The late Quaternary environmental and climatic history of North-East Greenland, inferred from coastal lakes

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“We only survived the physical strain
because we had enough chocolate with us.”

Roald Engebreth Gravning Amundsen
(1872-1928)

Preface

The research presented in this thesis was carried out within the framework of an international project that is targeted on a detailed reconstruction of the Late Quaternary glacial history and postglacial climatic and environmental changes in East and North-East Greenland. Collaborating institutes are the Geological Survey of Denmark and Greenland (GEUS), Copenhagen, Denmark; the Lund University, Department of Geology, Quaternary Sciences, Lund, Sweden; and the Utrecht University, Palaeoecology, Laboratory of Palaeobotany and Palynology, Institute of Environmental Biology, Utrecht, The Netherlands.

The fieldwork on Store Koldewey and Geographical Society Ø was carried out in summer 2003 during the expedition ARKTIS-XIX/4 of the Alfred Wegener Institute for Polar and Marine Research. I participated in this expedition as field assistant.

My research as well as my participation in a number of conferences and meetings, where I contributed with oral presentations, were financially supported by the German Research Foundation, grant no. ME 1169/10. Furthermore my Ph.D. studies benefited from my attendance at several courses and a short study visit at UNIS Longyearbyen that were financially supported by the European Union and the German Academic Exchange Service (DAAD), respectively.

Acknowledgements

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For many fruitful discussions, the friendly atmosphere, and for the unselfish help whenever needed I would like to thank my colleagues P. Hofmann, O. Juschus, V. Wennrich, E. Stefan, the PhD students H. Vogel, S. Berg, S. Ortlepp, and my part-time room mate S. Schmidt. My former colleagues in Leipzig namely Prof. Dr. W. Ehrmann, D. Leuschner, T. Kuhnt, and S. Krüger are thanked for many discussions and their help with literature, access to the laboratory and for the office during the final phase of writing. Special thanks go to the laboratory staff at the universities in Leipzig and Cologne namely S. Dorn, N. Mantke and A. Shahnazarian and the students M. Damascke, F. Boxberg and O. Diecke. Their helping hands allowed the preparation and measurement of large numbers of samples. I would also like to thank the authors and co-authors for their contribution to the results and interpretation presented in this thesis. The Alfred Wegener Institute and the ^{14}C laboratories at the University Lund and University Kiel gave either access to analytical devices or measured samples and/or gave valuable information for the interpretation of the obtained data; therefore, I would like to thank H. Hubberten, L. Schirrmeister, U. Bastian, C. Hjort, G. Skog, P. M. Grootes and M.-J. Nadeau. A. Dehnert, K. Theuerkorn, K. Stolz, R. Pierau, and K. Lehmann shared their knowledge and literature whenever needed - thank you!

Jochen Knies (Norwegian Geological Survey, Trondheim) and Katharina Stolz (University Bremen) read parts of an earlier version of this thesis and gave many comments and suggestions. Cristina Crawley carefully proofread most of the thesis and corrected my English – my sincere thanks are given to them all.

Finally and most important, I wish to express my deepest gratitude to Doreen for her enduring support and her all-encompassing respect for the time I needed to finish this thesis. Furthermore, our daughter Eva missed me during many hours I spent at work - adequate compensation is promised to you both!

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Kurzfassung

Elf größere und kleinere Seen sowie drei Seesedimentsequenzen aus dem Küstenbereich Ost- und Nordostgrönlands wurden hinsichtlich ihrer gegenwärtigen Hydrologie, dem Phytoplankton und im Falle der Seesedimente hinsichtlich der Paläolimnologie untersucht.

Die limnologische Untersuchung der Seen von der Insel Store Koldewey ergab kalte, monomiktische, thermisch durchmischte Wasserkörper mit basischen und oligotrophen Eigenschaften. Planktonische Diatomeen wurden in sechs Seen vorgefunden, wobei die Diatomeen zwischen den Seen variierten und vier Spezies dominierten. Diese Ereignisse liefern einen Beitrag über den gegenwärtigen ökologischen Zustand dieser hocharktischen Seen.

Die Seesedimentsequenzen wurden hinsichtlich der Chronologie, Veränderungen der physikalischen und biogeochemischen Eigenschaften, den darin enthaltenen Makro- und Mikrofossilien, der Korngrößenverteilung und für eine Sequenz der Elementverteilung mittels XRF-Analyse untersucht. Die paläolimnologischen Untersuchungen beziehen sich mit unterschiedlicher Auflösung auf verschiedene Zeitintervalle, wobei besondere Aufmerksamkeit auf die spätquartären Klima- und Umweltentwicklungen entlang der Küste Ost- und Nordostgrönland gelegt wurde. Die Ergebnisse enthalten Hinweise über die Eisrandlage auf Store Koldewey während des Spätweichsels sowie Informationen über die zeitliche und räumliche Entwicklung des grönlandischen Eisschildes westlich der Insel. Des Weiteren geben die Ergebnisse einer Sedimentsequenz Auskunft über die Entwicklung der Umwelt nach dem Eisrückzug und während des Holozäns. Die Bioproduktivität eines weiteren Sees unterlag lang- und kurzfristigen Schwankungen und gibt Auskunft über die klimatische Entwicklung auf Store Koldewey während des Holozäns. Diese Bioproduktivitätsveränderungen sind mit den Veränderungen der klimatischen Bedingungen in Ostgrönland verknüpft, zeigen aber auch die Sensibilität der Bioproduktivität in dem See mit der Veränderlichkeit der Meereisbedeckung im nördlichen Nordatlantik. Eine weitere Sedimentsequenz enthält die ersten in Ostgrönland nachgewiesenen Ablagerungen des Storegga Tsunamis. Diese Ablagerungen beweisen die Auswirkungen dieses Tsunamis in einem weiten Gebiet und zeigen, daß neben den bekannten Fundorten solcher Ablagerungen in Norwegen, Schottland, den Färöer und den Shetland Inseln auch die Küstenbereiche Bereiche Ostgrönlands betroffen waren.

