and low water content, and a high susceptibility, GRD, and P-wave velocity. Pollen concentrations in the facies are very low. This is also valid for the biogeochemical parameters, particularly in intervals with high sand contents. The C/N ratio shows distinct fluctuations, varying between 2 and 14, and carbonate contents amount to 1-6%.

The lenticular sediment bodies with their sharp pinch out towards the slope indicate gravity-driven sediment transport, characterizing most of the facies, in particular in its lower part. This is also indicated by the graded sediments, which may represent turbidites, and by the folding structures that are typical of slumps. Similar folding structures may also be created by rapid sedimentation of a dense, coarse-grained layer onto more fine-grained, muddy sediment with higher water content, leading to unstable conditions and postsedimentary overturning and mixing. All these mass movement processes often occur in front of deltas or glaciers, and normally are associated with high sedimentation rates (Nemec 1990). The single embedded gravel grains in a fine-grained matrix are interpreted as ice-rafted debris. The debris may originate from icebergs, indicating the occurrence of glacier ice at the lake shore. However, it could also have become incorporated into lake ice by basal freezing at the shore and supplied to the coring site by ice floes. The muddy horizons reflect more pelagic sedimentation. Their low contents in biogeochemical parameters and pollen point to a still high terrigenous sediment input, which may derive from glaciers, or from erosion processes as a result of the melting of extensive snow fields in the catchment. Thus, the glaciolimnic facies represents the deglaciation of the area.

The limnic facies forms the uppermost 2.7 m of core PG1205 and the uppermost 1.75 m of core PG1212. This facies is characterized by laminated gyttja of a greyish and brownish colour. In the seismic profile crossing the coring location PG1205 the limnic facies is indicated by relatively transparent layers with a few weak internal reflectors. This is associated with low and more constant values in magnetic susceptibility, density and P-wave velocity, and high water contents of up to about 80% in both cores. Some high-amplitude reflections within this facies give rise to acoustic diffractions below them. The biogeochemical parameters and the pollen concentrations show higher values than in the underlaying sediments in both cores and the C/N ratio stabilizes at values between 6 and 12. Carbonate disappears almost completely in the limnic facies of core PG1205, but is still present in that of core PG1212.

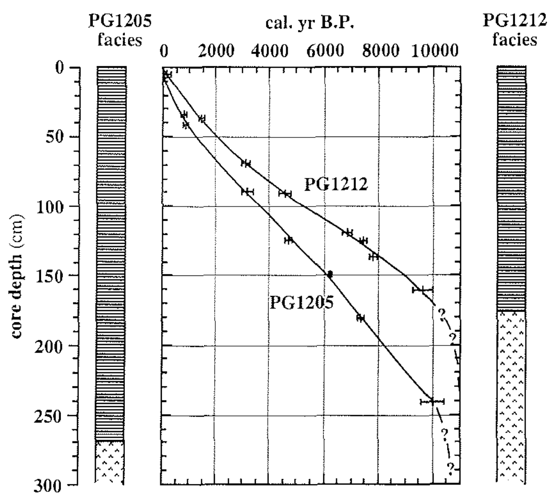


Fig. 4.6: Ages of radiocarbon-dated plant remains (given in calibrated years B.P.) versus depth and facies in cores PG1205 and PG1212 (for legend, see Fig. 4.4). Note that only the upper parts of the cores are shown.

The spotty high-amplitude reflections and diffractions below them are typical for sedimentary gas that may originate from bacterial decomposition. In combination with a generally weak acoustic backscatter, an enrichment of organic matter in the sediment is indicated. In addition, the horizontal lamination and the fine grain size in both cores document quiet, pelagic sedimentation without any significant disturbance by mass movement processes. Sediment supply by glaciers or snow fields in the catchment has diminished, and biogenic sedimentation has gained in importance. The onset of limnic sedimentation is documented by radiocarbon dates from the basal part of the limnic facies in both cores (Tab. 4.1, Figs. 4.4-4.6). Sample OxA-7264 from 1.61 m depth in core PG1212 has an age of 9620±360 cal. yr BP, and sample OxA-7287 from 2.41 m depth in core PG1205 was dated to 10,010±410 cal. yr BP. The radiocarbon dates from lower sediments show decreasing ages with decreasing depth, and thus allow age models to be created (Fig. 4.6).

4.6. Climate history

4.6.1. Pleistocene/Holocene transition

During the Last Glacial Maximum, the mean annual temperature in central Greenland was 23°C lower than today, as estimated from temperature measurements in the GRIP borehole (Dahl-Jensen et al. 1998). According to Funder et al. (1998), precipitation at that time was about 20% of the present value. At the end of the Last Glacial Maximum, glaciers receded from the shelf and the major inlets, caused predominantly by calving (Funder and Hansen 1996).

The Younger Dryas obviously was too cold and too dry to produce glacier advances (Funder et al. 1998). The subsequent warming trend to the Preboreal was interrupted by the Preboreal oscillation, a cold period several hundred years after the termination of the Younger Dryas (Johnsen et al. 1992, Larsen et al. 1995). This oscillation is dated to 11,300-11,150 cal. yr B.P. (Björck et al. 1997). It seems to be consistent with the timing of readvances of fjord glaciers during the Milne Land stade (Björck et al. 1997, Funder et al. 1998), which led to a repeated ice coverage of the outer fjord region of East Greenland (Hjort 1979, Funder 1989). In Vega Sund, glaciers reached their outer limit during this period approximately at the Nordenskjölds Ø (Hjort 1979). Due to its small distance to Basaltsø (about 4 km), and a limnic sedimentation postdating the Milne Land stade (Fig. 4.6), the till layers at the base of core PG1205 in all likelihood derive from this event.

The climatic warming following the Preboreal oscillation was affected not only by the northern hemisphere insolation maximum, but also by an enhanced northward flow of warm Atlantic waters at that time (Koç et al. 1993, Henrich et al. 1995, Hald and Aspeli 1997, Haflidason et al. 1998, Henrich 1998). The Polar front, separating dominating polar surface waters from mixed polar and Atlantic surface waters (Hebbeln et al. 1998), according to Koç et al. (1993) has reached a position off Scoresby Sund at ca. 10,100 cal. yr B.P. (Tab. 4.1).

Initiated by the Preboreal temperature increase, glacier recession in East Greenland commenced between 11,200-10,050 cal. yr B.P. (Hjort 1979, Funder 1989, Björck et al. 1994a, 1994b; Tab. 4.1). The annual rate of recession was dependent on local effects such as basin morphology and on climatic changes (Funder 1989). Periods of standstills and renewed small advances are likely to have occurred. The minimum times of deglaciation in the Vega Sund region are documented by dating of bivalve shells, found in raised marine deposits (Håkansson 1973, Håkansson 1974). The oldest sample from Kap Laura, 35 km to the west of Basaltsø (Fig. 4.1), showed an age of about 10,330 cal. yr B.P. (Håkansson 1973, Tab. 4.1). Another bivalve dating from Kap Elisabeth on Ella Ø, 42 km to the west of Kap Laura, resulted in an age of 8400 cal. yr B.P. (Håkansson 1973; Tab. 4.1). Based on the assumption that the dated bivalves have colonized the deglaciated areas rapidly, the rate of glacier recession between both locations can be calculated to about 22 m/yr, which coincides with the early Holocene recession rate of 20-40 m/yr in the Scoresby Sund region (Funder 1989).

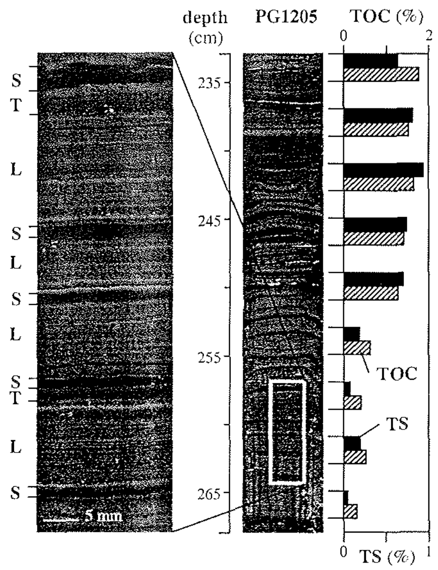


Fig. 4.7: Photographs of the basal part of the limnic facies (233-268 cm) in core PG1205 from Basaltsø, and concentrations of total organic carbon (TOC) and total sulphur (TS) in this sequence (measured in 2 cm thick intervals). The occurrence of black sulphidic layers (S), turbidites (T), and laminated sequences (L) is indicated on the left margin. An increase of sulphidic layers upwards corresponds with a rise of the TS and TOC.

The glacier recession is mirrored in the glaciolimnic facies of Basaltsø and lake B1. In core PG1205 from Basaltsø, the overall upward decrease in grain size, sediment deformation, ice-rafted debris content, and mass movement deposits reflects the increasing distance from a glacial sediment source (Fig. 4.4). In core PG1212 from lake B1, this process is shown by a fining upward from sand to mud (Fig. 4.5). The progressive deglaciation is also indicated by a change in terrigenous sediment supply in the upper parts of the glaciolimnic facies in both cores, reflected by a decrease of the carbonate content. The carbonate has a marine origin, shown by its $\delta^{13}C$ values between 0 and 3‰ in core PG1212 (Clark and Fritz 1997; Fig. 4.5). Since both lakes are located above the limit of the postglacial marine transgression in this region (Hjort 1979), the carbonates in the glacial and glaciolimnic facies most likely derive from glacially redeposited carbonate rocks. The material may originate from bedrocks exposed to the west of Basaltsø, either directly supplied by ice or first deposited in the catchment area and later eroded by brooks draining into the lakes.

The change in sediment supply at the end of the glaciolimnic sedimentation was associated with decreasing sedimentation rates (Fig. 4.6). Due to diminishing terrigenous supply biogenic accumulation gained in importance at the transition to the limnic facies in both cores (Figs. 4.7-4.9). A significant increase of TOC, TN and TS, of

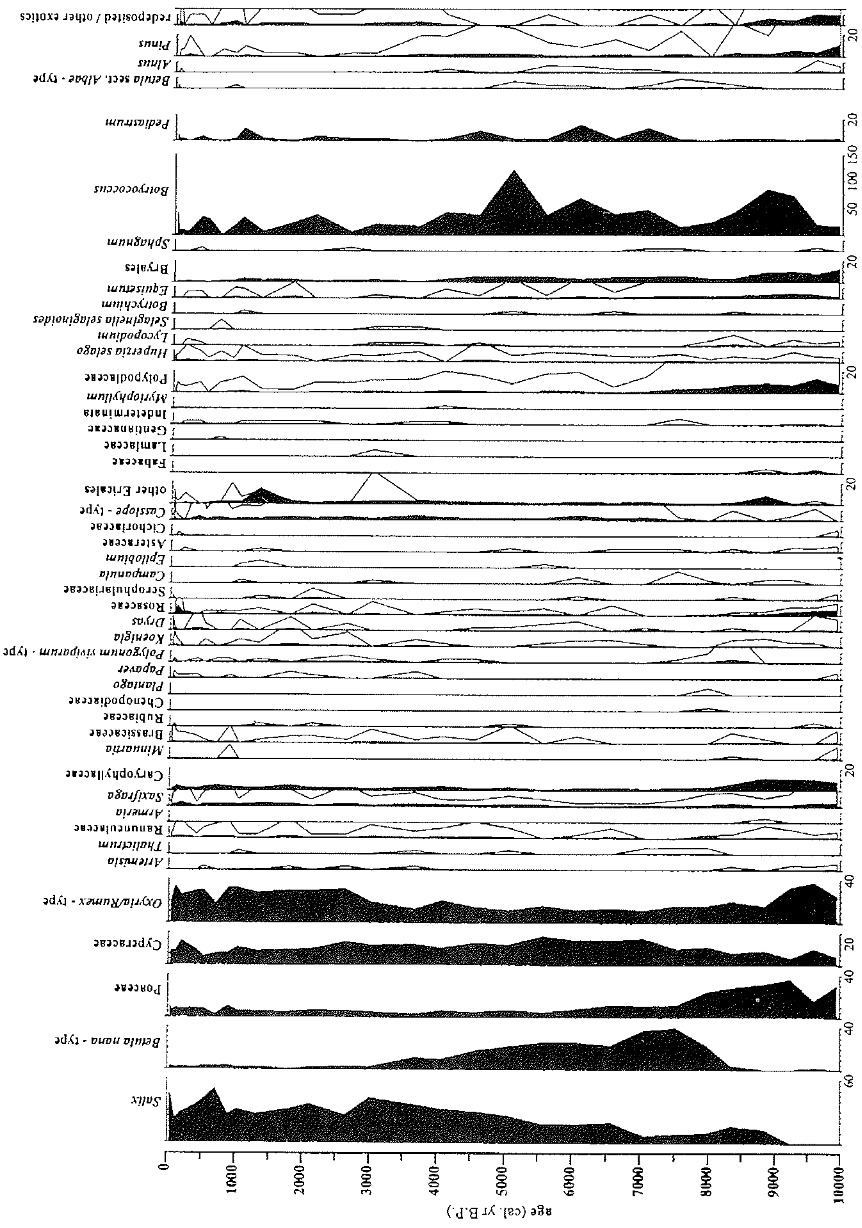


Fig. 4.8: Pollen percentage diagram from Basaltsø, plotted versus sediment ages.

opal, and of pollen first takes place at about 10,000 cal. yr B.P. This increase post-dates the temperature rise since the Preboreal oscillation that is indicated by the shift in $\delta^{18}O$ values in the Renland ice core by ca. 1000 years (Fig. 4.10). Usually, lakes are rapidly reacting systems on environmental changes due to the short lifetime of most aquatic organisms, their high reproduction rate, and their easy transportation by wind or birds over great distances. Hence, the adjustment of the lake equilibrium with the climatic conditions may have been delayed by environmental processes in the catchment.

One reason for delayed biogenic accumulation may be the presence of dead ice or extensive snowfields in the surrounding of the lakes, which may have persisted after the general deglaciation (Hjort, pers. commun.). This may have led to cooling and light limitation due to turbid melt water supply and thus hampered biogenic production. The occurrence of melt water sources in the region is documented by a distinct, fine lamination at the onset of the limnic facies. Most laminae in core PG1205 have thicknesses of less than 1 mm (Fig. 4.7), and are formed by couplets with light, silt-sized basal and dark, clay-sized top layers. This is typical for glacial varves (Leemann et al. 1991), but could also derive from erosion processes of melting snowfields that may result in seasonal changes in the transport energy of brooks entering the lake (Larsen et al. 1998). An interpretation of the varve thicknesses with respect to climatic or glaciological variations, as presented by Leemann and Niessen (1994b) for Silvaplana Lake, Swiss Alpes, and by Leonard (1997) for Hector Lake, Canada, was not possible due to their partly indistinct type (Fig. 4.7). However, a decreasing frequency of thin

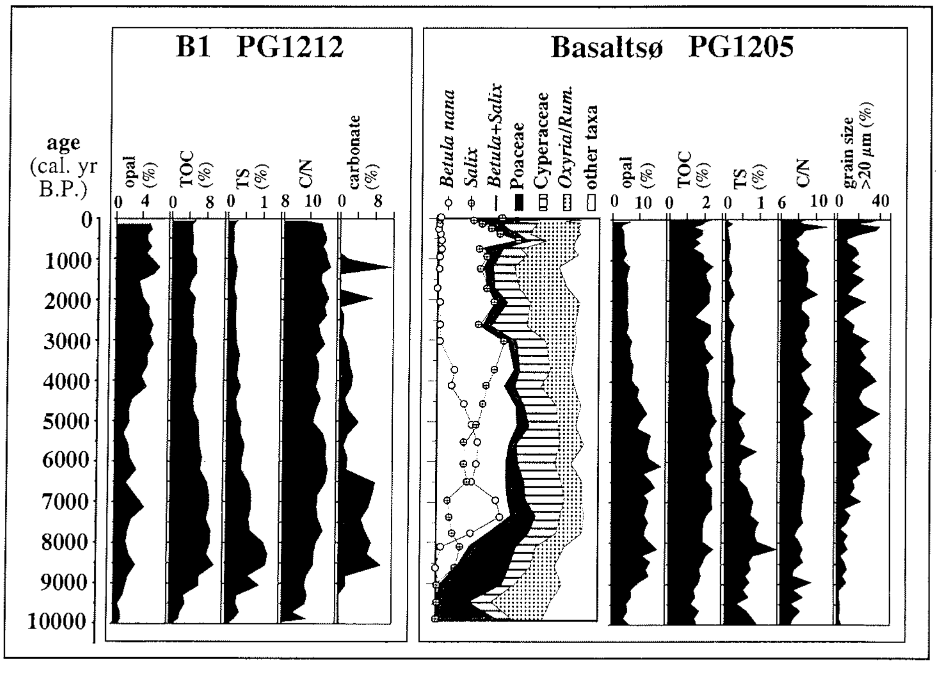


Fig. 4.9: Most important biogeochemical parameters in cores PG1205 and PG1212, and coarse-grained matter and pollen assemblage in core PG1205, plotted versus sediment ages.

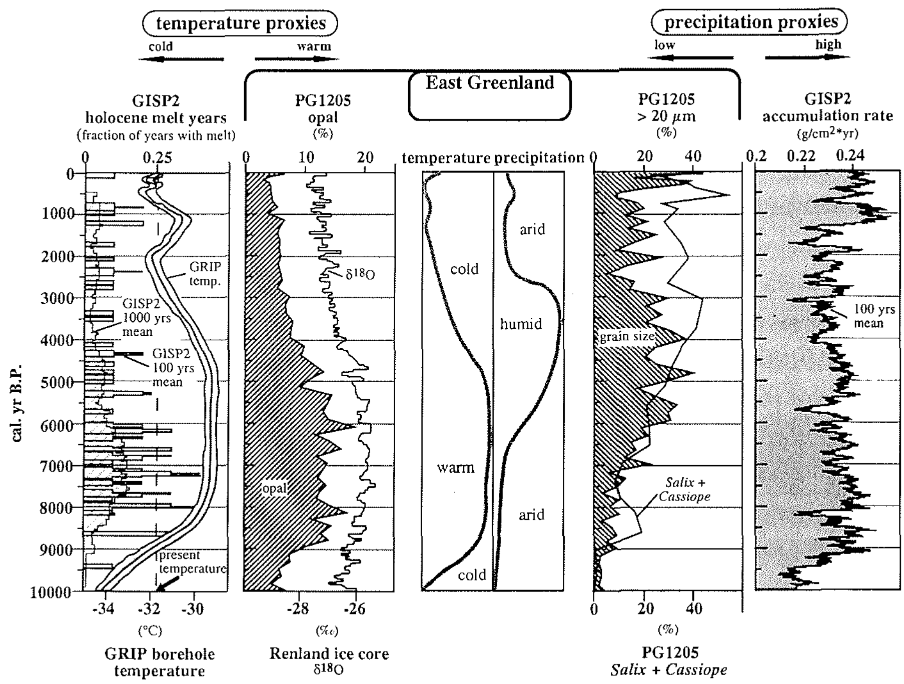


Fig. 4.10: Reconstruction of the temperature and precipitation history in East Greenland (central part of figure), as based on the most important climate proxies in the core PG1205. Latitudinal and longitudinal similarities and differences in the climate history of central and eastern Greenland are indicated by comparison of the Basaltsø record with the GISP2 melt year record (after Alley and Anandakrishnan 1995), the GRIP palaeotemperature record (after Dahl-Jensen et al. 1998), the Renland ice core isotopic record (after Larsen et al. 1995), and the GISP2 snow accumulation record (after Meese et al. 1994).

turbidites, along with an increase of biogenic accumulation (Figs. 4.4 and 4.5), indicates a gradual decrease of clastic sediment supply due to prograding diminution of dead ice or snowfields in the catchment.