Die hier angeführten Untersuchungen zeigen die Verschiedenartigkeit von Informationen, die aus den Untersuchungen von Wasserproben und Seesedimenten gewonnen werden

können. Abhängig von der Zielsetzung, den angewandten Methoden sowie der genommenen Probendichte liefern die Untersuchungen von Seesedimenten aus dem hocharktischen Bereich Grönlands wertvolle Informationen über Klima- und Umweltveränderungen in der Vergangenheit, welche auch unser Verständnis von damaligen, gegenwärtigen und zukünftigen Umwelt- und Klimaveränderungen nicht nur in den polaren Regionen erweitern. Die Ergebnisse unterstreichen jedoch auch die Schwierigkeiten bei der Interpretation von Seesedimentdaten.

Abstract

Eleven lakes and ponds and three lacustrine sediment sequences from coastal East and North-East Greenland were studied for recent hydrological and phytoplankton characteristics and for palaeolimnology, respectively.

The limnological survey of lakes and ponds from Store Koldewey in summer 2003 revealed cold, monomictic, thermally unstratified, alkaline and likely oligotrophic water bodies. The diatom phytoplankton, present in six lakes and dominated by four species, varied distinctly between the lakes. The results give information about the recent status of the ecology of these High Arctic freshwater bodies.

The sediment sequences were investigated for their chronology, changes in physical and biogeochemical properties, macro- and microfossils, grain-size distribution and in the case of one sequence for its elemental profiles using XRF. The palaeolimnological studies address different time intervals at various temporal resolutions with particular attention to the latitudinal differences of late Quaternary climatic and environmental changes along the coast of East and North-East Greenland. The results include information about the Late Weichselian ice-front environments on southern Store Koldewey with information about the temporal and spatial evolution of the Greenland Ice Sheet margin west of the island. Furthermore, the results and the interpretation of one sediment sequence comprise information about the postglacial and Holocene environmental history of central Store Koldewey. Detailed information about the Holocene climatic history on Store Koldewey are obtained from another record. Changes of the lacustrine biogenic productivity varied on long-term and short-term scale and correspond to the Holocene climatic condition in East Greenland. It also revealed the sensitivity of coastal lacustrine bioproductivity to changes in sea-ice coverage of the northern North Atlantic. Additionally, one sedimentary record contains the first evidence of the Storegga Tsunami deposits of East Greenland. These deposits show the long distance impact of tsunami waves and indicate that besides coastal areas of Norway, Scotland and the Shetland and Faroe Islands also low elevated areas of East Greenland were affected by this event 8200 years ago.

The studies presented here demonstrate the diversity of information obtained from water samples and lacustrine records. Depending on focus, methods applied and sampling resolution investigation of lacustrine records from the High Arctic provides valuable information about past climatic and environmental changes and will contribute to our understanding of previous, recent, and future processes of the environment not only in Polar

Regions. However, the results also emphasize the difficulties associated with the interpretation of lacustrine records.

List of Abbreviations (in alphabetic order)

Alt.	- Altitude
AMS	- Accelerator Mass Spectrometry
a.s.l	- above sea level
Atet	- Aulacoseira tethera
Benth	- Benthic diatoms
BP	- Before Present (i.e. before 1950)
BSi	- Biogenic Silica
cal. kyr	- calendar thousand years
Cc	- Chrysophyte cysts
Corr. age	- Corrected age
Cpse	- Cyclotella pseudostelligera
Cros	- Cyclotella rossii
DIC	- Differential Interference Contrast
Dp	- Diatom phytoplankton
DVC	- Diatom valve concentration
E	- Evenness
EGC	- East Greenland Current
EIC	- East Iceland Current
Ften	- Fragilaria tenera
GISP	- Greeland Ice Sheet Project
GRD	- Gamma Ray Density
GRIP	- Greenland Ice-core Project
H (S)	- Diversity Index
IC	- Irminger Current
IRD	- Ice Rafted Debris
LGM	- Last Glacial Maximum
MS	- Magnetic Susceptibility
Mwd	- Maximum water depth
NADW	- North Atlantic Deep Water
NAO	- North Atlantic Oscillation
NGRIP	- North Greenland Ice-core Project
NP	- Not Present
oPlank	- other Planktonic diatoms
P	- Present
SeDe	- Secchi Depth
SD	- Standard Deviation
SST	- Sea-surface temperature
TC	- Total Carbon
TIC	- Total Inorganic Carbon
TN	- Total Nitrogen
TOC	- Total Organic Carbon
TS	- Total Sulphur
WBD	- Wet Bulk Density
Wd	- Water depth
XRF	- X-ray fluorescence spectrometry

1 Introduction

1.1 State of the Art

During recent decades an increasing number of investigations were carried out to investigate the temporal and spatial variations of the Greenland Ice Sheet and the climatic evolution during the late Weichselian, the onset and procedure of deglaciation after the Last Glacial Maximum, the uplift history of previously glaciated areas and the postglacial climatic and environmental history of Greenland. The existing knowledge is based on the investigation of ice cores, glacial-related geomorphological features, ages of cosmogenic nuclides, and marine and lacustrine sediment sequences (Fig. 1.1).

Studies of ice cores from central and coastal Greenland (Fig. 1.1) give detailed information about temperature changes during the Weichselian and the Holocene. Changes in $\delta^{18}O$ isotope compositions document repeated distinct warming events during the Weichselian and the distinct temperature increase of about 10°C at the Weichselian-Holocene transition (e.g. Johnsen *et al.*, 1992, Johnsen *et al.*, 1995, Alley, 2000, Johnsen *et al.*, 2001, Andersen *et al.*, 2004). About the last 11 kyr are characterized by comparatively more stable and warmer conditions (Dahl-Jensen *et al.*, 1998, Johnsen *et al.*, 2001, Vinther *et al.*, 2006, Vinther *et al.*, 2008) but repeated short-term fluctuations occur of which the cooling event at 8.2 kyr BP with a temperature decrease up to 8°C is the most pronounced (Alley *et al.*, 1997, Johnsen *et al.*, 2001, Thomas *et al.*, 2007).

Investigations of geomorphological features, the occurrence of fossils, cosmogenic nuclides, and marine sedimentary sequences enabled the reconstruction of the glacial extensions and the periods of glacial advances and recessions in North-East (NE) and East (E) Greenland. The Weichselian glacial chronology for NE Greenland with a succession of three glaciations of gradually smaller extent was established by Hjort (1981) and Hjort and Björck (1984) and largely corresponds to the glacial history of E-Greenland (Funder, 1989). Recent investigations of cosmogenic exposure ages revealed a more complex Weichselian glacial history for NE and E Greenland with the presence of less-erosive, local ice caps, and temporal ice-free areas on Jameson Land, Scoresby Sund area (Håkansson *et al.*, 2009) and either local cold-based plateau ice caps or extensive polythermal ice sheets on Store Koldewey (Håkansson *et al.*, 2007). In general, the last glaciation covered almost entirely the present ice-free NE and E Greenland with some unglaciated parts in coastal and high-elevated areas and an ice-margin extending onto the continental shelf (e.g. Hjort, 1979, Hjort, 1981, Hjort and Björck, 1984, Funder, 1989, Björck *et al.*, 1994a, Björck *et al.*, 1994b).

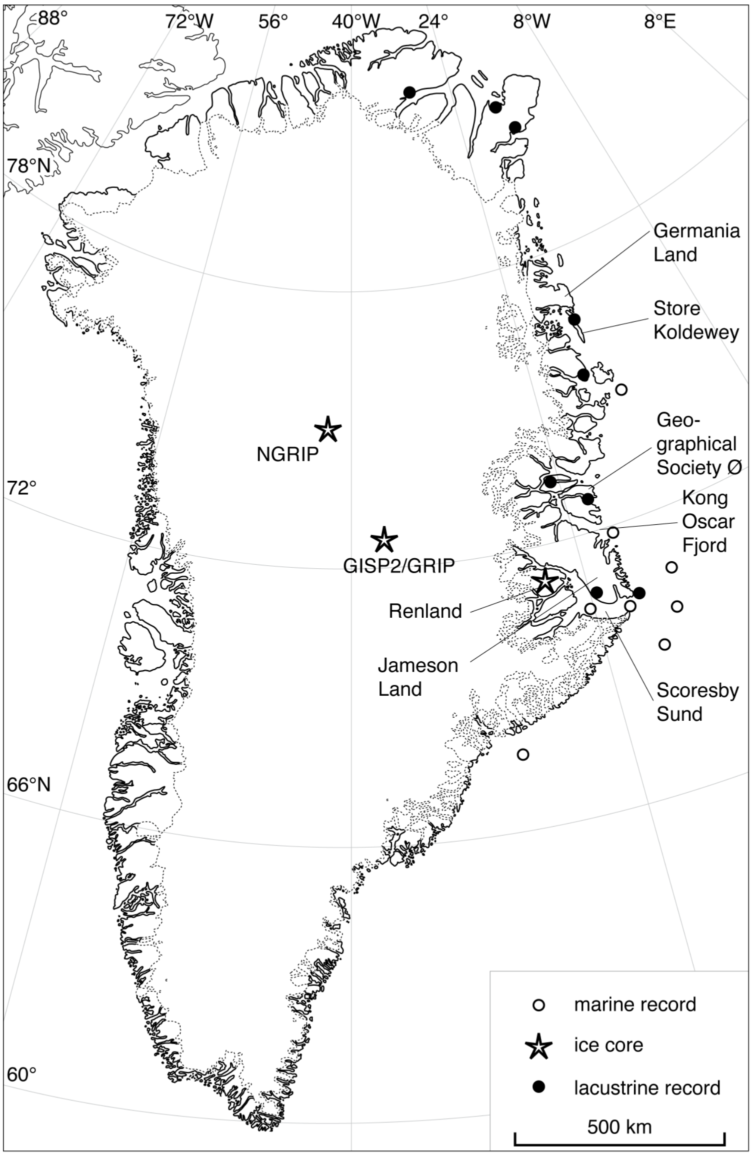

Fig. 1.1: Map of Greenland with present ice margin (dashed line) and positions of selected marine studies, ice cores (NGRIP - North Greenland Ice-core Project, GRIP - Greenland Ice-core Project, GISP2 - Greenland Ice Sheet Project II), and lacustrine sediment records. For references see text.

According to Hjort and Björck (1984) the youngest glaciation corresponds with the Last Glacial Maximum (LGM) during the Late Weichselian (Fig. 1.2).