Further reasons for the delayed reaction of the lakes to the climatic conditions may be a sparse vegetation in their surroundings due to a slow soil formation or a delayed immigration of landplants (Björck et al. 1994b). Even if aquatic organisms are capable of gaining nutrients from minerals for their bioproduction, the decomposition of detrital vegetation supplies nutrients on a large scale. In the sediments of Basaltsø, the pollen concentration is extremely low before 9900 cal. yr B.P. (Figs. 4.8 and 4.9). Until ca. 9000 cal. yr B.P., the pollen assemblages were characterized by only few taxa. A maximum of *Pinus* coincides with a high percentage of redeposited/exotic pollen grains. This suggests high wind velocities (Fredskild 1984), since *Pinus* did not occur in East Greenland (Funder 1978, Funder and Fredskild 1989, Björck et al. 1994a, 1994b, Fredskild 1995, Böcher and Bennike 1996). The vegetation cover was sparse, composed of Poaceae, Cyperaceae, Caryophyllaceae, Saxifragaceae, Rosaceae, *Oxyria* and Polypodiaceae, *Equisetum* and green mosses (Bryales). This pollen assemblage represents fell field vegetation, an association that is adapted to unstable soils, a high amount of soil erosion during snow melting, and problems concerning the water balance of soils due to the underlaying permafrost (Funder and Fredskild 1989). In northern Iceland, the development of a similar vegetation, with dominating *Oxyria*, Poaceae and Caryophyllaceae, characterizes a progressive closure of the vegetation cover during the early Holocene (Rundgren 1998).

The end of rising concentrations of most biogeochemical parameters at ca. 9000 cal. yr B.P., and a subsequent correspondence of some of them to climatic variations mirrored in ice cores from central and eastern Greenland (Fig. 4.10), imply that Basaltsø and lake B1 have since then been directly affected by climatic changes. The deglaciation of the catchment area has finished, and the vegetation in the surrounding of the lakes may supply sufficient nutrients for a rapid reaction of the lake organisms.

4.6.2. Early Holocene

Following deglaciation at the Pleistocene/Holocene transition, the remaining part of the early Holocene (9000-6500 cal. yr B.P.) is characterized by an increase of dwarf shrub pollen to a high level in core PG1205 from Basaltsø, and by a high amount of organic matter in both sediment cores (Fig. 4.9). Differences in the absolute values and variations of individual biogeochemical parameters within and between both cores indicate that the climate is not the only factor controlling their concentration. Important factors masking the climate signal could be differing morphologies of the lakes and their surroundings, a differing supply of organic and terrigenous matter, and differing diagenetic processes in the sediment columns.

Differences in the absolute values between the lakes are most pronounced for TOC and TN, which reach distinctly higher concentrations in the sediments in core PG1212 than in those of same age in core PG1205 (Figs. 4.4 and 4.5). The lower concentrations in Basaltsø can be explained by its larger catchment area with a steep south-facing slope north of the lake, where a thick snow cover accumulates in the winter, assisted by predominant northern winds (Fredskild 1998, Meltofte and Rasch 1998). This leads to a high meltwater inflow into the lake during summer, supplying a higher amount of minerogenic matter that dilutes the biogenic accumulation. Lake B1, in contrast, is located on a weakly inclined north-facing slope and has a much smaller catchment. In addition, the smaller and shallower lake B1 warms up in summer more rapidly than the larger and deeper Basaltsø, leading to increased lake bioproduction.

Despite the significant differences in TOC and TN values between the sequences, the C/N ratios of ca. 8-9 in Basaltsø and 9-11 in lake B1 show rather small differences. The C/N ratio reflects the relative amounts of autochthonous versus allochthonous organic matter on the bulk organic accumulation. In contrast to C/N ratios of between about 4 and 10 for non-vascular aquatic plants, terrestrial plants may have a ratio of more than 20 (Meyers and Ishiwatari 1995, Hassan et al. 1997). Hence, the C/N ratios in both cores indicate a predominance of autochthonous biogenic accumulation, slightly more pronounced in Basaltsø than in lake B1.

Distinct differences in the concentration patterns of the biogeochemical parameters throughout individual sediment successions indicate diagenetic processes. These processes in the two investigated lakes have influenced different parameters.

In core PG1205 from Basaltsø, biogenic opal and TS have a broad maximum during the early Holocene. Biogenic opal in limnic systems is predominantly formed by diatoms, whilst TS originates from the total buried biomass in the sediment. Hence, both parameters indicate a maximum in biogenic accumulation. However, this is matched neither in the C/N ratios nor in the TOC and TN concentrations, which remain rather stable throughout the entire Holocene (Figs. 4.4 and 4.9). The reasons for the depletion of TOC and TN remain uncertain, but a possible explanation is given by an enhanced postdepositional decomposing of organic matter by bacteria during the early Holocene, leading to the release of CO_2 and N_2 and subsequent fixation of sulphur as sulphide under strongly reducing conditions (Clark and Fritz 1997). The decomposition may particularly have affected the autochthonous matter (Meyers and Ishiwatari 1995), and thereby may have kept the C/N ratio rather constant.

In core PG1212 from lake B1, the early Holocene sediments are characterized by a maximum in TOC, TN and TS. This mirrors the high biogenic accumulation in this period and indicates a smaller bacterial decomposition. Another difference from Basaltsø is the minimum in the opal contents in core PG1212 in this period, which may derive from different diatom assemblages or opal dissolution. Indication for the latter comes from the negative correlation between the amount of opal and carbonate in this core. In contrast to the detrital carbonates in the glaciolimnic facies, whose $\delta^{13}C$ values of 0-3‰ indicate a marine origin, those of the limnic facies with values of -28 to -29‰ have to be of limnic origin, with methane as the carbon supplier (Clark and Fritz 1997; Fig. 4.5). The limnic carbonate suggests a shift in pH conditions towards higher values, which may have led to a dissolution of biogenic opal in lake B1. Geochemical measurements have identified the carbonates as siderite. The formation of siderite is strongly dependent on reducing conditions, caused by decomposition of organic matter in the sediment. In the sections of core PG1212 with most carbonate, reducing conditions may have been enhanced by an enrichment of more easily decomposable autochthonous matter that is reflected in lower C/N ratios. Additional evidence for strongly reducing conditions comes from the presence of vivianite concretions that occur in the sediments deposited about 8000 to 3000 cal. yr B.P., and in particular 8000 to 5000 cal. yr B.P. The formation of vivianite is dependent on the presence of dissolved iron in the interstitial water and a high amount of phosphorus supplied by concentrated organic matter.

Summarizing the discussed influences of lake external and internal factors on biogenic accumulation, the maxima, particularly in opal and TS in core PG1205 and in TOC, TN, and TS in core PG1212, indicate an early Holocene maximum in limnic biogenic production. Its association with a coinciding high allochthonous biogenic supply, indicated by the fairly constant C/N ratio, indicates high bioproduction in the lake catchment due to the establishment of a more dense vegetation cover in this period.

The vegetation succession is mirrored by the pollen assemblages in core PG1205. The palynological data show an increase in *Salix* pollen percentage between 9200 and 8800 cal. yr B.P. (Figs. 4.8 and 4.9). *Salix herbacea* was probably the only *Salix* species present in East Greenland at this time (Funder 1979, Björck et al. 1994a, Böcher and Bennike 1996). A subsequent depletion in *Salix* pollen percentages between 8300 and 7900 cal. yr B.P. coincides with a strong increase in *Betula* pollen, and with a gradual retreat of the fell field vegetation that was common in this area since 9900 cal. yr B.P. The maximum in *Betula* pollen was reached between 7800 and 6500 cal. yr B.P. This may reflect extensive spreading of birches in the catchment, though birches are known to produce high amounts of pollen and therefore may be overrepresented in the pollen record (Rundgren 1998). The delayed spreading of birches at Basaltsø, compared to the onset of biogenic sedimentation in the lakes is explained by its delayed immigration to East Greenland. *Betula nana*, coming from Europe via Scotland, the Faroe Islands, and Iceland, arrived in the region of Scoresby Sund at about 8900 cal. yr B.P. (Funder 1978, Fredskild 1991; Tab. 4.1) or even ca. 200 years earlier (Bennike and Funder 1997). The immigration of *Betula nana* is recorded by its arrival at Basaltsø, ca. 300 km further north, at 8300 cal. yr B.P. and by its first occurrence on Hochstetter Forland, another 300 km to the north, at ca. 7800 cal. yr. B.P. (Björck et al. 1994b; Tab. 4.1).

Today, the abundance of *Betula* in the outer coastal zone of East Greenland is restricted due to the maritime climate (Fredskild 1991, Fredskild and Mogensen 1998). In the continental climate of the interior fjord regions, in contrast, *Betula* settles on south-west-exposed slopes, which have only a thin snow cover in winter, and a long-lasting snow-free period, high temperatures, and dry conditions in summer (Fredskild

1991, 1998, Bay 1992). Hence, the early Holocene maximum of dwarf birches in the Basaltsø region indicates higher summer temperatures than today and rather dry conditions, which is supported by the slightly earlier established maximum of biogenic production in Basaltsø and lake B1.

A comparison of the early Holocene climate history in the Basaltsø region with published data from central and east Greenland supplies information concerning latitudinal and longitudinal differences in the climatic development. The concentration patterns of the temperature-significant biogeochemical parameters in both lake sediment cores, in particular of opal in core PG1205, the isotopic record of the Renland ice core in the interior Scoresby Sund region, the reconstructed palaeotemperatures from the GRIP ice core, and the GISP2 Holocene melt year record, both from central Greenland, fit very well (Fig. 4.10). The Renland and GRIP ice core records indicate 2-2.5°C higher temperatures than those of today for the early Holocene climatic optimum (Johnsen et al. 1992, Larsen et al. 1995, Dahl-Jensen et al. 1998), which suggests a similar warming amplitude in the Basaltsø region. Hence, longitudinal differences in the temperature likely exist in the absolute values, but cannot be detected in temperature trends and the temperature amplitude.

Latitudinal differences, in contrast, may have occurred at the onset of the early Holocene climatic optimum. This is indicated by a delay of ca. 1000 years between the onset of high biogenic accumulation in a lake on Hochstetter Forland (Björck et al. 1994b) and in Basaltsø. A similar delay exists in the vegetational development along the coast, not only in the immigration of birches but also in their maximum abundance. At the outer Scoresby Sund, *Betula nana* reached highest values at 8300 cal. yr B.P., being the dominant species in a rich dwarf shrub heath indicating warm conditions (Funder 1978; Tab. 4.1). In Basaltsø, the *Betula* maximum is recorded at 7800 cal. yr B.P., and on Hochstetter Forland at 7400 cal. yr B.P. (Björck et al. 1994b; Tab. 4.1). This latitudinal shift cannot exclusively be caused by the progressive settlement of birches, but must also be due to a delayed warming towards the north.

Both the early Holocene warming, and the latitudinal trend in its onset in East Greenland, could be due to a change in the oceanic and atmospheric circulation patterns. A broad instreaming of warm water masses into the North Atlantic at 10,000 cal. yr B.P. (Tab. 4.1), culminating about 2000 years later, was reconstructed by Koç et al. (1993). This caused a shift in the oceanic circulation pattern with a retreat of the marine Polar front towards the northwest. The resulting higher water temperatures during the early Holocene enabled the settlement of the bivalves *Mytilus edulis* and *Chlamys islandica* along the coast of East Greenland (Hjort and Funder 1974, Israelson and Buchardt 1991). *Mytilus edulis* is presently only found further south, whereas a relict population of *Chlamys islandica* is still present in this region. The early Holocene shift of the marine water front coincides with a northwards movement of the atmospheric circulation pattern (Koç et al. 1993), which results in predominant southern winds and thus in a supply of warmer air masses to East Greenland at ca. 8900 cal. yr B.P. (Böcher and Bennike 1996; Tab. 4.1).

4.6.3. Middle Holocene

The middle Holocene (6500-3000 cal. yr B.P.) is characterized by decreasing values of most biogeochemical parameters in both cores. Along with an increase in grain sizes in core PG1205 from ca. 6500 cal. yr B.P., that culminates between 5500 and 2800 cal. yr B.P. (Figs. 4.9 and 4.10), a change in the climatic conditions during the middle Holocene is indicated.

The pollen record shows a decrease of *Betula* pollen percentages between 7000 and 6500 cal. yr B.P. (Figs. 4.8 and 4.9). Passing a plateau, which lasts to ca. 5500 cal. yr B.P., a further decline leads to an almost complete disappearance of *Betula* pollen in the Basaltsø sequence at about 3000 cal. yr B.P. A simultaneous rise of *Salix* and *Cassiope* dominated poor dwarf shrub heaths with a maximum at 3000 cal. yr B.P. resembles the record from Jameson Land, Scoresby Sund (Funder 1978). This rise may be an artefact of the coinciding decline in *Betula* pollen percentages. However, as suggested by Funder (1978), the middle Holocene *Salix arctica* spreading could also be a climate response, in contrast to the earlier birch spreading that was more dependent on their date of arrival. *Salix arctica* and *Cassiope tetragona* arrived first in northwestern Greenland from North America and expanded, in contrast to *Betula nana*, southwards along the coast of East Greenland (Funder 1978, Funder and Fredskild 1989). Whilst the arrival of *Salix arctica* in the Basaltsø region cannot be clearly dated due to the difficult distinction between *Salix arctica* and *Salix herbacea*, that for *Cassiope* is dated to about 7300 cal. yr B.P. (Fig. 4.8).

Today, *Salix arctica* and *Cassiope tetragona* prefer sites in East Greenland where a thick snow cover in winter prevents their damage by deep frost (Fredskild et al. 1986, Bay 1992, Fredskild and Mogensen 1998), but causes a shorter vegetation period in summer. High pollen percentages simultaneous to that of *Salix* and *Cassiope* show the Cyperaceae, that comprise *Eriophorum* and *Carex* species, most of which prefer wet soils (Fig. 4.8). Moist conditions are also recorded in the high contents of *Botryococcus* and *Pediastrum*, both green algae, whose blooming indicates a greater influx of nutrients due to larger amount of snowmelt in the catchment area (Fredskild 1995). Hence, the indications for increasing winter snow cover, simultaneously with increasing summer soil moisture, may be due to increasing precipitation during the middle Holocene.

In addition to the vegetational development, a coinciding increase in grain size in core PG1205 supports the indication of increasing precipitation since about 6500 cal. yr B.P. (Figs. 4.9 and 4.10). The higher amount of coarse-grained sediments indicates an increased erosion in the catchment, probably the result of decreasing vegetation cover and a higher transport energy of streams due to the melting of larger snow fields and increased rain. The summer meltwater pulse forms the major part of the annual lake inflow (Retelle and Child 1996, Hasholt 1997). However, in addition, single high-amplitude rain events may have sporadical effects on the sediment input, because the permafrost prevents deep water penetration in soils (Bradley et al. 1996, Hardy 1996, Meltofte and Rasch 1998).

The increase in snow accumulation might be the result of the warm temperatures, and the glacier retreat into the inner fjords during the early Holocene climatic optimum (Funder 1989). Both enabled a high evaporation rate in summer due to open water conditions in the fjords. Supplementary snow accumulation may derive from increased wind activities, indicated by a rise of *Pinus* pollen in Basaltsø between 5500 and 3500 cal. yr B.P., simultaneous with a record from southern Greenland (Fredskild 1984). A transfer of snow masses would have affected in particular the steep slope to the north of Basaltsø, forming the lee side of predominant northern winds (Fredskild 1998, Meltofte and Rasch 1998). The distinctly different patterns of the precipitation proxies in the Basaltsø region from the accumulation rate of the GISP2 ice core (Meese et al. 1994; Fig. 4.10) indicate longitudinal differences in Greenland.

Summarizing, the climatic deterioration in East Greenland started at about 6500 cal. yr B.P., at this time primarily induced by a rise in snow accumulation rather than by a temperature decline. Since at least 5000 cal. yr B.P., cooling is indicated by the ongoing changes of the palynological assemblages and by the decrease of organic accumulation (Fig. 4.9) during times of constant sedimentation rates (Fig. 4.6). The three temperature minima postulated by Funder (1990), at ca. 5700, 4500 and 3200 cal. yr B.P. (Tab. 4.1), are not seen in the records from Basaltsø and lake B1. The lake records correspond more to the Renland isotopic record (Larsen et al. 1995) and to the GRIP palaeotemperature record (Dahl-Jensen et al. 1998; Fig 4.10), in which a gradual temperature decrease since 5000 cal. yr B.P. is indicated.

The reason for the middle Holocene cooling is probably a retreat of warm Atlantic water masses to the southern central part of the Greenland, Iceland and Norwegian Seas. Simultaneously, the East Greenland Current strengthened and induced both a southwards shift of the sea ice margin to the area off Scoresby Sund between ca. 5800 and 3300 cal. yr B.P. (Koç et al. 1993; Tab. 4.1) and a decrease or disappearance of the bivalves *Mytilus edulis* and *Chlamys islandica* in East Greenland (Hjort and Funder 1974). Also, the cooling is likely responsible for the renewed glacier advance in

Northeast Greenland at ca. 5800 cal. yr B.P. (Hjort 1997; Tab. 1), and in the Scoresby Sund region about 2000 years later (Funder 1989).

4.6.4. Late Holocene

The late Holocene (< 3000 cal. yr B.P.) is characterized in both cores by a progressive decrease in the amount of organic matter in fine-grained sediments, interrupted only by a coarser interval between 500 and 200 cal. yr B.P. in core PG1205 (Figs. 4.9 and 4.10).

The pollen record of core PG1205 shows very low contents of *Betula* pollen from 3000 cal. yr B.P. and a sharp decrease in the *Salix* pollen percentages in the following 400 years (Fig. 4.8). From 2600 cal. yr B.P., the pollen spectra are characterized by high frequences of *Oxyria*, Caryophyllaceae, Cyperaceae, Saxifragaceae, and Poaceae. By that, it is similar to the fell field vegetation in the early Holocene, with the exception that *Salix* pollen occur at higher percentages. A very weak return of birches from 900 to 500 cal. yr B.P. is accompanied by a distinct increase in the *Salix* values at 650 cal. yr B.P. The following diminution in *Salix* values stops at about 100 cal. yr B.P., when a recent increase is observed. The small content of *Betula* pollen in the youngest sediments suggests a spreading of birches from their refugia areals, or wind transport from the inner fjord region.

A combination of the biogeochemical data with the palynological record indicates very cold and dry conditions from 3000 to 1000 cal. yr B.P. Probably, the cold conditions caused a freezing of the fjords also in summer and thus diminished the evaporation rate. Off East Greenland, a closed sea-ice cover is supported by a southwards movement of the Polar front due to a restricted influence of warm water masses to the Greenland, Iceland and Norwegian Seas from ca. 3300 cal. yr B.P. (Koç et al. 1993; Tab. 1). The palaeotemperature record from the GRIP borehole documents the shift in the North Atlantic circulation pattern by decreasing temperatures until 2000 cal. yr B.P. (Dahl-Jensen et al. 1998; Fig. 4.10).