After the culmination of the Late Weichselian ice sheet the glacial retreat started at ~ 16 cal. kyr BP, however, most of the present ice-free parts of Greenland were not deglaciated until the early Holocene (Bennike and Björck, 2002). Increased Northern Hemisphere insolation (Berger and Loutre, 1991) with rising temperatures (e.g. Johnsen *et al.*, 2001, Andersen *et al.*, 2004) and rising sea-level (Fairbanks, 1989, Hald and Aspeli, 1997, Haflidason *et al.*, 1998, Peltier and Fairbanks, 2006) mainly caused the deglaciation in NE and E Greenland and the shelf and major fjords gradually became ice-free (Dowdeswell *et al.*, 1994, Hubberten *et al.*, 1995, Nam *et al.*, 1995, Funder and Hansen, 1996, Funder *et al.*, 1998). However, the glacial retreat was interrupted during the Younger Dryas and the Preboreal Oscillation when fjord glaciers in central NE and E Greenland re-advanced (Hjort and Björck, 1984, Björck *et al.*, 1997, Funder *et al.*, 1998, Kelly *et al.*, 2008). Subsequently, deglaciation continued until the ice sheet and the glaciers in NE and E Greenland re-advanced in association with decreasing temperatures from about 5 kyr BP and onwards (Funder, 1989, Andrews *et al.*, 1997, Jennings *et al.*, 2002, Reeh, 2004).

The postglacial uplift history of NE and E Greenland is closely interrelated to the deglaciation history and shows longitudinal and latitudinal differences in conjunction with timing of ice-decay and partly with local ice thickness (e.g. Hjort, 1979, Funder and Hansen, 1996, Funder *et al.*, 1998, Bennike and Weidick, 2001, Wagner and Melles, 2002). Highest uplift rates of more than 110 m above present sea-level have been reconstructed for the inner Kong Oscar Fjord and the inner Scoresby Sund, coastal areas raised up to 40 m (Hjort, 1979, Funder and Hansen, 1996, Funder *et al.*, 1998).

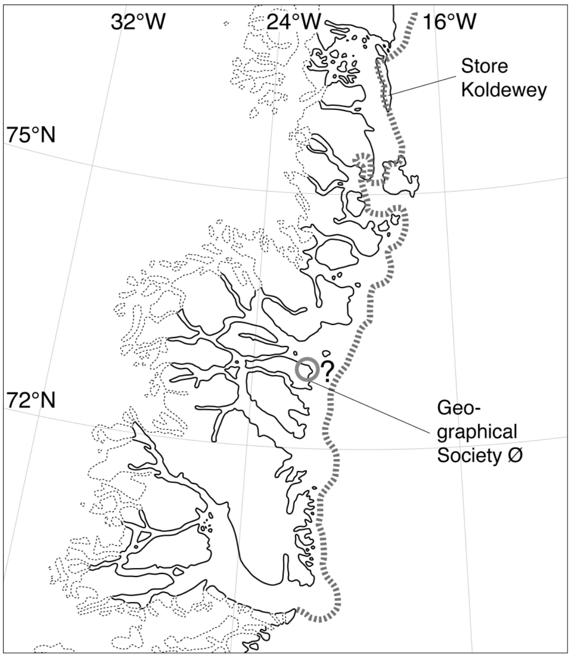

Fig. 1.2: North-East Greenland Fjord zone with present coastline (full line), present ice margin (thin dashed line) and reconstructed ice margin during the Late Weichselian (thick dashed line) according to Hjort (1981), Hjort and Björck (1984), Funder (1989), and Landvik (1994) with suggested ice-free areas on Store Koldewey and on Geographical Society Ø (circle).

In contrast to marine records off the coast of NE and E Greenland (e.g. Dowdeswell *et al.*, 1994, Hubberten *et al.*, 1995, Nam *et al.*, 1995, Funder *et al.*, 1998), sediment sequences from presently existing lakes have a much higher time resolution, similar to those of ice cores, and often provide reliable minimum ages for the deglaciation (Cremer *et al.*, 2008). In addition, lake sediment records provide high-resolution information on climatic and environmental changes on NE and E Greenland, which supplements knowledge from ice cores and marine records (Wagner and Melles, 2001). The existing lacustrine sediment records from NE and E Greenland (Fig. 1.1) were predominantly investigated for palynology, partly supplemented with sedimentological, biogeochemical and fossil analyses. Studies were carried out on lake sediments from the Scoresby Sund region (Funder, 1978, Björck *et al.*, 1994a, Cremer *et al.*, 2001, Wagner and Melles, 2001), Geographical Society Ø (Wagner *et al.*, 2000, Klug and Wagner, 2008), Ymer Ø (Wagner and Melles, 2002), on Hochstetter Foreland (Björck and Persson, 1981, Björck *et al.*, 1994b), on Store Koldewey (Wagner *et al.*, 2008) and in eastern North Greenland (Fredskild, 1969, Funder and Abrahamsen, 1988, Fredskild, 1995). Early studies from Scoresby Sund region and eastern North Greenland focused mainly on changes in the pollen assemblages; ensuing studies used a multidisciplinary approach and detailed information about the deglaciation, the postglacial uplift but especially about the climatic and environmental history during the Holocene were obtained (Wagner *et al.*, 2000, Wagner and Melles, 2001, Wagner and Melles, 2002, Wagner *et al.*, 2008). However, latitudinal differences in the onset and in the duration of warm and cold periods during the Holocene apparently exist.