The maximum of dwarf shrub pollen in Basaltsø between about 900 and 500 cal. yr B.P. indicates a short warming. This maximum is slightly delayed to the so-called medieval warm period at 1000 cal. yr B.P. that is very vaguely recorded in the Renland isotopic record (Johnsen et al. 1992), but well expressed in the palaeotemperature record from the GRIP borehole, indicating temperatures 1°C warmer than today (Dahl-Jensen et al. 1998).

The following Little Ice Age cooling period is reflected by lowest values of most biogeochemical parameters in Basaltsø and lake B1, starting at about 800 cal. yr B.P. Simultaneously, *Salix* and *Cassiope* pollen reach high percentages. A comparable reaction of the vegetation took place in the middle Holocene, when a spreading of *Salix* and *Cassiope* coincided with climatic deterioration. In contrast to the middle Holocene, the spreading of these taxa at the beginning of the Little Ice Age was not associated with an accumulation of coarser sediments (Fig. 4.10). Hence, the delayed peak in grain-size distribution between 500 and 200 cal. yr B.P. has to be affected not exclusively by an increase of snow accumulation. One explanation might be given by a nival basin record in the Zackenberg region (Fig. 4.1). There, Christiansen (1998) suggested that lower temperatures, and an increase of wind activity, at the beginning of the Little Ice Age led to expansive snowpatches at the lee sides of hills and in depressions. Erosion of the vegetation cover due to rising northern wind activities started at about 450 cal. yr B.P., and caused increased soil erosion. The maximum of niveo-aeolian activity was measured between 400 and 260 cal. yr B.P., simultaneous to the rise of coarse-grained matter in Basaltsø. This suggests that the increase of niveo-aeolian activity was not a local phenomenon.

Cool conditions in East Greenland between 500 and 100 cal. yr B.P. are also mirrored in the $\delta^{18}O$ record of the Renland ice core (Johnsen et al. 1992; Fig. 4.10). A simultaneous temperature decrease is indicated in central Greenland, in the palaeotemperature record from the GRIP borehole (Dahl-Jensen et al. 1998), and northeastern Greenland, in a palynological record (Fredskild 1995). The cooling is traced back to a reduced summer insolation (Overpeck et al. 1997).

An increase in the dwarf shrub pollen percentage and a weak rise in most biogeochemical parameters at the sediment surface in both lakes indicate a recent warming. Many of the warm-adapted plants, however, remain in their small refugia areals in East Greenland (Fredskild et al. 1986, Fredskild 1998). This is probably due to the particular cool and moist climate at the outer coastal region of East Greenland, being a consequence of frequent fog formation (Bradley et al. 1996, Funder et al. 1998).

4.7. Conclusions

From a multi-disciplinary geoscientific investigation of two sediment cores from two lakes on southeastern Geographical Society Ø, East Greenland, the following conclusions can be drawn concerning (1) the last glaciation and deglaciation, (2) the usefulness of the proxies for palaeoclimatic reconstructions, (3) the Holocene climate development, and (4) longitudinal and latitudinal similarities and differences in the climate history.

The last glaciation of Geographical Society Ø probably took place during the Milne Land stade (11,300-11,150 cal. yr BP; Björck et al. 1997). The deglaciation of the area was associated with a high sediment supply to the lakes until 10,000 cal. yr B.P.

The opal concentrations in the sediment core from Basaltsø show an excellent correspondence to the $\delta^{18}O$ ratios in the Renland ice core, indicating that they best reflect the temperature development during the Holocene. In addition, variations in the pollen assemblage in this core, in TOC and TN in the core from lake B1, and in TS in both cores provide evidence for temperature changes. The precipitation development is best mirrored by the grain-size distribution and by the pollen assemblage in the core from Basaltsø.

The Holocene climatic variations are well reflected in the composition of the lake sediments formed after 9000 cal. yr B.P. Prior to that date, sedimentation was strongly influenced by glaciers, dead ice or snow fields in the catchments, leading to a high sediment supply and a local cooling effect. The early Holocene climatic optimum from 9000 to 6500 cal. yr B.P. was characterized by dry summers and temperatures higher than today. Climatic deterioration commenced at 6500 cal. yr B.P. with an increase in precipitation. High precipitation until 3000 cal. yr B.P. is accompanied by a cooling trend starting at 5000 cal. yr B.P. Cool and dry conditions lasted from 3000 to about 1000 cal. yr B.P., when the weak medieval warming took place. The Little Ice Age from about 800 to 100 cal. yr B.P. is characterized by cooling to the lowest level within the Holocene and, at its beginning, by slightly increased precipitation.

Longitudinal and latitudinal similarities and differences in the climate history of central and eastern Greenland are revealed from a comparison of the reconstructed history in the Basaltsø region with that reflected in published climate records. Similarities in the Holocene temperature development on Geographical Society Ø, on the Renland ice cap, and on the central Greenland ice sheet demonstrate the longitudinal coincidence of temperature changes. In contrast, differences in the precipitation development in these areas are indicated, reflecting changes in the atmospheric and oceanic circulation. Latitudinal differences along the East Greenland coast occur in the chronology of climate changes, best reflected in a northward delay of the Holocene climatic optimum, probably due to changes in the oceanic circulation.

5. Holocene climatic and oceanic changes at East Greenland – evidences from seabird affected lake sediments on Raffles Ø

Bernd Wagner & Martin Melles

(**Boreas,** submitted, revised version)

5.1. Abstract

A 3.5 m long sediment sequence from a lake on Raffles Ø, off Liverpool Land, East Greenland, was investigated for chronology, lithology, palynology, and biogeochemistry. Radiocarbon dating of plant remains and the lithology of the sediment succession indicate continuous sedimentation since deglaciation of the area prior to 10,000 cal. yr B.P. The postglacial palynological record shows little variation and evidence of a wind transported pollen supply, both resulting from the geographical characteristics of Raffles Ø. Significant variations in the biogeochemical data reflect changes in aquatic bioproduction. These changes depend to some extent on climatic changes; however, they are mainly due to variations in seabird breeding colonies in the catchment which influence nutrient and cadmium supply to the lake. Large seabird breeding colonies were present between 7500 and 1900, from 1000 to 500, and since ca. 100 cal. yr B.P. Their absence prior to 7500 cal. yr B.P. is most likely the result of unsuitable feeding conditions close to Raffles Ø caused by a too dense or too open sea-ice cover. In contrast, between 1900 and 1000 and from 500 to 100 cal. yr B.P., the seabird settlement probably was restricted by an insufficiently long breeding season due to cold climate conditions in East Greenland.

5.2. Introduction

The climate of East Greenland during the Holocene is influenced by changes in the atmospheric and oceanic circulation patterns. Archives for these changes are the ice cores from the central ice sheet and the coastal Renland ice cap (e.g., Johnsen et al. 1992, Dansgaard et al. 1993, Meese et al. 1994, Dahl-Jensen et al. 1998), and marine sediment cores (e.g., Marienfeld 1991, Koç et al. 1993, Nam 1997, Notholt 1998). The onshore studies in East Greenland, functioning as a link between the ice core and the marine records, so far focussed on the geomorphology and the lithology, palynology, biogeochemistry and macrofossils of lake sediments (Funder 1978, Björck et al. 1991, Bennike et al. 1994, Björck et al. 1994a, 1994b, Bennike 1997, Wagner et al. in press).

Another terrestrial approach to reconstruct the palaeoenvironmental history of coastal polar areas is based on the historical settlement of birds. Bird settlements were successfully reconstructed in Antarctica, for instance by a nutrient and heavy metal enrichment in the sediments of Lake Boeckella, Antarctic Peninsula region (Zale 1994). Historical seabird populations were also reconstructed from ornithogenic soils, and from fossilized stomach oil at the nesting sites (Baroni and Orombelli 1994, Verkulich and Hiller 1994, Hiller et al. 1995). These studies revealed that the occurrence and the size of seabird colonies depend on the accessibility of breeding sites, primarily a result of the climatic conditions, and of feeding areas, primarily affected by the sea-ice cover. Thus, the historical settlement of birds combines information about atmospheric and oceanic changes. In Greenland, little is known about the historical breeding sites of birds, despite a good knowledge of their recent distribution (e.g. Norderhaug 1970, Meltofte et al. 1981, Nettleship and Evans 1985, Elander and Blomqvist 1986, Mehlum 1989, Isaksen and Bakken 1995, Mehlum and Isaksen 1995, Falk et al. 1997).

The aim of this study is to reconstruct the Holocene settlement history of seabirds on Raffles Ø, East Greenland, a small island off Liverpool Land that is known for the high abundance of modern seabird breeding colonies (Nettleship and Evans 1985). For this purpose, a sediment record from a lake on Raffles Ø was studied by a multi-disciplinary approach, including radiocarbon dating of terrestrial plant remains, palynological and biogeochemical investigations. A comparison between the sediment data with published data from ice-core, marine and terrestrial records supplies information on historical seabird populations on the island, which are discussed in the light of the known Holocene climatic and oceanic changes off East Greenland.

5.3. Geographical setting

The sediment core investigated in this study was recovered from a lake on Raffles Ø, an island situated close to the mouth of the Scoresby Sund in front of Liverpool Land, East Greenland (Fig. 5.1). The unnamed lake, here referred to as Raffles Sø, has a length of about 1 km and a width of about 300 m (Fig. 5.2). The lake is located at an altitude of about 40 m a.s.l. in a cirque that exhibits steep slopes up to 550 m a.s.l. in all directions except the southwest. This results in a very small catchment, estimated to cover an area between 1.0 and 1.6 square kilometers. The main inflow to Raffles Sø today originates from a perennial snowfield to the northeast of the lake. The outflow is situated at the opposite lake shore.

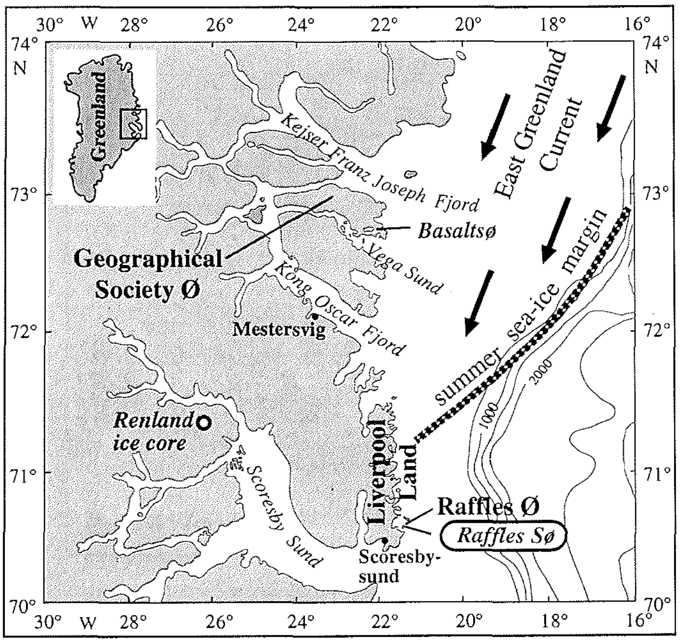


Fig. 5.1: Map of East Greenland showing Raffles Sø on Raffles Ø in front of Liverpool Land, the recent position of the summer sea-ice margin (Koç et al. 1993), the coring location of Renland ice core, and geographical terms mentioned in the text (settlements = black dots).

A bathymetric survey of the lake bottom, conducted by single-point measurements with an echosounder through ice holes, revealed a maximum water depth of 63 m close to the northeastern lake shore (Melles et al. 1995). Towards the southwest, the lake bottom gently rises after crossing a low ridge in the central part (Fig. 5.2).

The climate on Raffles Ø depends on the atmospheric circulation, and moreover on oceanic circulation patterns. The East Greenland Current transports cold (< 0°C) and low-salinity (34.4 ‰) polar waters from the Fram Strait southwards along the coast of East Greenland (Swift and Aagaard 1981). This leads to the present location of the average summer sea-ice margin in front of Scoresby Sund (Hebbeln et al. 1998; Fig. 5.1). The cold water of the East Greenland Current causes temperature inversions along the coast and a frequent formation of fog, which hampers land warming by insolation. Thus, temperatures of 3°C during the warmest month are recorded at the Scoresbysund settlement (Funder 1989) close to Raffles Ø, whilst further inland temperatures are higher (Harpøth et al. 1986).

Precipitation in East Greenland is mainly supplied by maritime air masses, which follow cyclone tracks along the coastline from southwest to northeast (Funder

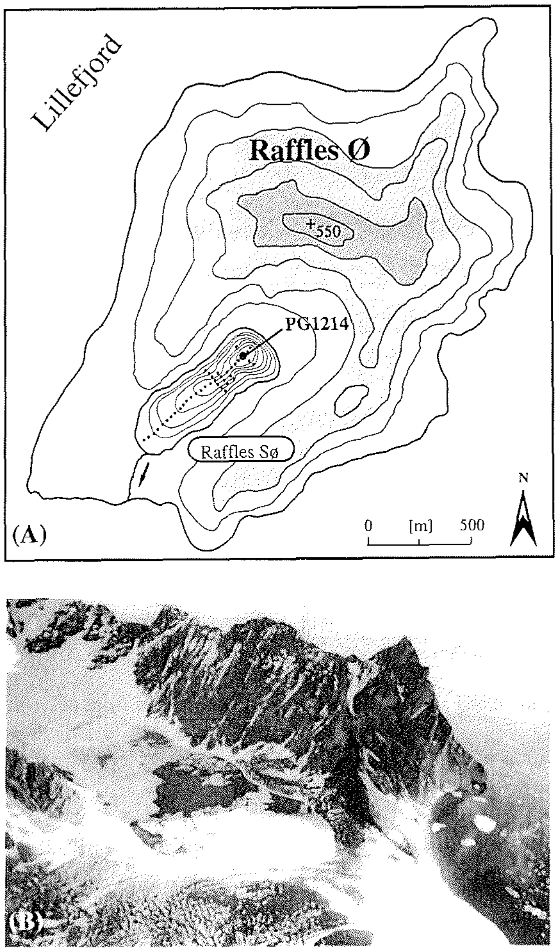


Fig. 5.2: (A) Topographic map of Raffles Ø with bathymetric contours (in 10 m intervals) of Raffles Sø, based on single point measurements (small dots). The heavy dot marks the coring position. (B) Photograph of Raffles Sø (view from the west), taken from a helicopter, reveals the steep slopes of the cirque in which Raffles Sø is embedded and its outlet in the front.

1989). In consequence, humidity decreases both from the south to the north and from the coast to the interior (Reeh 1989). The mean annual precipitation at Scoresbysund amounts to 550 mm/yr (Funder 1989). Precipitation is predominantly during the winter, but occasional rain events and cold spells with snowfall are also common in the summer months (Meltofte et al. 1981, Meltofte and Rasch 1998).

Raffles Ø is formed from massive and hornblende-bearing migmatites of Proterozoic-Caledonian origin. It is sparsely covered by vegetation. Despite several observations of seabird breeding colonies at the mouth of Scoresby Sund and on the islands off Liverpool Land, including Raffles Ø (Nettleship and Evans 1985, Mehlum

1989, Falk et al. 1997), no populations were on the island at the end of September 1994, which is very late in the season. Nevertheless, the occurrence of two arctic foxes at that time indicates feeding conditions sufficient to survive the summer period, when the island is separated from Liverpool Land by open water.

5.4. Material and methods

The coring position is located in the deepest part of the lake. Coring was carried out in summer 1994, through holes in the lake-ice cover (Melles et al. 1995). The sediments were recovered with two drives of a piston corer (UWITEC Corp.), comprising 0-230 cm (PG1214-3) and 106 cm below that (PG1214-4). The coring depth is controlled by the release of a piston that is fixed at the lower end of the corer during its way through the water column and, for deeper cores, the overlaying sediments. A more detailed description of the coring technique is given by Melles et al. (1994). The cores were cut into pieces of up to 1 m length, and stored at + 4°C prior to their opening in the laboratory in 1997.

For opening, the PVC core tubes of 6 cm diameter were scratched along their axis at two opposing sides by an electrical saw, fully cut by a knife in order to avoid pollution of the sediments by shavings, and finally divided into two halves with a nylon fishing line. Core description and photographic documentation were carried out immediately after core opening. One core half was then used for continuous subsampling in 2 cm intervals, whilst the other core half was kept complete and stored as archive for possible future work. The 2 cm thick subsamples were freeze-dried, and their water contents (% of wet bulk sediment) calculated from the mass differences between the wet and dry samples.

For the biogeochemical analyses, an aliquot of the sediment subsample was ground (to < 63 μm) and homogenized using a planet mill. The major part of the analytical measurements was conducted in intervals of 4 cm in the uppermost 230 cm, and in intervals of 8 cm in the deeper sediments. The contents of total carbon (TC), total nitrogen (TN) and total sulphur (TS) were measured with a CHNS-932 determinator (LECO Corp.). Total organic carbon (TOC) was analyzed with a Metalyt-CS-1000-S (ELTRA Corp.) in corresponding samples, which had been treated with HCl (10 %) at a temperature of 80°C to remove carbonate. Biogenic silica (opal) contents were measured according to the wet chemical method described by Müller and Schneider (1993). In addition, the contents of cadmium were determined in intervals of 4 to 24 cm with an inductively coupled plasma optical emission spectrometer (ICP-OES, Perkin-Elmer) after complete acid solution of the bulk sediment.

The palynological analyses were carried out by J. Hahne. For a qualitative and quantitative overview of the palynological inventory in core PG1214, about 1 g of dry bulk sediment was treated with HF (70 %) in intervals of 10 to 20 cm. Subsequently, pollen were further enriched by sieving (6x8 μm mesh). The determination of the absolute pollen values in each sample was enabled by adding a *Lycopodium*-tablet (12,548 spores). An exact description of the method is given in Hahne and Melles (1999). The amount of counted pollen grains per sample was more than 325 in the uppermost 225 cm, in deeper sediments the pollen concentration was too low for significant palynological investigations.

The palynological results are presented as total pollen diagrams, i.e. the pollen sum (100 %) includes all pollen, with the exception of aquatics, spores, and algae. *Betula* is assumed to represent *Betula nana*, the dominant taxon in East Greenland from the early Holocene (Fredskild 1991). Due to the difficult distinction between *Salix arctica* and *Salix herbacea*, both common in East Greenland during the Holocene, the willows are grouped to the term *Salix*. For the same reason *Oxyria* and *Rumex* were combined to the *Oxyria/Rumex*-type, though the majority of grains are believed to belong to *Oxyria*.

Radiocarbon dating was conducted on handpicked terrestrial plant remains by accelerator mass spectrometry (AMS) at the Van de Graff Laboratory, University of Utrecht. The material was separated under a binocular microscope, and mechanically cleaned from other sediment particles. The cleaning avoids contamination by coal particles, which are reported from lakes of East Greenland and which would cause an overestimation of the ^{14}C dates (Funder 1989, Björck et al. 1994a). By dating terrestrial plant remains the influence of reservoir effects in the lake water body on the radiocarbon dates is excluded. Fourteen samples from core PG1214 contained sufficient organic matter for AMS dating. All ^{14}C ages were calibrated into calendar years before present (cal. yr B.P.) using the bidecadal tree-ring dataset of Stuiver et al. (1998).

5.5. Results and discussion

5.5.1. Lithology and chronology

The base of the lower segment of core PG1214 is formed of unconsolidated coarse-grained sediments (Fig. 5.3). They indicate deposition of glacially supplied bedrock debris, and a depletion of the muddy matrix by flowing water. Thus, they probably originate from the deglaciation of the Raffles Sø cirque. The deglaciation may have taken place after the Milne Land stade or Preboreal oscillation, dated to 11,300-11,150

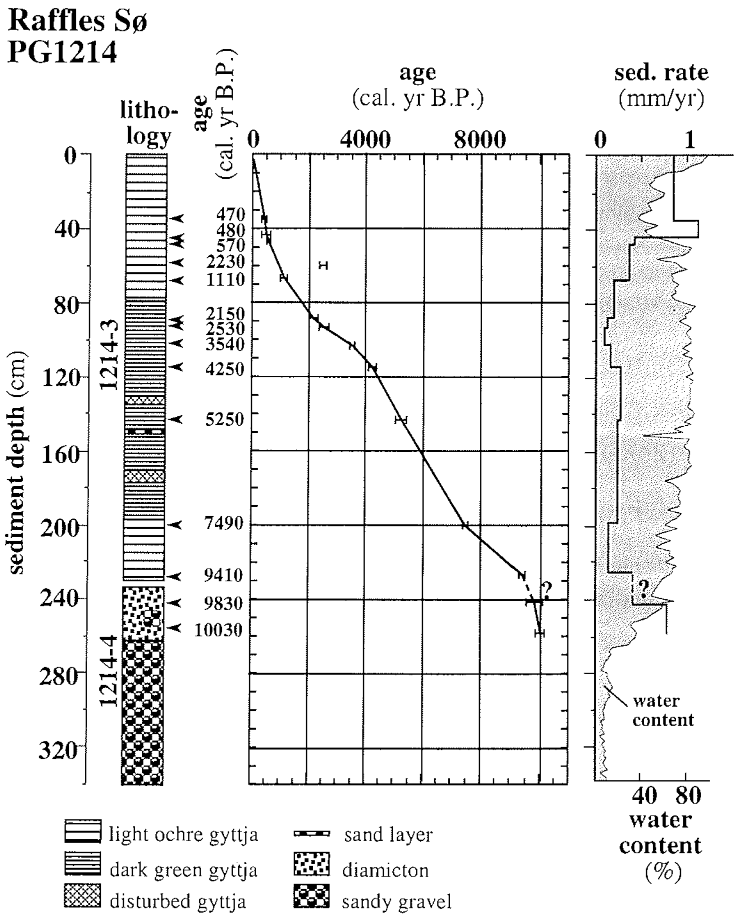
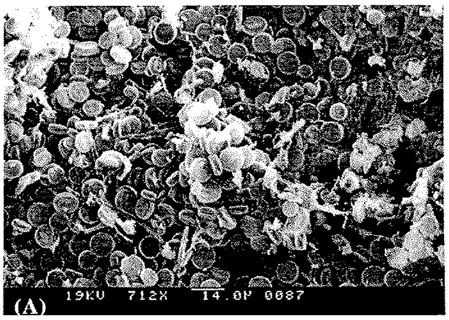


Fig. 5.3: Lithology and chronology at the coring site in Raffles Sø. Core PG1214 consists of two non-overlapping segments (PG1214-3 and PG1214-4) with an appoximated gap of 5 cm between both cores. Black arrows mark the horizons of radiocarbon dated plant remains. Linear interpolation between single datings, except the age of 2230 cal. yr B.P. in a depth of 59 cm, was used for calculation of the sedimentation rate, being negatively correlated to the water content.

cal. yr B.P., when the outer coastal region of East Greenland, explicitly Liverpool Land, was glaciated (Funder and Hansen 1996, Björck et al. 1997). In the uppermost 30 cm of the lower segment of core PG1214 an increase of the muddy matrix and a coinciding decrease of gravel resulted in a diamicton that is characterized by a poor sorting and weak stratification. The diamicton may reflect decreasing current velocities and the establishment of a lake in the cirque on Raffles Ø.



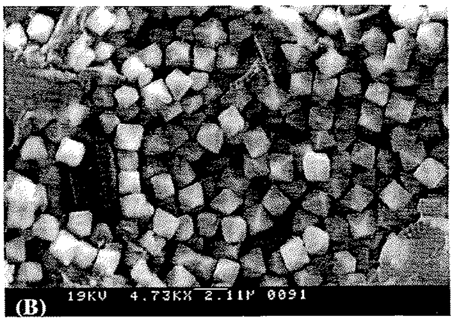


Fig. 5.4: SEM- photos of bulk sediment at a depth of 90 cm in core PG1214. The sediment is (A) extremely enriched in diatoms and depleted in terrestrial matter and (B) contains aggregations of framboidal pyrites are embedded in this sediment sequence.

The upper segment of core PG1214 (230-0 cm) consists almost completely of well stratified gyttja, formed by algal layers alternating with thin clayey and silty layers (Fig. 5.3). These sediments differ from both sediment types in the lower core segment, suggesting that the segments do not overlap each other. The gyttja in the upper segment shows distinct changes in colour from light ochre below 195 cm and above 80 cm to dark green between 195 and 80 cm. The horizontally bedded sediments are weakly deformed at depths of 177 to 170 cm and 135 to 132 cm. The gyttja, in particular in its lower part, contains single gravel grains and sandy layers of grey colour, reaching up to 2 cm thickness (at 150 cm depth). The gravel suggests ice-rafted transport, probably by ice floes with material frozen to the base or delivered to the surface at the shore. The thin sandy layers may be a result of melt-water events or turbidity currents. The coarse-grained terrigenous sediment supply plays only a small role in the gyttja formation. The predominance of fine grain sizes, and the high contents of organic matter, including diatoms (Fig. 5.4A) suggest calm conditions in the lake and a significant contribution of autochthonous matter to the bulk sediment accumulation.

The radiocarbon dates in both core segments show, except sample UtC-7421 (Tab. 5.1), increasing ages with increasing depths, and a maximum age of 10,030 cal. yr B.P. in the diamicton (sample UtC-7428; Fig. 5.3). The young age of sample UtC-7420 at a depth of 35 cm testifies that a recent reservoir effect on the dates can be excluded. The small age difference between sample UtC-8256 (9410 ± 120 cal. yr B.P.), 2 cm above the base of the upper core segment PG1214-3, and sample UtC-7463 (9830 ± 300 cal. yr B.P.), 9 cm below the top of the lower segment PG1214-4, suggests a small but unknown gap between both core segments. Assuming a gap of 5 cm, and neglecting sample UTC-7421 because of its obvious misplacement in the sediment, the resulting age model for the sediment sequence PG1214 in Raffles Sø indicates relatively high

Tab. 5.1: Radiocarbon and calendar ages determined on terrestrial plant remains isolated from cores PG1214-3 and PG1214-4, Raffles Sø. Uncertainties are reported at the 2σ level.

core	depth (cm)	material	sample no.	$\delta^{13}C$	^{14}C age (B.P.)	age (cal. yr B.P.)
1214-3	35	plant remains	UtC-7420	-24.4	366 ± 40	409 ± 93
1214-3	42-44	wood	UtC-7461	-28.0	470 ± 90	480 ± 166
1214-3	46-48	plant remains	UtC-7462	-24.3	530 ± 60	573 ± 77
1214-3	59	wood	UtC-7421	-28.9	2205 ± 38	2226 ± 101
1214-3	66-68	wood	UtC-7422	-27.7	1198 ± 36	1113 ± 119
1214-3	88	wood	UtC-7423	-27.7	2130 ± 40	2149 ± 153
1214-3	93	wood	UtC-7424	-28.0	2451 ± 37	2534 ± 176
1214-3	102-104	plant remains	UtC-7425	-25.9	3286 ± 41	3537 ± 95
1214-3	115	wood	UtC-7426	-29.1	3830 ± 42	4249 ± 156
1214-3	142-144	plant remains	UtC-7427	-26.1	4589 ± 43	5254 ± 203
1214-3	198-200	plant remains	UtC-8255	-27.4	6559 ± 38	7491 ± 72
1214-3	226-228	plant remains	UtC-8256	-27.7	8403 ± 46	9408 ± 117
1214-4	244	plant remains	UtC-7463	-26.7	8720 ± 100	9825 ± 295
1214-4	259-261	plant remains	UtC-7428	-26.7	8900 ± 50	10030 ± 162

sedimentation rates of more than 0.5 mm/yr below 244 cm and above 43 cm, and significantly lower rates in the intervening sediments. These sedimentation rates are negatively correlated to the water content (Fig. 5.3).

5.5.2. Vegetation

The pollen concentration in core PG1214 was sufficient for significant analyses down to a depth of 225 cm, which corresponds to about 9270 cal. yr B.P. (Fig. 5.5). Throughout the entire Holocene the species diversity is very sparse and stable. Poaceae, with 55-85 % of the total pollen sum, is the dominant taxon, accompanied by relatively frequent *Oxyria/Rumex*, Caryophyllaceae, Crassulaceae, Rosaceae and Cyperaceae.

The arboreal pollen concentration amounts to less than 10 % throughout the sequence. The maxima of *Alnus* and *Pinus* pollen from 9500 to 6500 cal. yr B.P. and from 4800 to 500 cal. yr B.P. indicate long distance transport of pollen grains, since both taxa did not grow in East Greenland (Funder 1978, Fredskild 1984, Björck et al. 1994a, Wagner et al. in press). The coinciding weak maxima of *Betula* pollen suggest that they also originate from long distance transport. The earlier maximum between 8500 and 5000 cal. yr B.P. corresponds to a period of general high abundance of *Betula nana* in East Greenland, indicative for warmer conditions than today (Funder 1978, Björck et al. 1994a, Wagner et al. in press). This may have led to a high influx of *Betula* pollen from Liverpool Land, only 2 km to the west of Raffles Ø. The second weak

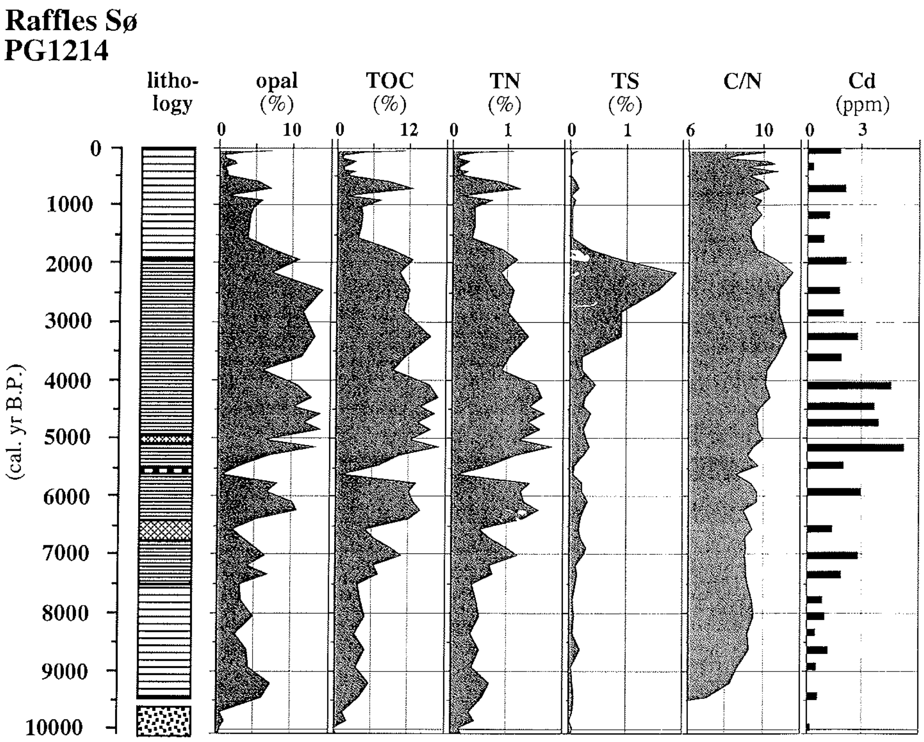


Fig. 5.6: Lithology and investigated biogeochemical parameters of core PG1214 versus age (cal. yr B.P.), for legend see Fig. 5.3.

The C/N ratio in bulk organic matter generally indicates changes between the relative input of autochthonous and allochthonous matter. In contrast to C/N ratios between about 4 and 10 of non-vascular aquatic plants, terrestrial plants may have a ratio of more than 20 (Meyers and Ishiwatari 1995, Hassan et al. 1997). However, the C/N ratio may also be affected by decomposition and dissolution processes. From 8500 cal. yr B.P. the C/N ratio in core PG1214 is relatively stable, varying between 8 and 10.5, which suggests that variations in the accumulation of autochthonous matter coincide with those of allochthonous matter. A weak maximum with values around 11 from 4500 to 1800 cal. yr B.P., during times when TS shows enrichment due to reducing conditions, may be a further indicator for decomposition and dissolution in the sediment.

The parameters opal, TOC, TN and cadmium in core PG1214 correlate well throughout the Holocene (Fig. 5.6). After an increase in the earliest Holocene they remain rather stable until ca. 7500 cal. yr B.P. Whilst the concentrations of opal, TOC and TN at this time are still relatively low, those of cadmium with values between 1 and

2 ppm are certainly on a high level in comparison to other sediments from unpolluted lakes (Håkanson and Jansson 1983). Nevertheless, interrupted by two short-term decreases at 6600 cal. yr B.P. and at 5600 cal. yr B.P. that both reflect grey sandy horizons, a further rise to more than 5 ppm takes place, when opal concentrations increase to more than 13 % and TOC to more than 17 % at 5200 cal. yr B.P. These high levels indicate a high enrichment of organic matter, but also an extraordinary pollution during this period. A subsequent decrease commences at ca. 4000 cal. yr B.P., more pronounced in the cadmium concentrations. Later, at 1900 cal. yr B.P., a sharp drop also in opal, TOC, and TN is mirrored by a sediment colour change from dark green to light ochre. The decrease leads to relatively low values comparable to those of the earliest Holocene, except for two intervals of higher concentrations at 1000 and 500 cal. yr B.P., and in the uppermost sediments deposited after 100 cal. yr B.P.

This Holocene biogeochemical lacustrine record from Raffles Sø is not conform with those from other regions in East Greenland, neither in timing nor in the amount of biogenic accumulation. Other lacustrine records from East Greenland reflect the establishment of an early Holocene climatic optimum at least 1000 years earliër than in Raffles Sø (Funder 1978, Björck et al. 1994a, Wagner et al. in press; Fig. 5.7). The delayed changes of Raffles Sø are confirmed by comparison with ice core records from central and eastern Greenland (Johnsen et al. 1992, Larsen et al. 1995, Dahl-Jensen et al. 1998), and by marine records (Hjort and Funder 1974). Whilst biogeochemical parameters in Raffles Sø reach a broad maximum at ca. 5200 cal. yr B.P., the other records indicate the end of the climatic optimum and a subsequent deterioration at that time. Only in the last 1900 cal. yr B.P. the Raffles Sø sediments mirror the general climatic development of East Greenland. Explicitly, the concentration patterns of opal, TOC, and TN indicate a short medieval warming around 1000 cal. yr B.P. and a subsequent Little Ice Age (Fig. 5.7).

Besides the timing, also the amount of biogenic accumulation in the Raffles Sø sediments, in particular in those sediments deposited when the early Holocene climatic optimum is weakening in East Greenland, is unexpected. The contents of TOC in Raffles Sø sediments are comparable to the loss-on-ignition values of other lakes from East Greenland (Funder 1978), however, the gentle surroundings and shallow water bodies of these lakes create much better conditions for the biological production and accumulation. This was revealed, for instance, by a comparison of the sediment records in the 21 m deep Basaltsø and the 9 m deep Lake B1 (Wagner et al. in press). Raffles Sø with a maximum water depth of 63 m is much deeper, and surrounded by steep, stony and shading slopes, which today are only sparsely covered by vegetation and partly snow-covered even during summer. Nevertheless, the maximum contents of biogenic opal are similar to those in Basaltsø, and the contents of TOC are even distinctly higher

(Fig. 5.7). This, and the delay in timing indicate that the high biogenic accumulation in Raffles Sø must have been partly controlled by non-climatic factors. A likely explanation is the input of nutrients into the lake due to the occurrence of seabird breeding colonies in the catchment of Raffles Sø, which are also known to contaminate sediments with heavy metals as, for instance, cadmium (Zale 1994). This may also apply in historical times.

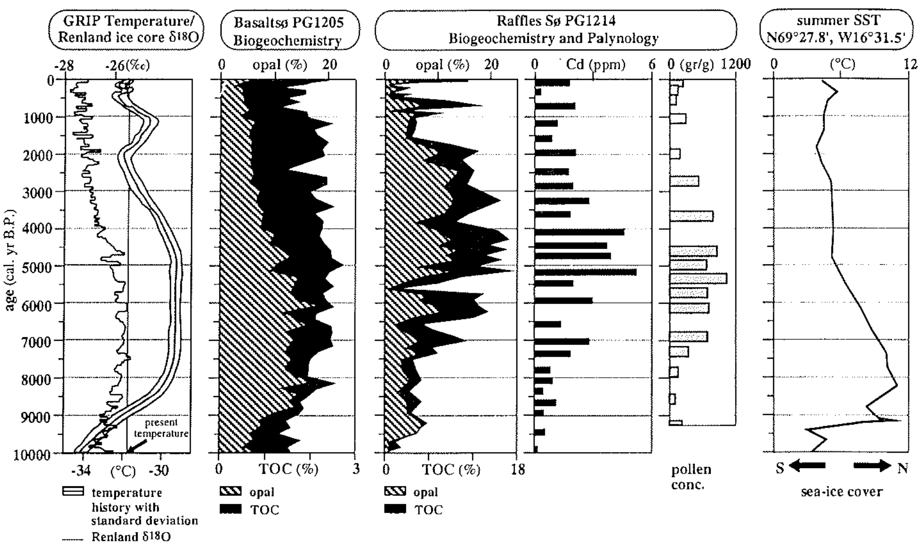


Fig. 5.7: TOC, opal, cadmium and pollen contents in core PG1214 from Raffles Sø for the last 10 000 cal. yr B.P., compared with the TOC, opal contents in core PG1205 from Basaltsø (Wagner et al. in press), the GRIP palaeotemperature record (Dahl-Jensen et al. 1998), the isotopic record of the Renland ice core (Larsen et al. 1995), and the reconstructed summer sea-surface temperatures off East Greenland (after Koç et al. 1993, ages converted into cal. yr B.P. using the marine calibration dataset of the CALIB4.0 program of Stuiver et al. 1998).