Lacustrine sediment sequences from Store Koldewey and Geographical Society Ø are of special importance for a better understanding of latitudinal differences in the postglacial climatic and environmental development between the relatively well-studied sequences from central E Greenland (Funder, 1978, Björck *et al.*, 1994a, Wagner and Melles, 2001) and the few existing sequences from eastern North Greenland (Fredskild, 1969, Funder and Abrahamsen, 1988, Fredskild, 1995) but also between the high-resolution ice core records from inland and the relatively low-resolved marine records offshore. Additionally, parts of both islands were suggested to have remained ice-free during the LGM (Funder and Hjort, 1973, Hjort, 1979, Hjort, 1981). Therefore lacustrine sedimentary records from these islands may reach back far into the Younger Dryas epoch or even into earlier times such as the LGM. Lakes, ponds and rivers are prominent features of high arctic landscapes. These ecosystems have a simplified ecology and are subject to natural and human-induced climate and environmental change (Vincent and Hobbie, 2000) but have been poorly investigated in the High Arctic.

1.2 Aims of the Studies

This thesis comprises the investigation of eleven lakes and ponds from Store Koldewey for their recent hydrology and the diatom phytoplankton and palaeolimnological investigations of three sediment sequences from Store Koldewey and Geographic Society Ø. The presented studies attempt, (i) to describe the recent conditions concerning the hydrology and diatom phytoplankton of lakes in the coastal High Arctic, (ii) to verify the existence of ice-free regions on the outer coast of NE and E Greenland during the Last Glacial Maximum, (iii) to reconstruct the onset and pattern of the deglaciation after the LGM, and (iv) to reconstruct the climatic and environmental changes on the outer coast of NE and E Greenland during Late Weichselian and Holocene times.

2 Study Area

2.1 Geography and Geology

The lakes studied for recent hydrology and palaeolimnology are located on two coastal islands in the North-East Greenland Fjord zone. The Fjord zone comprises the present ice-free area between Germania Land in the north and Scoresby Sund in the south. It is bordered in the west by the present ice margin of the Greenland Ice Sheet and towards the east by the northern North Atlantic (Fig. 2.1). The area is characterised by numerous fjords of which some are about 300 km long and extend to the calving front of outlet glaciers of the Greenland Ice Sheet. The interfjord plateaus are of 1500 to 2000 m elevation with the

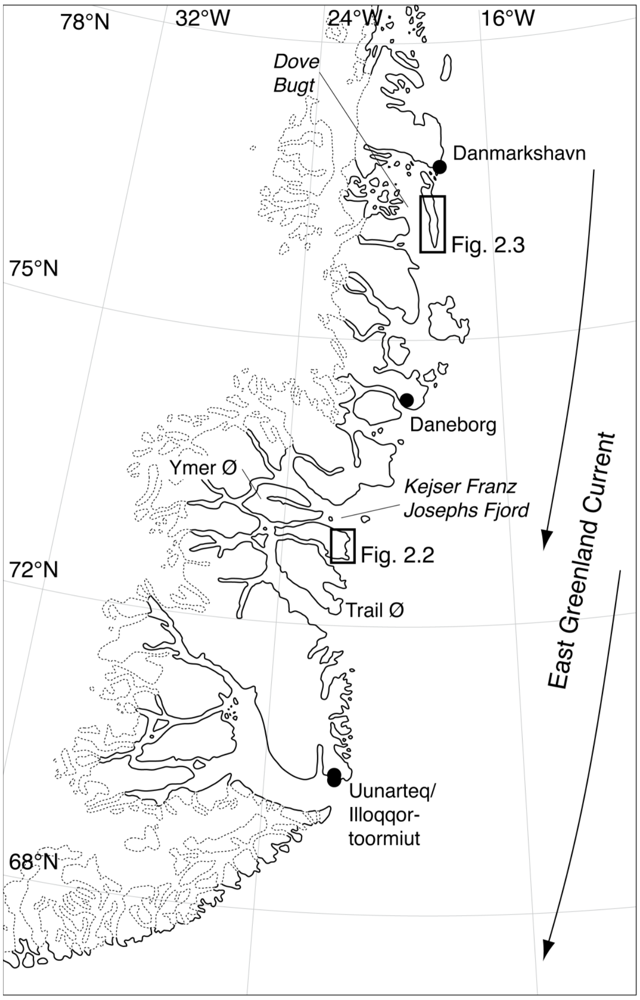

Fig. 2.1: North-East Greenland Fjord zone with present ice margin (dashed line), positions of weatherstations (dots) and localities mentioned in the text.

occurrence of local ice caps on higher altitudes. From the inland elevations the terrain gently slopes to partially extensive coastal lowlands (Funder, 1989). The general geology of East Greenland, comprising an intermittent succession of crystalline, sedimentary and volcanic rocks throughout approximately the last 2 billion years, has been described in detail by Henriksen *et al.* (2000) and can be summarised as follows: Mid to Late Proterozoic and Early Palaeozoic sediments, deformed and metamorphosed during the Caledonian orogenese with mainly north to south orientated structural elements, compose the crystalline basement in East Greenland. Continental to marine-epicontinental sediments, including clastic sediments, carbonates and evaporites of Carboniferous to Mesozoic age, overlie the crystalline basement. Tertiary volcanic rocks, in form of plateau basalts, sills and dykes, are widespread. Quaternary sediments, including discontinuous tills and erratic boulders, cover vast areas of the present ice-free land; in coastal areas raised marine sediments are common.