5.5.4. Ecology of seabirds

The islands in the North Atlantic are important breeding areas for seabirds. Most observations of breeding populations in East Greenland are summarized in Nettleship and Evans (1985). The Little Auk (*Alle alle*) is thought as the most common seabird in the North Atlantic today. More than 500,000 pairs were observed in the Scoresby Sund region and along the coast of Liverpool Land. Their breeding areas are within boulder screes on mountain slopes close to the sea, conditions as found on Raffles Ø. The Thick-Billed Murre (*Uria lomvia*) has three breeding colonies in East Greenland; one, with about 3000 pairs, is located on Raffles Ø. The Black Guillemot (*Cepphus grylle*) is common at Scoresby Sund with 2000 pairs, and along the outer coast of Liverpool Land with about 2500-10,000 pairs. A breeding population of the Atlantic Puffin (*Fratercula arctica*) on Raffles Ø was observed at the begin of this century, but not recently, though the bird is very frequently found in northwestern Iceland. Colonies of Fulmars (*Fulmarus glacialis*) are present in northeastern Greenland, and at the mouth of Scoresby Sund (Falk et al. 1997). The Kittiwake (*Rissa tridactyla*) commonly breeds on Spitsbergen. In East Greenland it has its northermost distribution area at the Scoresby Sund (Mehlum 1989).

All these species have an arctic or subarctic distribution. They are adapted to cool conditions, e.g. a short breeding and nesting period in response to only a few weeks of snow-free conditions on land and ice-free conditions at sea during summer (Bédard 1985). For example, the Little Auk arrives at the breeding sites in April. The laying of eggs takes place in the second half of June or early July, depending on snow melt. The incubation period is 29 days and the young leave the nests when they are 27 days old (Norderhaug 1970, Stempniewicz 1980, Isaksen and Bakken 1995). The Murres, which have an incubation period of 32 days, leave the nests after 21 days, but in contrast to the Little Auk, the fledging weight of the young Murres relative to the adults is very small (Birkhead and Harris 1985). A minimum of eight weeks of ice-free conditions is thus required by the Murres for their breeding cycle, and for other species even longer (Nettleship and Evans 1985).

The selection of a breeding area is dependent on the accessibility of suitable breeding sites within range of an adequate food supply (Nettleship and Evans 1985, Birkhead and Harris 1985). The maximum distance to the feeding area is ca. 100-150 km for Little Auks (Isaksen and Bakken 1995), and less for the other seabirds (Bradstreet and Brown 1985). The best feeding areas are mixture zones of cold polar waters with warm Atlantic waters, in the pack-ice, and at the sea-ice margins. In these areas, the marine bioproduction in spring and summer is very high, and wind-induced upwelling along the ice margin may additionally enlarge the amount of plankton, in particular in the upper 20-30 cm of the sea water (Mehlum and Isaksen 1995). The ice-marginal phytoplankton is grazed by zooplankton, which itself is fed upon primilary by fishes. Directly under the sea surface, zooplankton and fish are available for diving seabirds. The Little Auk, whose food consists almost exclusively of zooplankton, prefers about 50 % ice cover in its feeding area, as do the other seabirds, most of which are fish feeders (Isaksen 1995, Mehlum and Isaksen 1995, Falk et al. 1997). Changes in the water temperature may cause a shift in the range of some plankton or fish species and a subsequent alteration of predator-prey relationships (Evans and Nettleship 1985).

The period of ice-free conditions on the ocean off East Greenland is dependent on the intensity of the East Greenland Current, and on the air temperature. Changes in both may affect seabird breeding at coastal sites. For instance, prolonged suitable feeding conditions on the ocean allow a longer food supply, and increased air temperatures lengthen the accessibility of breeding sites; both may enlarge the number of breeding species (Birkhead and Harris 1985).

The turn-over rate of nutrients in a breeding seabird colony is immense. According to Norderhaug (1970) during just four summer weeks a colony of ca. 100,000 pairs of the Little Auk transports about 70 tons of plankton from the feeding area to the nestlings in the breeding area. A study from Svalbard revealed for the same colony size a consumation of about 30 tons of fresh zooplankton per day, and the subsequent production of ca. 3 tons guano per day (Gabrielsen et al. 1991). Another study at the Northeast Water Polynya in northeastern Greenland demonstrated that the five most abundant seabird species consumed an estimated total of 243 tons during their breeding season. The total transfer of carbon to the breeding population of all surface feeders in the Polynya amounted to about 27 tons (Falk et al. 1997). This indicates the enormous role of seabirds on nutrient supply into the catchment area of Raffles Sø today, inferring that nutrient supply to Raffles Sø by seabirds had also an impact in earlier times. The extraordinary enrichment of cadmium in the Raffles Sø sediment supports this hypothesis, since particularly zooplankton and, in consequence, the zooplankton feeding seabirds and their guano are known to accumulate high amounts of heavy metals (Frank 1986, Zale 1994, Wenzel and Gabrielsen 1995, Hawke et al. 1999). Thus, periods of high biogenic production in the lake could match times of bird settlement.

5.5.5. Climatic and oceanic implications

As illustrated by the comparison of core PG1214 from Raffles Sø with a core from Basaltsø (Fig. 5.7), changing concentrations in the amount of organic matter in the sediment of Raffles Sø are not only a response on climatic trends, but also influenced by changing occurrences of seabirds during the Holocene. The occurrence of seabirds depends on the climate, which affects the period of snow-free breeding sites, and on the sea ice-conditions, which affects the distance to the ice marginal feeding areas. Therefore, changes in the enrichment of organic matter in the Raffles Lake sediments originating from the nutrient input of seabird colonies point to both climatic changes and changes in the position of the sea-ice margin.

Between about 9500 and 7500 cal. yr B.P., the rather stable and low content of organic matter and cadmium in the sediment of Raffles Lake implies that extensive seabird colonies were still absent. The presence of seabird breeding colonies on Raffles Ø today indicates a snow-free period at the breeding sites of at least two month in summer, and we expect it to be longer during the early Holocene climatic optimum, when conditions were warmer than present. Therefore, a probable explanation for the absence of seabirds is an unsuitable feeding area around Raffles Ø. This may be caused by either a too closed sea-ice cover or too open water conditions around the island. A widely closed sea-ice cover until about 9200 cal. yr B.P. is suggested by the reconstructed low summer sea-surface temperatures at a position that is today located at the eastern margin of the East Greenland Current (Koç et al. 1993; Fig. 5.7). At 9200 cal. yr B.P., a distinct shift in the sea-surface temperatures towards warmer conditions suggests a weakening of the East Greenland Current intensity, leading to a summer sea-ice margin far north of Raffles Ø. This would have also created poor feeding conditions around Raffles Ø, and therefore refuse the extensive settlement of breeding seabird colonies.

Increasing pollen concentration and a rise in organic matter and cadmium content in the lake sediment at ca. 7500 cal. yr B.P. indicate the establishment of the earliest significant breeding populations (Figs. 5.5, 5.6 and 5.7). The higher pollen contents, reflecting a denser vegetation on Raffles Ø, can hardly be explained by just enhanced soil formation in consequence of prevailing warm conditions. Firstly, soil formation is hampered by the rough environment with steep and stony slopes on the island. Secondly, enhanced soil formation would have supported the settlement of dwarf shrubs, whose pollen are not enriched in the palynological record of Raffles Sø though being common in East Greenland at that time (Funder 1978). Therefore, the increase of fell field vegetation on Raffles Ø at ca. 7500 cal. yr B.P. likely was favoured by a warm climate, but initiated by enhanced nutrient supply from seabird excrements (Norderhaug 1970). A gradual amelioration of the feeding conditions for seabird colonies on Raffles Ø from this time could have taken place due to a new intensification of the East Greenland Current. This may be reflected in the decreasing sea-surface temperatures in the western North Atlantic (Koç et al. 1993; Fig. 5.7) which may have led to an increase of a mixture of open water and pack-ice around Raffles Ø, feeding conditions as preferred by most seabirds.

The two sharp collapses in the amount of organic matter and cadmium at 6600 cal. yr B.P. and at 5600 cal. yr B.P. (Fig. 5.7), in particular the latter, are probably artefacts of mass movement processes from lateral deposited sediments. This is inferred from the lithological changes that indicate turbidites by the deposition of sandy layers with a fining-upwards trend in the corresponding horizons. Additionally, short-term fluctuations in the deposited diatom assemblages from general planktonic species to benthic species in these horizons confirm the redeposition of lateral sediments (Holger Cremer, pers. communication 1999).

6. Deglaciation and Holocene marine inundation on western Ymer Ø, East Greenland

Bernd Wagner & Martin Melles

(in prep.)

6.1. Abstract

Sediment records from two lakes on western Ymer Ø, East Greenland, were investigated for chronology, lithology, biogeochemistry and marine fossils. The data evidence that deglaciation of this interior coastal region has commenced prior to 10,000 cal. yr. B.P., suggesting a rapid ice retreat after the Milne Land stade. Deglaciation was followed by a marine inundation. The inundation led to marine sedimentation in Noa Sø, which at present is located 32 m a.s.l., and to accumulation of marine bivalve shells up to 60 m a.s.l. in the modern lake surroundings, but did not inundate a small lake (N1) at about 120 m a.s.l. to the north of Noa Sø. Radiocarbon dating of the marine sediments and bivalves indicate that sea level was still more than 30 m a.s.l. at about 6200 cal. yr. B.P. Together with published data this points to an uplift rate of ca. 1.3 cm/yr between about 8300 and 4700 cal. yr B.P., being lower and more constant than assumed by previous investigations in the region.

6.2. Introduction

During Last Glacial Maximum (LGM) the water stored in continental ice masses and the cooling of oceanic water led to a sea level about 120 m lower than today. Warming and ice melt following the LGM resulted in a global eustatic sea level rise that is well documented, for instance, in the growing rates of coral reefs of Barbados (Fairbanks 1989, Broecker 1998). From ca. 18 ka B.P. the sea level rose double s-shaped with maximum rates between ca. 14 ka B.P. and 6 ka B.P., interrupted by an intermediate weak decline between 11 and 10.5 ka B.P.

In East Greenland, as in other glaciated regions, the global eustatic sea-level rise after the LGM was superimposed by an isostatic rebound of the continent, due to a decrease in the extent and thickness of the Greenland ice sheet and outlet glaciers. Deglaciation started at ca. 16-14 ka B.P. (Hjort and Björck 1984, Marienfeld 1991, Funder and Hansen 1996, Nam 1997, Funder et al. 1998). The ice retreat was interrupted during the Milne Land stade at ca. 10.3 ka B.P., when glaciers covered the outer fjord region in East Greenland (Hjort 1979, Funder 1989). Renewed glacier recession to their present positions, being asynchronous in the different fjords and interrupted by standstills and small advances, commenced between 10.2 and 9.5 ka B.P. (Funder 1989).

The combination of eustatic and isostatic movements in wide coastal areas of Greenland resulted in a postglacial marine inundation. The sea-level changes are documented in the ages and altitudes of marine fossils and of marine geomorphological features along the coast (Hjort 1979, Funder 1989). Based on their mapping the uplift history of East Greenland was reconstructed, exhibiting different uplift rates in the outer and in the interior coastal regions (Funder 1989, Funder 1990, Björck et al. 1994a, Funder and Hansen 1996).

In this paper we present new data on the deglaciation and sea-level history on western Ymer Ø, derived from marine and lacustrine sediments in two Holocene lake sediment records, and from marine bivalve shells in the lake surroundings. A comparison of these data with published data, based on age calibration into calendar years B.P., supplies new information about the Holocene deglaciation and uplift of the interior region of East Greenland.

6.3. Study area

The study area on western Ymer Ø, East Greenland, is an anticline that forms the morphological extension of the Dusens Fjord in the east to the Kejser Franz Josephs Fjord in the west (Fig. 6.1).

Noa Sø, located in the center of the anticline, has an irregular oval shape with a length of ca. 3.8 km in W-E direction and a width of ca. 2.7 km in N-S direction (Fig. 6.2). The lake is situated at an altitude of 32 m a.s.l. Its main inflow, fed by an ice cap to the south, enters the lake at the southeastern shore, where it has formed a large delta. The outflow at the eastern lake shore via a gently inclined slope and a smaller lake finally drains into the Dusens Fjord. A bathymetric and shallow seismic survey has shown that Noa Sø has a complicated bathymetry, with several basins being separated by subaquatic ridges which mostly trend in WNW to ENE direction (Melles et al. 1995). The highest sediment fill of more than 10 m was observed in front of the delta and in the northern lake part, where the deepest depression with more than 120 m water depth occurs (Niessen and Melles 1995). At coring site PG1200, in western Noa Sø, the sediment thickness is ca. 2 m at a water depth of 12 m.

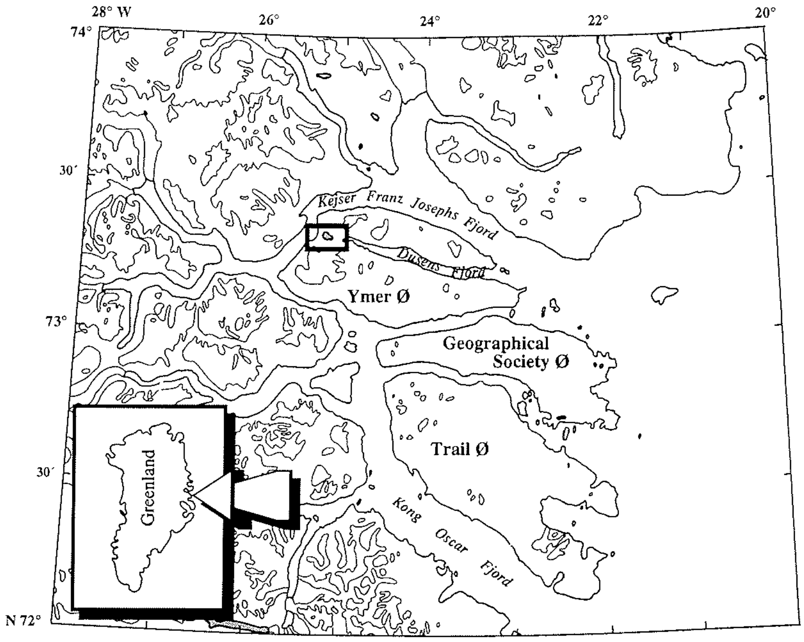


Fig. 6.1: Map of central East Greenland showing the study area (rectangle) on western Ymer Ø and the surrounding fjord systems and present locations of outlet glaciers and local ice caps.

Lake N1 is located about 1 km to the north of Noa Sø at an altitude of ca. 120 m a.s.l. (Fig. 6.2). The lake fills an elongated, ca. 1 km long and 0.5 km wide basin which trends from east to west. The main inflow at the eastern shore drains a valley that is separated by a ridge from Noa Sø. The outflow is located at the western lake shore, draining to the Kejser Franz Josephs Fjord. Core PG1204 was drilled in the central, deepest part of the lake, at 27 m water depth.

The geology of the study site is characterized by numerous faults, which mostly trend from west to east through the anticline (Harpøth et al. 1986). The slopes towards the south and north of Noa Sø exhibit similar geological structures. A succession of older Quarzite Series, via Multicolored Series and Limestone Dolomite Series, towards a younger Tillite Group is exposed with increasing distance from Noa Sø. All these Series belong to the Eleonore Bay Group and are of Precambrian age. In high altitudes, they are overlain by Cambrian bedrock which is partly covered by the ice caps to the north and south of the anticline. The larger southern ice cap today supplies high amounts of turbid melt water from the Multicolored Series to Noa Sø, causing its red-coloured water. Lake N1, in contrast, today is not affected by the smaller northern ice cap and thus has clear water.

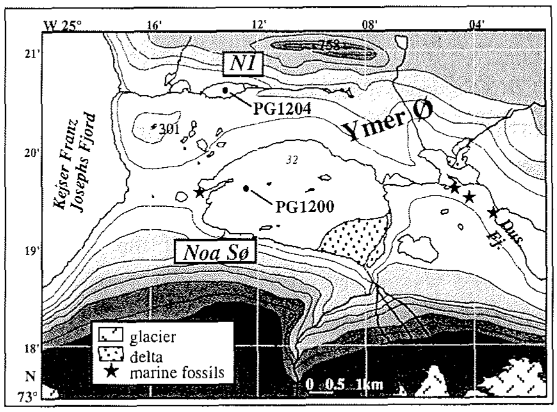


Fig. 6.2: Map of the study area showing Noa Sø and lake N1 embedded in the anticline between Kejser Franz Josephs Fjord to the west and Dusens Fjord to the east. Black dots mark the lake sediment coring locations, asterisks the bivalve shell sampling sites in the lake surroundings (contour lines are in 100 m).

6.4. Material and methods

Shells of marine bivalves (*Mya truncata*) were collected from four locations in the western and eastern Noa Sø surroundings (Fig. 6.2). These locations vary in altitude between about 10 and 60 m a.s.l. At all sites, the shells were embedded in poorly sorted, red-coloured sediments, which contained varying amounts of sand and gravels. The sediments predominantly occur at the modern surface of terraces. No paired shells in life position were found.

Sediment coring at sites PG1200 in Noa Sø and PG1204 in lake N1 was carried out from a floating platform by a piston corer. With this gear up to 3 m long sediment segments can be recovered. By coring of overlapping segments from defined depths, and their correlation on basis of core description and analytical data, a much longer continuous sequence can be obtained. A more detailed description of the coring technique is given by Melles et al. (1994).

The cores, stored in PVC liners of 6 cm diameter, were cut along their axes and split into two halves. Following core description and photographic documentation, one core half was continuously subsampled in 2 cm intervals. The subsamples were freeze-dried, and their water contents (in % of the wet bulk sediment) calculated from the mass differences between the wet and dry samples.

Biogeochemical and grain-size analyses were conducted in intervals of 4 cm on core PG1200 from Noa Sø, and in intervals of 8 cm on core PG1204 from lake N1. For the biogeochemical measurements, an aliquot of the sediment subsamples was ground (to < 63 μm) and homogenized in achate breakers by a planet mill. The contents of total carbon (TC), total nitrogen (TN) and total sulphur (TS) were measured with a CHNS-932 determinator (LECO Corp.). Total organic carbon (TOC) was determined with a Metalyt-CS-1000-S (ELTRA Corp.) in corresponding samples, which had been treated with HCl (10 %) at a temperature of 80°C to remove carbonate. The carbonate contents were calculated from the contents of carbonaceous carbon, derived from the difference between TC and TOC, and the atomic weights of the elements. The grain-size distribution of the samples was measured by a laser particle analyser GALAI CIS-1 (LOT Corp.) after adding NH_3 to all samples to diminish the surface tension, and after disaggregation of the sediment components in an ultrasonic sifter (15 sec). The maximum grain diameter to be measured with the CIS-1 is limited to 150 μm.

Sediment horizons bearing marine fossils were identified by microscopical inspection of the > 32 μm fraction that was isolated from large core intervals (0.1-1 m) by wet sieving. In horizons containing marine fossils, additional samples were taken in intervals of 2 cm, and the fossils isolated for their determination and radiocarbon dating. Terrestrial plant remains were isolated and mechanically cleaned from other sediment using tweezers and a needle under a binocular microscope.

Radiocarbon dating of the samples from the sediment cores was carried out by Accelerator Mass Spectrometry (AMS) at the Van de Graff Laboratory, University of Utrecht. In core PG1200, dating was conducted on bivalve shells (*Macoma* sp.) and their fragments from 175 and 165 cm sediment depths, on foraminifera from 150 to 146 cm depth, and on plant remains from the same horizon. In core PG1204, a pelvic spine of *Gasterosteus aculeatus* (fish, sticklebar) from 240 cm depth and terrestrial plant remains from 195 and 52 cm depths were dated. The ages of the bivalve shells (*Mya truncata*) collected in the field were determined by conventional radiocarbon dating at the Alfred Wegener Institute for Polar and Marine Research in Potsdam, after mechanical cleaning of the fragments.

The ^{14}C ages from the marine shells and foraminifera were corrected by 550 yrs for the marine reservoir effect in East Greenland (Tauber and Funder 1975). These reservoir corrected ages, and the ^{14}C ages determined on the terrestrial plant remains and the pelvic spine of the sticklebar, were calibrated into calendar years before present (cal. yr B.P.) using the calibration programme CALIB 4.0 (Stuiver and Reimer 1993, Stuiver et al. 1998). The means and uncertainties of the calibrated ages are calculated from the upper and lower boundaries of the probability distribution at the 2σ level. The same calibration was employed for published radiocarbon dates from Holocene marine fossils from East Greenland (based on their reservoir corrected ^{14}C ages) for a better comparison with our data.

6.5. Results and discussion

6.5.1. History of Noa Sø

The sediment core recovered at site PG1200 in Noa Sø has a length of 201 cm (Fig. 6.3A). A comparable thickness of the unconsolidated sediments visible in a sub-bottom profile crossing the coring site indicates that core PG1200 comprises the entire postglacial sediment record.

The basal core part (201-180 cm) is built up by a grey diamicton (Fig. 6.3A). Its normal consolidation and wide grain-size spectrum from mud to rounded gravel suggests subaquatic deposition in a proglacial environment. Low and stable current velocities are indicated by the lack of sorting and stratification. Significant contents of total organic carbon (TOC), total nitrogen (TN), total sulphur (TS), and carbonate in the diamicton indicate incorporation of reworked sediments and sedimentary rocks. These characteristics of the diamicton may be due to its formation by a slump, however, they could also be due to glacial deposition during deglaciation of the Noa Sø depression.

According to Hjort (1979), Björck et al. (1997), Funder et al. (1998); Wagner et al. (in press), Wagner and Melles (subm.) the last glaciation of the outer fjords of East Greenland occurred during the Milne Land stade, which corresponds to the Preboreal oscillation at 11,300-11,150 cal. yr B.P. (Björck et al. 1997). At that time, an outlet glacier flowing through the Kejser Franz Josephs Fjord likely had a branch eastwards into the Dusens Fjord (Fig. 6.1). This is indicated by the grey diamicton in Noa Sø, lacking the red sediments from the Multicolored Series which occur in the southern and northern lake surroundings.

The greyish sandy mud in 180 to 135 cm depth in core PG1200 differs from the underlaying diamicton by its smaller grain size, a higher water content, higher contents in TOC and TN, and decreasing contents of coarse sand, gravel, and carbonate with decreasing depth (Fig. 6.3A). This composition indicates a gradual decrease in glacial sediment supply simultaneously with an increase in biogenic accumulation. Relatively high contents of TS, and in particular the occurrence of marine bivalve shells (*Macoma* sp.) and benthic foraminifera (*Elphidium* sp.; Oberhänsli, pers com., 1999) indicate marine conditions in consequence of an inundation that affected the Noa Sø depression after its deglaciation. Comparable findings and interpretions were made in early Holocene lake sediments in East Greenland by Hjort (1979), Funder (1990) and Björck et al. (1994a, 1994b).

The age of the bivalve shell from 175 cm sediment depth, 26 cm above the core base, (10,270±440 cal. yr B.P.; Tab. 6.1) supports the suggestion that the marine

Tab. 6.1: Ages determined by radiocarbon AMS dating on micro and macrofossils isolated from core PG1200, Noa Sø, and core PG1204, lake N1. Calendar ages (cal. yr B.P.) were derived from calibration of the data, in marine fossils after correction of the marine reservoir effect. Uncertainties are reported at the 2σ level.

core	**depth** (cm)	**material**	**sample no.**	**weight** (mg)	$\delta^{13}C$ (‰)	^{14}C **age** (B.P.)	**cal. age** (cal. yr B.P.)
PG1200	146-150	plant remains	UtC-8053	0.07	-28.0	7980 ± 200	8860 ± 450
PG1200	146-150	foraminifera	UtC-8054	0.27	-1.4	8920 ± 110	9320 ± 330
PG1200	165	bivalve	UtC-7464	0.38	1.1	9630 ± 110	10200 ± 390
PG1200	175	bivalve	UtC-7465	0.20	-7.7	9730 ± 130	10270 ± 440
PG1204	52	plant remains	UtC-7417	1.25	-26.0	1880 ± 40	1800 ± 90
PG1204	195	plant remains	UtC-7459	0.10	-28.0	5420 ± 200	6220 ± 370
PG1204	240	pelvic spine	UtC-7460	0.41	-22.9	6880 ± 70	7710 ± 130

inundation of Noa Sø followed the Milne Land stade glaciation. A high sediment supply during the deglaciation is indicated by the similar age of the shells from 165 cm depth, 10 cm higher in the core (10,200±390 cal. yr B.P.). The difference between the ages of the terrestrial plant remains (8860±450 cal. yr B.P.) and the marine foraminifera (9320±330 cal. yr B.P.) from one horizon further upwards in the core (150-146 cm) may result from the relatively large sampling interval of 4 cm. However, the age difference could also be due to a higher reservoir effect than the assumed 550 yrs during the early Holocene. This could have been a consequence of the proximity to glacier ice at that time, which may have led to an input of ^{14}C-depleted melt water or a perennial sea-ice cover that prevented gas exchange with the atmosphere (Melles et al. 1997, Bondevik et al. 1999).

The sediments between 135 and 108 cm depth in core PG1200 are characterized by a succession from greyish sandy mud via reddish unlaminated mud to light greenish mud, and by the disappearance of the carbonate and IRD contents (Fig. 6.3A). The decreasing contents in the coarse fraction indicate proceeding distances from glacier fronts. The colour changes and decreasing carbonate contents implicate a gradual change in sediment sources. Whilst the supply of clastic sediments from the west by the outlet glacier diminished, supply of debris from the Multicoloured Series by the ice cap to the south of Noa Sø gained in importance.

A simultaneous decrease of the contents of TS and foraminifera indicates a gradual transition from marine via brackish to limnic conditions (Fig. 6.3A). In the upper part of the brackish sediments, slightly coarser grain sizes and, in particular, higher contents of TOC and TN occur. This is comparable with a sediment record from Peters Bugt Sø on Hochstetter Forland, NE Greenland, where a period of anoxic lagoon conditions or increased bioproduction reflects the end of the marine inundation (Björck

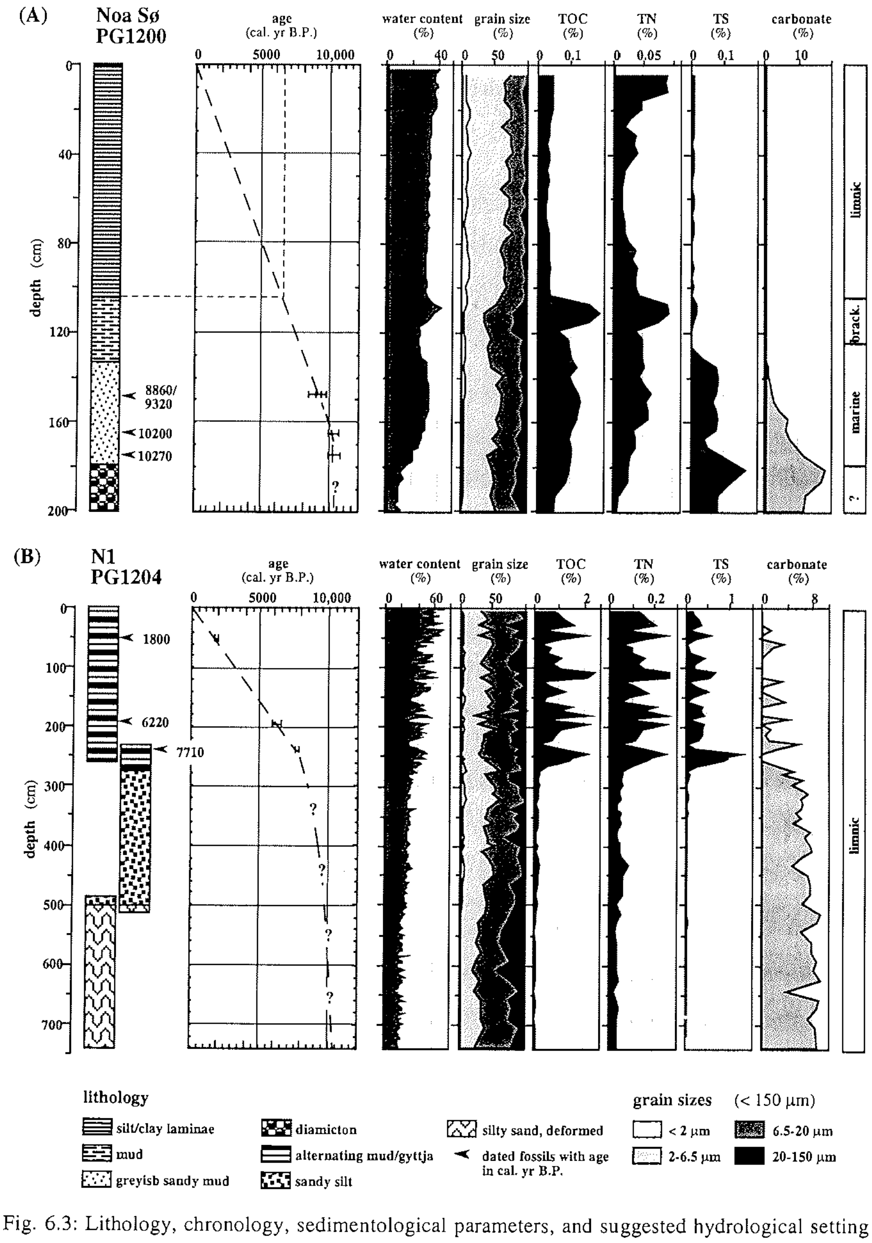


Fig. 6.3: Lithology, chronology, sedimentological parameters, and suggested hydrological setting of (A) core PG1200 from Noa Sø and (B) core PG1204 from lake N1. Black arrows mark the horizons of radiocarbon dated plant remains and marine fossils with their age in calendar years before present (cal. yr B.P.).

et al. 1994b). Assuming a rather constant sedimentation rate between the age of the glaciomarine sediments in Noa Sø and the surface sediments, the end of marine conditions is estimated to ca. 6600 cal. yr B.P. (Fig. 6.3A).

The uppermost 108 cm of core PG1200 are characterized by a light red colour and a fine alternation of silt and clay layers (Fig. 6.3A). This is typical for glacial varves formed in a lacustrine environment (Leemann and Niessen 1994a, Melles et al. 1995). Full limnic conditions are also reflected by low TS contents and the absence of marine fossils. The very low and stable contents of biogeochemical parameters in this facies result from the high suspension load, which often is observed in proglacial lakes, and which hampers bioproduction by light limitation. In addition, the increase of TN at the sediment surface, being not conform with the stable content of TOC, indicates that biogenic matter deposited in the sediment is partly oxidized.

6.5.2. History of lake N1

At site PG1204 in lake N1 a 7.4 m long sediment sequence was retrieved by three overlapping piston core segments (Fig. 6.3B). The lower part of the core (740-280 cm) is built up by stratified, strongly deformed silty sand and overlaying laminated sandy silt. Low contents of organic matter, a carbonate content similar to that in the oldest postglacial sediments of Noa Sø, and an obviously high sedimentation rate suggest that the sediments originate from the glacier retreat. In the upper part of this unit (500-280 cm) slightly decreasing contents in the coarse fraction (> 20 μm) and in carbonate, and increasing contents in clay and fine silt indicate a decreasing proximity to the glacial sediment source due to progressive glacier retreat in the catchment.

Above 280 cm depth in core PG1204 all sediment parameters show strong fluctuations (Fig. 6.3B). Ochre and black horizons with high contents in water, high values in the biogeochemical parameters, and lack of carbonate alternate with grey horizons of opposite composition. The ochre and black horizons represent times of significant biogenic accumulation. This is also documented in the occurrence of a pelvic spine from a sticklebar (*Gasterosteus aculeatus*) in 240 cm depth, dated to 7710±130 cal. yr B.P., and of terrestrial plant remains in 195 and 52 cm depths, dated to 6220±370 cal. yr B.P. and 1800±90 cal. yr B.P. (Tab. 6.1). During the formation of the intervening grey horizons, in contrast, the biogenic accumulation likely is diluted by enhanced terrigenous sediment supply. This is indicated by the lower contents in organic matter, but also by the occurrence of carbonate. As shown by $\delta^{13}C$ measurements (at depths of 60, 156, 188, 228, 276, 500 and 612 cm), the carbonates in the upper sediment sequence have the same range in carbon isotope ratios as those deposited in the basal sediments (3.88-4.29 ‰). This suggests that they originate from the same terrigenous source, likely the Limestone-Dolomite Series exposed to the northeast of lake N1. One reason for the fluctuating terrigenous sediment supply thus could be fluctuations in the size of the ice cap on the mountains to the northeast of the lake, leading to repeated ice advances into the catchment of lake N1. Today, the ice cap is located outsite the N1 catchment (Fig. 6.2), which corresponds with the lack of carbonate and relatively low concentration of biogenic matter in the surface sediments (Fig. 6.3B).

In contrast to Noa Sø, no evidence for marine or brackish conditions was found in the sediments from lake N1. Marine fossils do not occur, and the TS values during the assumed time of deglaciation are very low. The only hint for brackish conditions comes from the finding of the pelvic spine of a sticklebar (*Gasterosteus aculeatus*) at a depth of 240 cm (7710±130 cal. yr B.P.). The occurrence of sticklebars is often related to brackish water during the transition from marine to limnic conditions (Björck et al. 1994b, Bennike 1995). However, a marine to limnic transition in lake N1 is very unlikely, because sea level at 7700 cal. yr B.P. was already lowering in East Greenland (Funder 1990, Funder and Hansen 1996), and the older postglacial sediments show no indications for marine influences. Hence, more likely explanations are brackish conditions deriving from the spray of a close marine limit, or the occurrence of sticklebars in lake N1 was related to an ion and nutrient-rich limnic water body (Bennike 1997).

6.5.3. Holocene marine inundation

The upper limit of the marine inundation is indicated by the maximum altitudes of postglacial marine fossils, marine delta terraces, shorelines, fossil cliffs, or washed out boulders (Funder 1989). In East Greenland, highest marine features of Holocene age were found at about 130 m a.s.l., in the Kong Oscar Fjord region (Fig. 6.1) and in the Scoresby Sund region, 250 km further south (Funder 1978, Hjort 1979, Funder and Hansen 1996). On western Ymer Ø, the maximum uplift is documented by the altitudes of the marine bivalve shells (*Mya truncata*) and by the marine and lacustrine sediments in the lakes Noa and N1.

The highest altitude in the Noa Sø surroundings in which marine bivalve shells were found is ca. 60 m a.s.l. (Tab. 6.2). This altitude marks the minimum sea-level rise, because the bivalves do not reflect the sea surface but only a certain water depth during their lifetime (Funder and Hansen 1996). In addition, redeposition of the bivalves into deeper waters cannot be excluded, since none of the shells was found in life position. A postglacial sea-level rise of several 10 meters is also evidenced by the early to middle Holocene marine and brackish sedimentation in Noa Sø, which today is located 32 m a.s.l. However, the inundation probably did not reach lake N1 at an altitude of ca. 120 m a.s.l., indicated by the limnic sedimentation in this lake during the past at least 8000 cal. yr B.P. Hence, our data suggest a limit of the postglacial marine inundation between 60 and 120 m a.s.l. This confirms earlier work of Hjort (1979), who estimated an altitude

Tab. 6.2: Location and ages of shells collected on western Ymer Ø, East Greenland. Dating was conducted by the conventional radiocarbon method, calendar ages (cal. yr B.P.) were derived from a correction of the marine reservoir effect and subsequent calibration of the data. Uncertainties are reported at the 2σ level.

lat.	**long.**	**alt.** (m)	**sample** no.	**^{14}C age** (B.P.)	**cal. age** (cal. yr B.P.)
N 73°33	W 25°23	42	MM/G1	7480 ± 80	7800 ± 150
N 73°32	W 25°05	10	MM/G2	7190 ± 90	7530 ± 140
N 73°33	W 25°07	60	MM/G3	8020 ± 100	8340 ± 210
N 73°33	W 25°08	30	MM/G4	5930 ± 90	6180 ± 200

of ca. 70 m a.s.l. for this region, based on an extrapolation of the limit determined on sites to the north and south of Ymer Ø.

Not only the upper limit of the marine inundation differs along the coast of East Greenland but also its onset and duration. This is due to the general east-west direction of glacier retreat, and the regionally different retreat rates after the Milne Land stade, which are caused by different melting and calving rates (Funder 1989, Reeh 1989). The onshore locations of Holocene marine indicators in East Greenland were

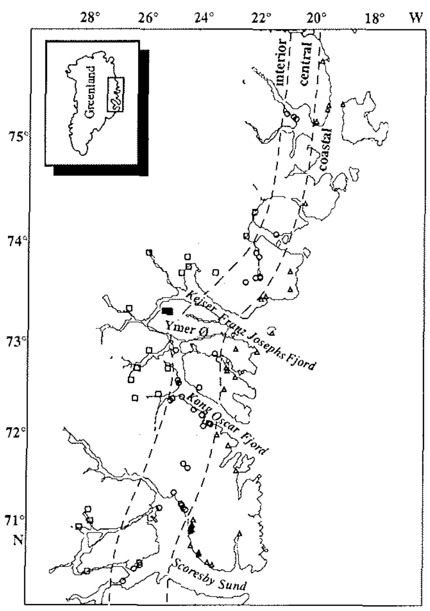


Fig. 6.4: Locations of marine fossils in East Greenland, with their separation in a coastal (triangles), a central (circles), and an interior region (open squares) according to Funder (1990). New locations from western Ymer Ø in the interior region are indicated by filled squares. Fossil altitudes and ages are presented in Tab. 6.3 and Fig. 6.5.

geographically grouped by Funder (1990) in a coastal, a central, and an interior region (Fig. 6.4). According to that, western Ymer Ø belongs to the interior region. A comparison of the new data from western Ymer Ø with the published data from marine fossils (Tab. 6.3, Fig. 6.5) supplies new information concerning the onset and duration of the marine inundation in the interior region of East Greenland.

Firstly, the ages of the oldest marine bivalves in the sediments in Noa Sø with 10,200 and 10,270 cal. yr B.P. (Fig. 6.3A, Tab. 6.1) pre-date the published data about the marine inundation into the interior region by about 700 yrs (Fig. 6.5). This may be due to a very rapid glacier retreat in the Ymer Ø region following the Milne Land stade, or to a location of the Milne Land stade ice margin much further inland than proposed in earlier studies (Hjort 1979, Funder 1989). Another possible explanation comes from the ridge separating the Kejser Franz Josephs Fjord and the Dusens Fjord (Fig. 6.1). The ridge may have diverted the entire ice flow through the Kejser Franz Josephs Fjord towards the north, and thus cut off the branch into the Dusens Fjord. This could have enabled the early deglaciation and marine inundation of the Noa Sø depression.