Geographical Society Ø is located off East Greenland at 72°40'-73°04'N and 021°52'-024°35'W (Figures 2.1 & 2.2). The E-W orientated island is about 90 km long and 25 km wide. Small sounds separate the island from Ymer Ø and Trail Ø towards the north and the east Geographical Society Ø is bordered by the Kejser Franz Josephs Fjord and the Greenland Sea, respectively. The topography of the central and western part of the island is dominated by a mountainous landscape with highest elevations of up to 1700 m above sea level (a.s.l.). The eastern and southern parts are characterised by a hilly topography, descending towards the coast (Fig. 2.2). The drainage system is adapted to the topography and consists of a large number of small rivers and streams. Lakes and ponds of various dimensions fill basins on the island. Geographical Society Ø is formed by Devonian and Carboniferous sediments in its western part and Triassic and Cretaceous sediments in the eastern part (Henriksen, 2003). The Mesozoic sediments consist of

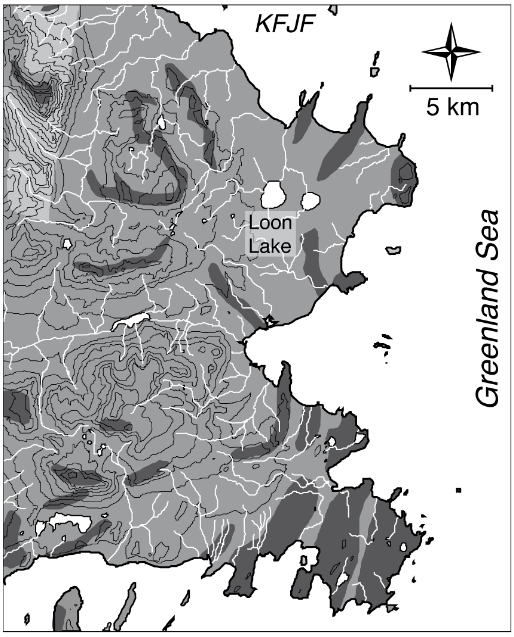

Fig. 2.2: Eastern Geographical Society Ø with topography (100 m contour lines), drainage system and major lakes (white lines and areas) and simplified bedrock geology after Henriksen *et al.* (2003) with Triassic (light gray) and Cretaceous (medium gray) sediments and Tertiary basalts (dark gray). Kejser Franz Josephs Fjord (KFJF) to the north.

coarse to fine-grained sandstones, mudstones and shale. Tertiary volcanic intrusions in form of sills and dykes are common in the eastern part of the island and crop out in form of ridges or plateaus. Unconsolidated marine sediments from the Cenozoic occur on the low altitude regions of the island (Bennike *et al.*, 2004).

Store Koldewey is located off North-East Greenland at 76°45'-75°55'N and 019°10'-018°27' W (Figures 2.1 & 2.3). It is separated from the Greenland mainland by the Dove Bugt in the west; towards the east the island is bordered by the Greenland Sea. The topography of Store Koldewey is dominated by a 3 to 5 km wide, flat-topped mountain range along the north–south axis with an elevation of up to 900m a.s.l. Steep slopes descend down to the shoreline on the western part of the island whereas a 2 to 5km wide plain at c. 150m a.s.l characterises the eastern part of the island (Fig. 2.3). The bedrock of the mountain range is Precambrian metamorphic rock, mainly gneiss. Jurassic and Cretaceous marine sediments occur on the eastern plain (Henriksen, 2003). Quaternary sediments, such as till and raised marine and littoral deposits, are widespread (Bennike *et al.* 2004). Lakes and ponds of various dimensions fill depressions on Store Koldewey.

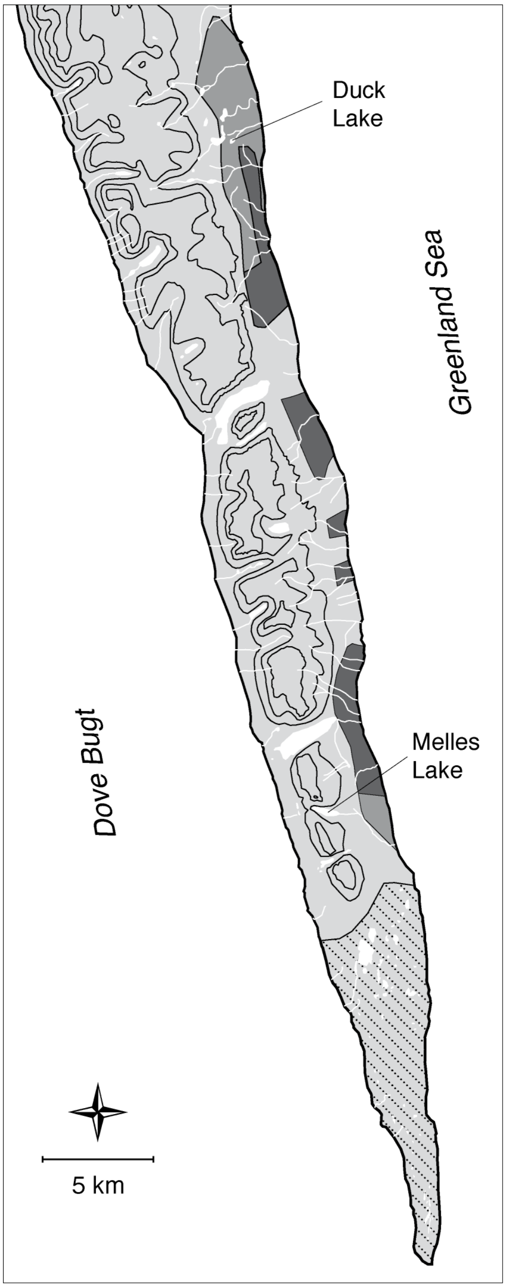

Fig. 2.3: Central and southern Store Koldewey with topography (200 m contour lines), drainage system and lakes (white lines and areas) and simplified bedrock geology after Henriksen *et al.* (2003) with Precambrian gneisses (light gray) and Jurassic (medium gray) and Cretaceous (dark gray) sediments; hatched area covered by quaternary sediments.

2.2 Present Climate and Biota

Along the coast of NE and E Greenland gradients in winter temperature and precipitation with colder and dryer conditions in the north occur (Fig. 2.4). These latitudinal differences are mainly related to atmospheric and oceanic circulation and annual insolation. Because of highest air pressure located in the coldest areas to the west and northwest, northerly winds are predominant in the winter (Cappelen *et al.*, 2001) and provide cold air downwards along the coast. Mean January temperatures at Danmarkshavn (76°46'N; 018°40'W) are approximately -25 °C whereas, about 700 km further to the south at Ittoqqortoormiut (Scoresbysund, 70°29'N; 021°57'W), temperatures in January average at approximately -15°C (Fig. 2.4). In contrast, July temperatures of about 4°C were measured at the three weather stations Danmarkshavn, Daneborg (74°18'N; 020°13'W) and Ittoqqortoormiut (Fig. 2.4). Although summer insolation increases towards the north, warming of the coastal area is hampered due to the influence of the East Greenland Current (EGC, Fig. 2.1), that flows along the east coast and transports cold, low saline polar waters and large quantities of ice from the Arctic Ocean through Fram Strait to Cape Farwell in the south. The cold waters also cause the condensation of water vapour in the air resulting in sea fog, a common phenomena from May to September in the coastal areas (Cappelen *et al.*, 2001). Precipitation in E and NE Greenland is related to the low pressure activity that moves moist air masses from southwest to the northeast

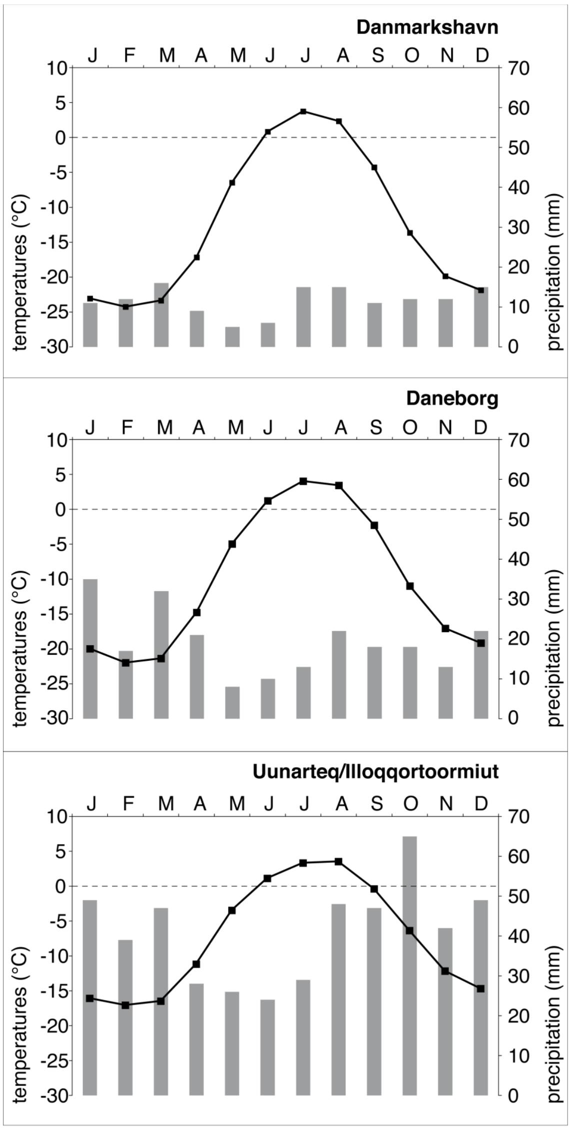

Fig. 2.4: Mean monthly temperatures (°C; curves) and mean accumulated precipitation (mm, columns) at weatherstations along the East Greenland coast according to intermittent data series between 1958 and 1990 after Cappelen *et al.* (2001).

along the south Greenland coast. Mean annual precipitation at Uunarteq (Kap Tobin 70°25’N;021°58’W, 7 km south of Scoresbysund) is about 500 mm with lowest precipitation rates during spring. At Daneborg, about 450 km north of Scoresbysund, the annual precipitation is about 230 mm and another 250 km further north at Danmarkshavn weather station, only 141mm were annually precipitated, mainly from drifting snow (Cappelen et al., 2001).

The vegetation in E and NE Greenland is adapted to the climatic conditions and is, according to the bioclimate zonation of Bliss (1997), High Arctic. Higher summer temperatures as occur in the inner fjord regions and/or increased precipitation as occur in E Greenland allow denser vegetation with plant cover of up to 80 % and hemi prostrate dwarf shrubs, such as *Cassiope* spp., grasses and sedges dominate the vegetation. In coastal areas or further to the north, plant cover becomes more open and patchy; dwarf shrubs are less abundant and mosses and lichens are more dominant (CAVM Team, 2003). The fauna in E and NE Greenland is represented by various migratory birds, and mammals like polar hare, polar fox, and polar bear. In areas with increased vegetation cover musk oxen are more frequent (Bennike *et al.*, 2004).

2.3 Lakes studied for Palaeolimnology

Loon Lake

Loon Lake is situated at 18m a.s.l. on north-eastern Geographical Society Ø, East Greenland in a undulating landscape with maximum elevations of up to 50m a.s.l.. Towards the west of the lake, the landscape gently ascends to 500 m a.s.l. The lake fills an almost circular basin of approximately 1.5 km in diameter (Fig. 2.5) with a maximum water depth of 11.7m in the centre of the lake. The estimated catchment area is about 15-20 km^2. Two smaller inlets on the western and north-western lakeshore and the major inlet on the southern lakeshore enter the lake. The outlet is in the north-eastern part of the lake and drains towards the Kejser Franz Joseph Fjord.