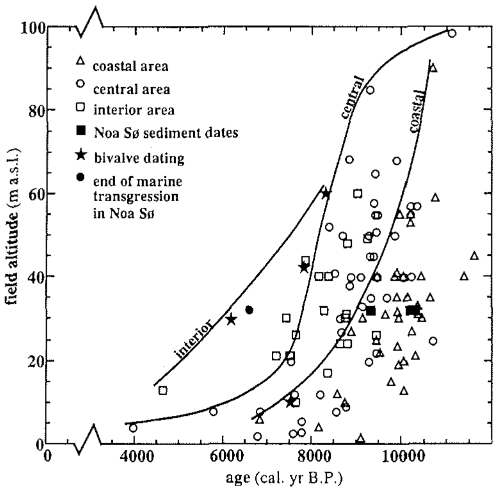


Fig. 6.5: Holocene uplift curves for East Greenland, deduced from the altitudes and ages of marine fossils from the coastal (triangles), central (circles), and interior (open squares) regions (Tab. 6.3, for locations see Fig. 6.4), modified after Funder (1990) on basis of the new data from western Ymer Ø (filled squares and asterisks).

In addition, the marine bivalve shells collected at about 30 m a.s.l. on western Ymer Ø with an age of 6180±200 cal. yr B.P. post-date the published data about this sea level in the interior region by ca. 1200 yrs (Fig. 6.5). A sea level close to 30 m a.s.l. at approximately that time is confirmed by the suggested transition from brackish to limnic conditions in Noa Sø (32 m a.s.l.) at ca. 6600 cal. yr B.P. (Fig. 6.3).

Tab. 6.3: Calendar ages (cal. yr B.P.) of marine shells in East Greenland, calculated from published radiocarbon ages and the new data from western Ymer Ø (Tab. 6.1 and 6.2). Sites are listed according to their occurrence in the coastal, central, and interior regions as defined by Funder (1990). Uncertainties are reported at the 2σ level. The shell locations are presented in Fig. 6.4, their altitudes and ages in Fig. 6.5.

lat.	long.	alt. (m)	sample no.	corr. ^{14}C age (B.P.)	cal. age (cal. yr B.P.)	reference
coastal						
N 70°50	W 23°45	40	K-3109	10010 ± 145	11400 ± 560	Funder 1990
N 70°53	W 23°63	1.5	K-3220	8140 ± 115	9120 ± 320	Funder 1990
N 70°63	W 23°93	27	K-3112	9100 ± 135	10210 ± 410	Funder 1990
N 70°65	W 23°93	23.5	K-2096	8860 ± 130	9920 ± 380	Funder 1978
N 70°68	W 24°05	55	AAR-965	8960 ± 150	9960 ± 360	Björck et al. 1994b
N 70°73	W 24°22	33	Lu-3400	9210 ± 110	10370 ± 400	Israelson et al. 1994
N 70°87	W 22°48	31.5	K-1919	8860 ± 140	9910 ± 390	Funder 1978
N 70°90	W 24°20	19	Lu-3282	8870 ± 110	9950 ± 350	Israelson et al. 1994
N 70°90	W 24°22	30	AAR-644	8180 ± 130	9140 ± 330	Israelson et al. 1994
N 70°90	W 24°22	21	AAR-643	9120 ± 180	10300 ± 500	Israelson et al. 1994
N 70°92	W 24°22	40		8760 ± 105	9870 ± 400	Funder and Hansen 1996
N 70°92	W 24°22	35	T-10382	8845 ± 110	9930 ± 360	Lyså and Landvik 1994
N 70°93	W 24°22	27	Lu-3283	8060 ± 100	8900 ± 210	Israelson et al. 1994
N 70°96	W 24°10	59		9590 ± 100	10760 ± 420	Funder and Hansen 1996
N 71°07	W 24°05	4		7310 ± 100	8160 ± 190	Funder and Hansen 1996
N 71°60	W 22°62	15	K-1462	8700 ± 120	9790 ± 380	Funder 1978
N 71°88	W 22°92	35	Lu-764	9330 ± 95	10640 ± 430	Håkansson 1974
N 72°00	W 23°33	40	Lu-485/486	8570 ± 90	9470 ± 370	Håkansson 1972
N 72°00	W 23°33	40	Lu-789	8960 ± 100	10050 ± 260	Håkansson 1974
N 72°50	W 23°08	90	Lu-712	9430 ± 95	10710 ± 410	Håkansson 1974
N 72°63	W 22°68	10	Lu-766	7870 ± 80	8760 ± 190	Håkansson 1974
N 72°70	W 22°97	25	Lu-714	8520 ± 90	9450 ± 360	Håkansson 1974
N 72°70	W 22°97	12	Lu-829	7700 ± 80	8590 ± 200	Håkansson 1974
N 72°73	W 23°00	55	Lu-645	9110 ± 95	10200 ± 380	Håkansson 1973
N 72°90	W 21°90	53	Lu-533	9060 ± 95	10200 ± 380	Håkansson 1973
N 72°93	W 22°67	40	Lu-710	9270 ± 95	10440 ± 360	Håkansson 1974
N 73°10	W 21°38	31	K-2376	9210 ± 110	10370 ± 400	Weidick 1976
N 73°45	W 21°78	35	I-9102	9055 ± 250	10150 ± 660	Weidick 1977
N 73°48	W 21°62	6	Lu-884	5970 ± 70	6840 ± 170	Håkansson 1975
N 73°55	W 20°75	45	Lu-882	10170 ± 150	11600 ± 740	Håkansson 1975
N 73°73	W 20°75	20	Lu-885	8990 ± 90	10060 ± 260	Håkansson 1975
N 74°40	W 20°25	31	I-9133	8685 ± 130	9630 ± 530	Weidick 1977
N 75°17	W 19°85	30	Lu-1384	9260 ± 95	10440 ± 370	Håkansson 1978
N 75°28	W 19°47	13	Lu-1300	8970 ± 90	10050 ± 260	Håkansson 1978
N 75°32	W 19°42	30	Lu-1386	8850 ± 90	9940 ± 340	Håkansson 1978
N 75°33	W 18°92	41	Lu-1389	8820 ± 90	9930 ± 350	Håkansson 1978
N 75°72	W 19°67	22	Lu-1289	8640 ± 90	9530 ± 430	Håkansson 1978
central						
N 70°30	W 26°58	5.5	K-1460	6890 ± 130	7770 ± 230	Funder 1978
N 70°45	W 26°20	2.5	I-5423	6690 ± 125	7590 ± 220	Funder 1978
N 70°50	W 26°00	45	I-9492	8375 ± 140	9380 ± 400	Funder 1978
N 70°53	W 26°03	68.5	I-8890	8010 ± 130	8830 ± 270	Funder 1978
N 71°15	W 24°42	27	I-9491	7790 ± 135	8670 ± 270	Funder 1978
N 71°17	W 25°33	65	K-1461	8490 ± 140	9430 ± 400	Funder 1978
N 71°17	W 24°50	50	I-9490	8750 ± 135	9850 ± 430	Funder 1978

Tab. 6.3 (continued)

N 71°20	W 24°53	55	HAR-2957	8470 ± 150	9420 ± 400	Funder 1990
N 71°22	W 24°58	40	HAR-2956	8120 ± 100	9110 ± 310	Funder 1990
N 71°35	W 24°83	98.5	K-1915	9750 ± 120	11110 ± 510	Funder 1978
N 71°63	W 24°35	51	K-2385	8520 ± 130	9450 ± 390	Street 1977
N 71°63	W 24°35	58	K-2386	8420 ± 130	9400 ± 390	Street 1977
N 71°68	W 24°50	50	W-1381	7750 ± 350	8680 ± 770	Levin et al. 1965
N 72°00	W 23°33	40	Lu-485	8570 ± 90	9470 ± 370	Håkansson 1972
N 72°10	W 23°83	40	Lu-649	8110 ± 85	9110 ± 300	Håkansson 1973
N 72°12	W 23°57	57	Lu-647	9060 ± 95	10200 ± 380	Håkansson 1973
N 72°13	W 23°63	55	Lu-528	9040 ± 75	10190 ± 370	Håkansson 1972
N 72°13	W 23°58	35	Lu-531	8390 ± 85	9330 ± 320	Håkansson 1972
N 72°22	W 23°88	4	Lu-831	3580 ± 60	3990 ± 160	Håkansson 1974
N 72°28	W 24°17	12	Lu-488	7340 ± 80	8190 ± 160	Håkansson 1972
N 72°38	W 25°00	41	Lu-530	7620 ± 80	8520 ± 200	Håkansson 1972
N 72°40	W 24°92	38	Lu-792	8000 ± 85	8860 ± 200	Håkansson 1974
N 72°42	W 24°58	55	Lu-790	8580 ± 90	9470 ± 370	Håkansson 1974
N 72°52	W 23°97	33	Lu-490	8160 ± 85	9140 ± 290	Håkansson 1972
N 72°52	W 23°97	68	Lu-529	8770 ± 90	9880 ± 390	Håkansson 1972
N 72°52	W 23°97	45	Lu-826	8360 ± 85	9310 ± 320	Håkansson 1974
N 72°52	W 23°97	20	Lu-827	6640 ± 75	7530 ± 120	Håkansson 1974
N 72°52	W 23°97	8	Lu-828	5020 ± 65	5780 ± 130	Håkansson 1974
N 72°57	W 24°70	50	Lu-648	8310 ± 90	9290 ± 330	Håkansson 1973
N 72°60	W 24°73	85	Lu-711	8370 ± 85	9320 ± 320	Håkansson 1974
N 72°82	W 23°28	22	Lu-768	8480 ± 85	9430 ± 360	Håkansson 1974
N 72°82	W 23°28	30	Lu-830	7720 ± 80	8630 ± 220	Håkansson 1974
N 72°88	W 23°42	57	Lu-646	9190 ± 90	10330 ± 370	Håkansson 1973
N 72°88	W 23°42	20	Lu-793	8280 ± 85	9280 ± 340	Håkansson 1974
N 72°92	W 24°80	52	Lu-608	7540 ± 80	8400 ± 180	Håkansson 1973
N 73°62	W 22°35	2	Lu-868	5910 ± 70	6780 ± 160	Håkansson 1975
N 73°67	W 21°83	8	Lu-867	5950 ± 75	6820 ± 170	Håkansson 1975
N 73°67	W 22°00	35	Lu-883	8670 ± 90	9710 ± 300	Håkansson 1975
N 73°68	W 21°83	40	Lu-886	8940 ± 90	10050 ± 260	Håkansson 1975
N 73°88	W 21°87	9	K-2573	7920 ± 120	8770 ± 250	Weidick 1977
N 73°92	W 21°97	12	K-2571	6740 ± 115	7620 ± 200	Weidick 1977
N 73°92	W 21°97	8	K-2572	7620 ± 120	8550 ± 290	Weidick 1977
N 74°10	W 21°27	40	I-9139	9105 ± 140	10230 ± 420	Weidick 1977
N 75°20	W 20°58	3	Lu-1292	6900 ± 75	7780 ± 140	Håkansson 1978
N 75°22	W 20°67	40	Lu-1390	8020 ± 85	8880 ± 200	Håkansson 1978
N 75°25	W 20°92	25	Lu-1303	9380 ± 90	10700 ± 430	Håkansson 1978
interior						
N 70°42	W 27°82	21	I-5422	6300 ± 125	7190 ± 260	Funder 1978
N 70°95	W 28°15	44	I-5421	6990 ± 130	7860 ± 250	Funder 1978
N 71°02	W 27°75	21	K-1459	6650 ± 130	7540 ± 230	Funder 1978
N 71°15	W 27°83	30	I-5420	6500 ± 125	7410 ± 220	Funder 1978
N 72°40	W 26°23	21	Lu-866	6590 ± 75	7490 ± 120	Håkansson 1975
N 72°45	W 25°42	40	Lu-492	7520 ± 80	8380 ± 180	Håkansson 1972
N 72°60	W 26°38	26	Lu-713	6760 ± 85	7640 ± 150	Håkansson 1974
N 72°73	W 26°17	40	Lu-767	7310 ± 80	8170 ± 170	Håkansson 1974
N 72°73	W 25°08	32	Lu-791	7420 ± 80	8270 ± 160	Håkansson 1974
N 72°73	W 25°08	24	Lu-825	7720 ± 80	8630 ± 220	Håkansson 1974
N 72°92	W 25°75	10	Lu-765	6770 ± 75	7640 ± 130	Håkansson 1974
N 73°35	W 26°47	17	I-9104	7480 ± 115	8350 ± 250	Weidick 1977

Tab. 6.3 (continued)

N 73°72	W 23°42	48	K-2417	7960 ± 120	8800 ± 260	Weidick 1976
N 73°72	W 24°62	49	K-2379	8250 ± 130	9260 ± 370	Weidick 1976
N 73°72	W 24°62	26	K-2418	8540 ± 130	9450 ± 390	Weidick 1976
N 73°78	W 24°38	31	K-2380	7920 ± 100	8790 ± 220	Weidick 1976
N 73°88	W 24°42	60	K-2381	8080 ± 130	9030 ± 390	Weidick 1976
N 73°92	W 25°78	13	K-2419	4090 ± 100	4650 ± 240	Weidick 1976
N 74°08	W 22°32	30	I-9131	7925 ± 130	8760 ± 280	Weidick 1977
N 74°32	W 22°05	24	I-9132	7970 ± 130	8800 ± 280	Weidick 1977
N 73°33	W 25°23	42	MM/G1	6930 ± 80	7800 ± 150	this study
N 73°32	W 25°05	10	MM/G2	6640 ± 90	7530 ± 140	this study
N 73°33	W 25°07	60	MM/G3	7470 ± 100	8340 ± 210	this study
N 73°33	W 25°08	30	MM/G4	5380 ± 90	6180 ± 200	this study
N 73°33	W 25°20	32	UtC-7464	9080 ± 110	10200 ± 390	this study
N 73°33	W 25°20	32	UtC-7465	9180 ± 130	10270 ± 440	this study
N 73°33	W 25°20	32	UtC-8054	8370 ± 110	9320 ± 330	this study

Under consideration of the new data from western Ymer Ø, an average uplift rate of ca. 1.3 cm/yr can be calculated for the interior region of East Greenland for the period 8340 (sample MM/G4; Tab. 6.2) to 4650 cal. yr B.P. (sample K-2419; Tab. 6.3). Compared to the central region, which is characterized by an earlier onset of uplift, by its s-shaped curve, and by its higher rate (ca. 3.7 cm/^{14}C-yr) between ca. 11,000 and 7500 cal. yr B.P. (Funder 1990, Funder and Hansen 1996), the data from Ymer Ø suggest a delayed, slower and more constant uplift. This is probably due to the east-west direction of deglaciation, and a lower isostatic rebound of areas with larger distance to the Milne Land stade margins, where maximum uplift was attained as a response to the ice decay (Funder and Hansen 1996). In the interior regions, the land is still down-pressed by the inland ice that survived the change from the last glacial to the present interglacial with a loss of only 50 % of its volume (Funder et al. 1998).

6.6. Conclusions

Based on lithological, biogeochemical and chronological investigations of two lake sediment records from western Ymer Ø, East Greenland, and on radiocarbon dating of marine bivalve shells occurring in their surroundings, the following conclusions can be drawn concerning the glacial history and the postglacial Holocene marine inundation in this region.

Deglaciation of the western Ymer Ø after the Milne Land stade commenced prior to 10,300 cal. yr B.P., about 700 yrs earlier than indicated by published radiocarbon dates from marine fossils in the interior East Greenland. This indicates that the glacial ice margin during the Milne Land stade was either further inland as expected so far or that the glacier retreat was very rapid in this region.

The limit of the marine inundation on western Ymer Ø is located between 60 m a.s.l., the maximum altitude of bivalve shells found in the region, and 120 m a.s.l., the approximate altitude of lake N1 that exhibits limnic sedimentation during at least the past 8000 years. These results confirm earlier studies of Hjort (1979), which indicate a marine limit of ca. 70 m a.s.l. in the investigated area, based on the extrapolation of marine features to the north and south of Ymer Ø. The middle Holocene uplift rate on western Ymer Ø was lower (ca. 1.3 cm/yr) and more constant than assumed by previous investigations in the interior region of East Greenland, evidenced by early to middle Holocene marine and brackish sedimentation in Noa Sø, and the coincident occurrence of bivalve shells in its surrounding.

7. Synthesis

7.1. The need of a multi-disciplinary approach for palaeoenvironmental reconstructions

To reconstruct a reliable environmental history of East Greenland based on lake sediments, a multi-disciplinary approach offers several advantages.

Shallow-seismic measurements in Basaltsø and Noa Sø allowed a first view into the sediment columns prior to coring. They depicted a succession of acoustic high transparent and low transparent layers in the seismograms, indicative for changes in sediment consolidation, grain-size distribution, or the presence of gas (Fig. 4.3). In addition, the geometry of the reflectors gave a hint on disturbed or undisturbed bedding of the sediments. Based on this information the coring positions were selected. By coring the entire sediment columns at locations, where a hard bottom reflector indicated the occurrence of a till or bedrock at the sediment base, the complete lake history was recovered.

Measurements of the physical properties prior to core opening enabled an estimation about the sediment consolidation or its amount of water-rich organic matter. Matching values of the physical properties between single core segments reflect their overlapping sequences and thus allowed the determination of the maximum sediment depth recovered with the piston corer (Figs. 4.4 and 6.3B).

The measurement of the grain-size distribution gives valuable information about the sediment origins, mass movement processes from lateral deposited sediments, the distance to an inflow, and its transport energy. The glacial and glaciolimnic sediments in the here investigated lakes are dominated by coarse-grained sediments and indicate the presence of glacier ice on the lake depressions or in their catchments (Figs. 4.4, 4.5, 5.3, and 6.3). A change in the grain-size distribution from coarser to finer sediments in the Noa Sø indicates a transition from a marine to a limnic environment (Fig. 6.3). The subsequent limnic sedimentation is dominated by a constant alternation of silt and clay layers, originating from the glacial meltwater inflow from the south, and without evidences for distinct changes in the meltwater supply. In contrast to Noa Sø, limnic sediments commonly are enriched in organic matter, which may blur the information derived from the grain-size distribution. This is suggested in lake B1 and in Raffles Sø, where high amounts of organic matter were revealed after core opening (Figs. 4.5 and 5.3). In contrast, Basaltsø and lake N1 have lower contents of organic matter in the sediments. Therefore, distinct changes in the grain-size distributions of their limnic sediments indicate mainly mass movement processes and fluctuations in the transport energy of the lake inflows (Figs. 4.4 and 6.3).

Different concentration profiles of total organic carbon, total nitrogen and total sulphur in the sediment, all predominantly originating from deposited organic matter, reveal diagenetic processes, decomposition and redox-limited movement processes. This is especially evident from the contents of total sulphur, which shows most distinct peaks at lithofacies changes, independently from the other biogeochemical parameters. For instance, a distinct maximum of total sulphur occurred at the transition from glaciolimnic to limnic sediments in the cores PG1205 from Basaltsø and PG1204 from lake N1 (Figs. 4.4 and 6.3B). A sulphur maximum in core PG1200 from Noa Sø was measured at the transition from glacial to glaciomarine sediments (Fig. 6.3A). In core PG1214 from Raffles Sø a sulphur peak indicates the transition from a biogenic gyttja to a mixed biogenic and clastic gyttja during the late Holocene (Fig. 5.6).

The concentration profiles of total organic carbon and total nitrogen are similar in the individual cores (Figs. 4.4, 4.5, 5.6, and 6.3). This indicates that both parameters are affected by a similar amount of postsedimentary decomposition. The decomposition of organic matter depends on the oxydizing conditions in the bottom waters, and on the availability of organic matter, which is easy to decompose and whose components are needed for new bioproduction within the lake. This is autochthonous formed organic matter, being rich in nitrogen and due to its depletion of lignin and cellulose easy decomposable (Meyers and Ishiwatari 1995, Clark and Fritz 1997). Loss of total organic carbon and total nitrogen was revealed in the early Holocene sediments of Basaltsø, where probably a high amount of autochthonous matter by decomposition was returned to the lake internal nutrient cycle. Even though a release of total organic carbon and total nitrogen from the sediment also plays a role in the other sediment sequences, both are necessary parameters to reflect the environmental histories of the lakes.

The C/N ratio indicates differences between the relative amounts of allochthonous and autochthonous organic matter deposited in the sediment. By decomposition of the preferred autochthonous matter changes in the C/N ratio may be blurred. Therefore, the C/N ratio has to be interpreted very carefully, on the one hand in the light of different supply of autochthonous and allochthonous organic matter, and on the other hand in the light of a change in decomposing conditions.

Biogenic opal in limnic sytems predominantly derives from diatoms and their fragments. Diatoms settle due to their high reproduction rate, their easy passive dispersal, and their wide ecological range in most circumneutral lakes. There, the opal concentration in the sediment may be a good indicator for environmental changes, such as in the sediments of Basaltsø (Figs. 4.4 and 4.10) or in those of Raffles Sø (Figs 5.3 and 5.7). Low opal concentrations in sediments may derive from reduced growing rates of diatoms due to light limitation in the water column. For instance, the high glacial meltwater supply in Noa Sø causes a very cloudy water and thus prevents diatom growing. A further reason for light limitation may be a perennial ice cover, as indicated in Raffles Sø sediment during cold periods (Fig. 5.3). Another explanation for low opal concentrations in sediments are alkaline conditions in lakes and sediments. They may prevent a settlement of diatoms or lead to postsedimentary dissolution of their silicatic skeleton. The former is suggested in the sediment of lake N1 by the occasional occurrence of terrigenous carbonate (Fig. 6.3B), probably leading to a widely absence of diatoms. The latter is assumed in lake B1, where periods of lower opal concentrations correspond to periods of higher autigenous carbonate concentrations (Fig. 4.5).

The autigenous carbonates in the sediment of lake B1 were identified as siderites by isotopic measurements on the carbonaceous fraction and by determination of the cations (Fig. 5.5). The formation of siderites strongly depends on the presence of iron, and on reducing conditions. Hence, the occurrence of siderite in the sediment may serve as an indicator for a period of high bioproduction, when the decomposition of the organic matter causes reducing conditions. This is also valid for the occurrence of vivianite concretions in the same sediment sequence, which were formed under anaerob conditions, when large amounts of iron and organic phosphorus were available.

Palynological investigations of the Basaltsø and Raffles Sø sediments indicated different vegetation assemblages along the outer coast of East Greenland. Whilst the palynological record from the Basaltsø region showed distinct changes during the Holocene (Fig. 4.8), that from Raffles Sø inferred a rather uniform vegetation in the surrounding (Fig. 5.5). Both Holocene records were influenced by long distance transport of pollen grains, slightly more pronounced in the Raffles Sø probably due to a sparser vegetation cover its surrounding. As shown by a delayed onset of both records, the vegetation assemblage not only depends on the climate, but also on the immigration of plants and the state of soil formation. A delayed immigration of terrestrial plants also influences the accumulation of organic matter in the lake sediments. These restrictions for the use of pollen grains for palaeoenvironmetal reconstructions may also be valid for the occurrence of other micro and macrofossils in the sediments.

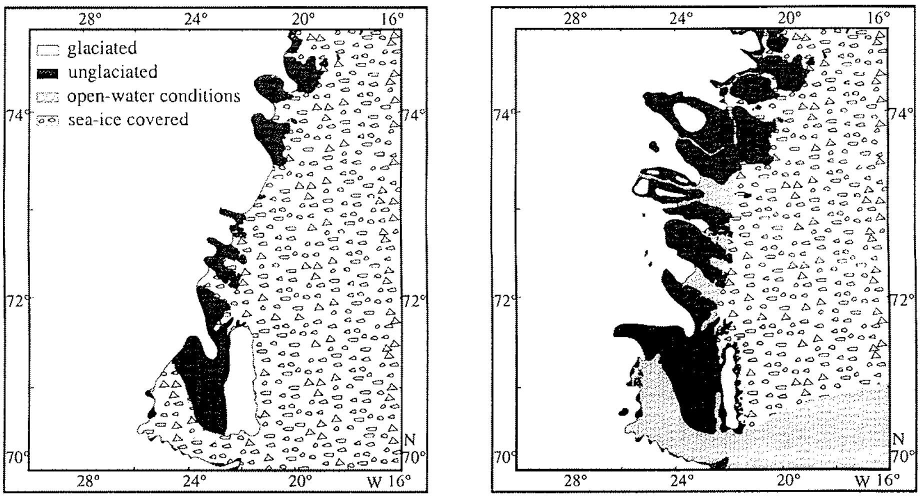
In summary, a multi-disciplinary approach for the study of lake sediments to reconstruct the palaeoenvironmental history reveals distinct differences in the applicability of individual parameters. Most individual parameters are influenced by local effects or by diagenic processes. Therefore, it is absolutely necessary to combine the information of individual parameters from one site for its palaeoenvironmental reconstruction. The subsequent combination of information from different sites enables the palaeoenvironmental reconstruction of a larger area.

7.2. Reconstruction of the Holocene environmental history of East Greenland

All sediment sequences studied have glacial, glaciolimnic or glaciomarine sediments at their base. That implies an ice covering of all coring locations during the last glaciation of the coastal East Greenland, which occurred during the Milne Land stade dated to 11,300-11,150 cal. yr B.P. (Hjort 1979, Björck et al. 1997, Funder et al. 1998). As evidenced by geomorphological features, for instance in the Scoresby Sund area by large moraine structures, the fjords were filled by glacier tongues at this time (Funder and Hansen 1996). The mountain plateaus in the interior regions were also covered by ice. Nevertheless, a small belt of coastal upland was ice free, and also numerous nunataks in the more interior region have existed, probably relicts of the dry conditions during the Last Glacial Maximum (Funder 1979; Fig. 7.1A). Laminated sediments in the Scoresby Sund indicate a perennial ice cover of the fjords during this period (Marienfeld 1991, Dowdeswell et al. 1994), supported by a close sea-ice cover off East Greenland (Koç et al. 1993).

Increasing biogenic accumulation at about 10,000 cal. yr B.P. in the sediments of the lakes Basaltsø, B1 and Raffles Sø reflects the glacial recession from the outer coastal areas (Figs. 4.4-4.7, 5.3 and 5.6). The high rate of glacial recession at this time is indicated by prevailing marine conditions at the Noa Sø depression in the interior region of East Greenland, even though the glacier retreat from this area is probably supported by the surrounding morphology. Due to the marine transgression, which followed the deglaciation and inundated the flat coastal areas, Ymer Ø was separated in a northern and a southern part (Fig. 7.1B). Warm and dry summers due to a maximum in the Northern Hemisphere insolation during this period (Berger 1978) created suitable conditions for the immigration of terrestrial plants and for increasing bioproductivity in the lakes. These conditions also led to open-water conditions in the fjords during summer (Marienfeld 1991), indicated in high amounts of ice-rafted debris in the sediments due to extensive calving of the receeding glaciers. Changes in the North Atlantic circulation pattern supported the open-water conditions, in particular in the southern study area. A strong inflow of warm Atlantic water masses in the Greenland, Iceland and Norwegian Seas led to a northward migration of the summer sea-ice margin (Koç et al. 1993).

The period between 10,000 and 7500 cal. yr B.P. is characterized by highest contents of organic matter in the sediments of the lakes Basaltsø and B1 (Fig. 4.9). This indicates warm and dry summers, as also reflected by the establishment of a birch dominated dwarf shrub heath in the surrounding of both lakes. The prevailing warm conditions led to a rapid glacier recession to approximately their present positions in the inner fjords (Hjort 1979, Funder 1989, Funder and Hansen 1996; Fig. 7.1C). At Noa Sø,

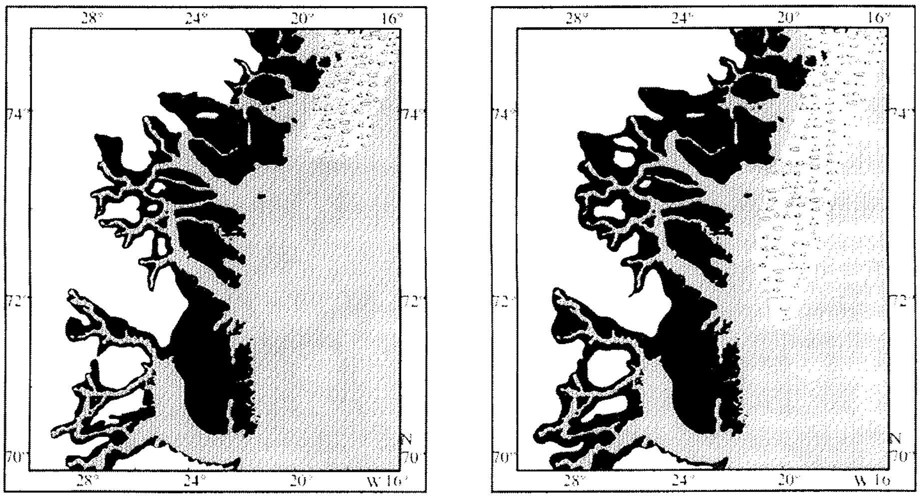


(A) 11,000 cal. yr B.P. **(B) 10,000 cal. yr B.P.**

Fig. 7.1: Reconstruction of major glacier fronts in East Greenland, the marine limits and the summer sea-ice cover from the end of the Milne Land stade until today (A-H), based on new data from the investigated lakes and published data mentioned in the text. Note that the boundaries between the glaciated areas, unglaciated areas, areas with open-water conditions and those with sea-ice cover are generalized.

the sediment supply from an outlet glacier through the Kejser Franz Josephs Fjord was gradually replaced by a supply from the local southern ice cap. Simultaneously, the marine conditions on the Noa Sø depression changed into brackish conditions due to the decreasing influence of marine conditions (Fig. 6.3). That was the consequence of the slow and constant isostatic rebound in the interior region after the ice decay, leading to an exposure of the anticline between the Dusens Fjord to the east and the Kejser Franz Josephs Fjord to the west above the marine limit. Offshore, the inflow of warm Atlantic water masses into the Greenland, Iceland and Norwegian Seas caused a weakening of the ice-transporting East Greenland Current (Koç et al. 1993), and thus led to a summer sea-ice margin far to the north of Liverpool Land. In consequence, the remaining unattractive feeding conditions for seabirds on Raffles Ø are indicated in low contents of organic matter in the sediments of Raffles Sø (Figs. 5.6 and 5.7).

Remaining favourable climatic conditions between 7500 and 6500 cal. yr B.P. are well reflected in the sediments of the lakes Basaltsø and B1, and in the vegetation assemblage in their surrounding (Fig. 4.9). The climatic conditions probably caused a retreat of glaciers behind their present positions (Hjort 1979, Funder 1989; Fig. 7.1D). In the Noa Sø depression, the transition from brackish to limnic conditions mirrors the isolation of the lake due to the constant uplift of the interior region (Figs. 6.3A and 6.5).



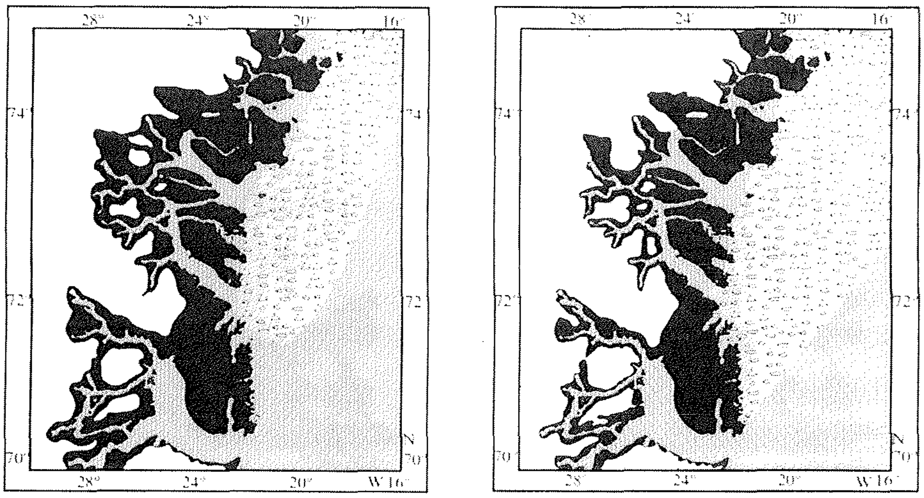
(C) **7500 cal. yr B.P.** (D) **6500 cal. yr B.P.**

Fig. 7.1 continued.

The strong increase of organic matter in Raffles Sø sediment indicates a progressive intensification of the East Greenland Current (Figs. 5.6 and 5.7). This probably reflects a settlement of seabird colonies on Raffles Ø during the summer, requiring at least two month mixture zones of open-water and pack-ice at sea for a breeding success.

The climatic deterioration at the outer coast between 6500 and 5000 cal. yr B.P. increased the transport energy of the brooks draining into Basaltsø (Fig. 4.10). Additional evidence for a rise in precipitation comes from the surrounding vegetation. The warm and dry conditions indicating birches were progressively replaced by moister and cooler conditions indicating willows (Fig. 4.8). However, the climatic deterioration at this time was more induced by an increase in snow accumulation than by a temperature decline, since the occurrence of the two bivalves *Mytilus edulis* and *Chlamys islandica* along the coast of East Greenland indicates that temperatures were still higher than present (Hjort and Funder 1974). However, diatom investigations offshore indicated that the ice-transporting East Greenland Current during this period intensified to a small belt along the coast, which led to a summer sea-ice margin as far south as Kong Oscar Fjord (Koç et al. 1993; Fig. 7.1E). This is confirmed by a further increase of the organic matter content in the sediment of Raffles Sø, inferring increasing settlement of seabirds on Raffles Ø (Fig. 5.6).

The period between 5000 and 2500 cal. yr B.P. was characterized by a cooling and remaining of moist conditions, recorded in a deceasing organic matter content in the sediments of the lakes B1 and Basaltsø, and in coarse sediments of the latter (Fig. 4.10).



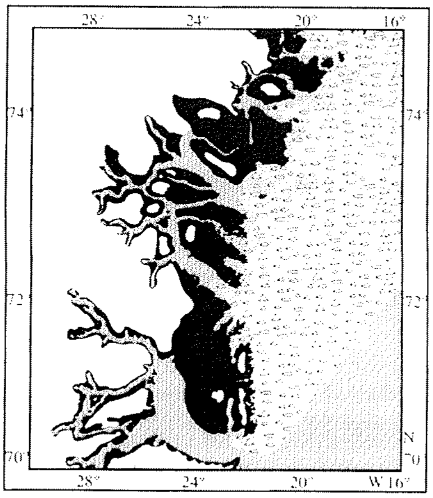
(E) **5000 cal. yr B.P.** (F) **2500 cal. yr B.P.**

Fig. 7.1 continued.

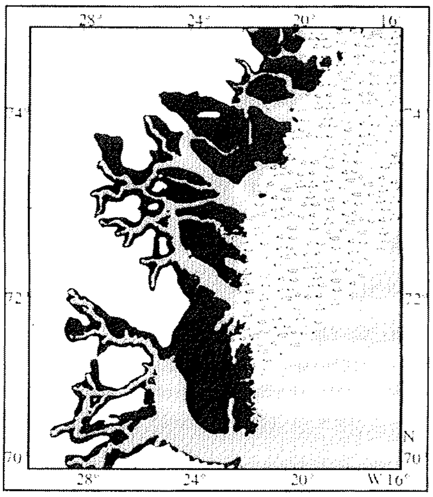
The birches in the lake surrounding almost completely disappeared, suggesting that the summer temperatures at the outer coast were too low for their growing. The cooling and the moist conditions may have led in northeastern Greenland, and, with a small delay, also in eastern Greenland to a renewed glacier advance (Funder 1989, Hjort 1997; Fig. 7.1F). The climatic deterioration was probably the result of decreasing northern Hemisphere summer insolation and an intensification of the East Greenland Current (Berger 1978, Koç et al. 1993). However, a remaining high organic matter content in the sediment of Raffles Sø suggests two months of snow-free breeding sites and attractive feeding conditions for seabirds on Raffles Ø during the summer (Fig. 5.6).

Rather cool and dry conditions from 2500 to 1500 cal. yr B.P. are indicated by a change from coarser towards finer sediments in Basaltsø and by low contents of organic matter in both sediment records from the Geographical Society Ø (Fig. 4.10). A distinct change of the vegetation assemblage from the willow dominated dwarf shrub heath to a more grass and herb dominated fell field vegetation reflects cool climatic conditions similar to the earliest Holocene period (Fig. 4.8). An abrupt decrease of the organic matter content in Raffles Sø indicates a strengthening of the East Greenland Current or a distinct cooling, the latter confirmed by a perennial lake-ice cover. Both led to a lower attractivity of Raffles Ø for breeding seabirds (Figs. 5.7 and 7.1G).

The Holocene period from 1500 cal. yr B.P. to the present is characterized by two distinct climatic changes, mirrored in the lake sediments on Geographical Society Ø and on Raffles Ø (Fig. 5.7). The first change may correspond to the medieval warm



(G) **1500 cal. yr B.P.**



(H) **recent**

Fig. 7.1 continued.

period and lasted from ca. 1000 to 500 cal. yr B.P. Slightly increased contents of organic matter in the sediments during this period and a short-term readvance of a willow dominated dwarf shrub heath indicate warmer conditions on Geographical Society Ø (Fig. 4.10). They are confirmed by the record from Raffles Sø, where renewed increase of nutrients in the sediment may reflect a higher acceptance of Raffles Ø as breeding site. Possible explanations are either a prolonging of the open-water conditions to more than two months during summer, or a better accessibility to the breeding sites on the island due to a less thick snow cover. The second half in the last millenium is dominated by the cool conditions of the Little Ice Age, reflected in lowest contents of organic matter in all studied cores from the outer coast since the early Holocene. This period finished, when recent warming is indicated in increasing organic matter in the sediments at ca. 100 cal. yr B.P. Today, the summer sea-ice margin is located in front of northern Liverpool Land, creating suitable conditions for breeding colonies on Raffles Ø (Fig. 7.1H).

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