Duck Lake

Duck Lake is located at 118m a.s.l. on the eastern plateau on central Store Koldewey, North-East Greenland (Fig. 2.5). The surrounding area is characterised by an undulating surface. The lake fills a small basin of about 0.032km^2 and is mainly fed by meltwater from surface runoff of its catchment area of about 0.3km^2. The outflow is towards the northeast into the Greenland Sea.

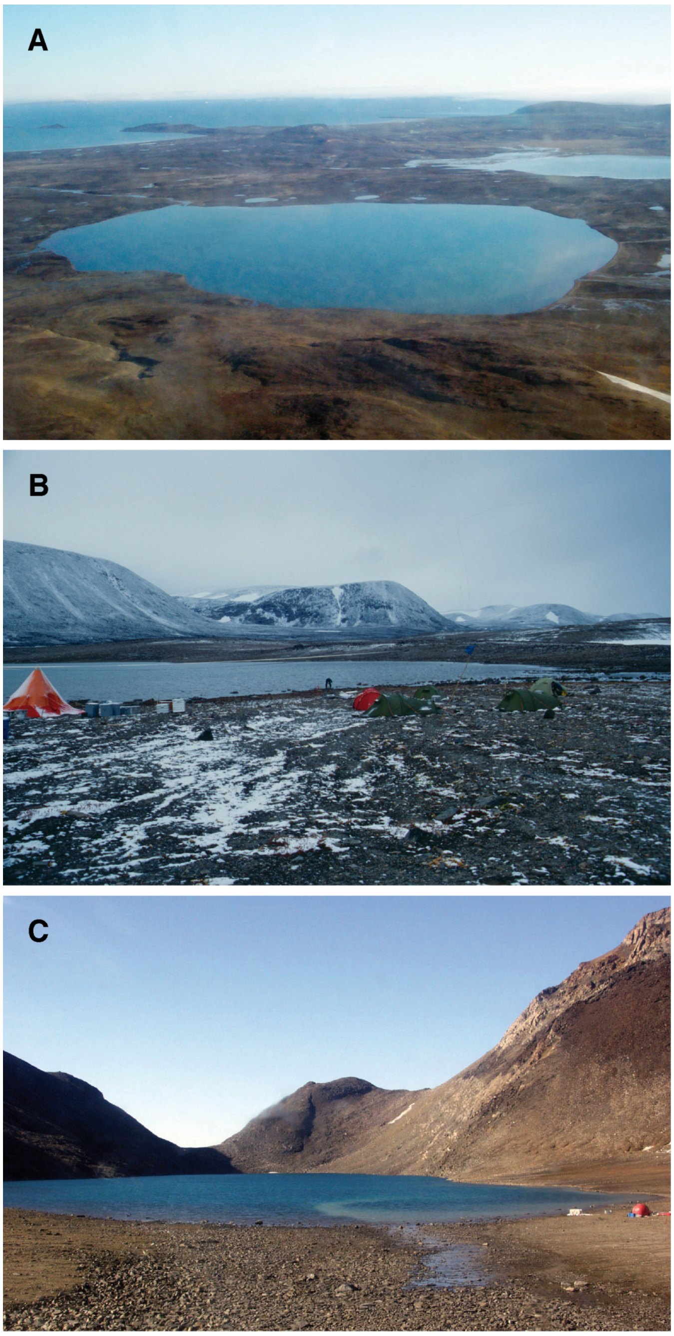

Fig. 2.5: Lakes investigated for palaeolimnology; A) View on Loon Lake on Geographical Society Ø with Kejser Franz Josephs Fjord in the background, B) Northern part of Duck Lake on Store Koldewey with the mountain range in the background, C) View on Melles Lake on southern Store Koldewey, surrounding slopes are covered by blockfields.

Melles Lake

Melles Lake fills an E-W elongated, c. 1.5 km long and c. 0.35 km wide basin on southern Store Koldewey at an elevation of 166 m a.s.l. Steep slopes surround the lake in the west whereas its eastern part stretches to the lower plateau (Fig. 2.5). The western slopes surrounding Melles Lake are covered by blockfield, the eastern plateau by fine to coarse sediments forming raw to initial soils. The lake is exclusively fed by meltwater from its estimated catchment area of about 1.6 km^2 and drains towards the east.

3 Methods

3.1 Field Work

Field work on Store Koldewey and on Geographical Society Ø was conducted in August and September 2003. Bathymetrical measurements were carried out from a floating platform along crossing profiles using a hand-held echo sounder and a GPS. Water samples were taken along vertical profiles in deepest parts of the lakes using a 5 L water sampler (UWITEC Corp., Austria). Following sampling, temperature (WTW Oxi196 probe; WTW GmbH, Weilheim, Germany), oxygen saturation and concentration (WTW Oxi196 probe), conductivity (WTW LF197 probe) and pH (pHScan WP2 tester, Eutech Instruments Ltd., Singapore) were measured and water samples prepared for laboratory analyses (Fig. 3.1). In deeper lakes the transparency was recorded using a Secchi disc.

The sediment cores were recovered using a piston and a gravity corer (both UWITEC Co., Austria) from a floating platform. The gravity corer was used to retrieve undisturbed surface sediments. The corer is equipped with a 60 to 120 cm long PVC plastic tube of 6 cm in diameter. To obtain deeper and especially longer sediment sequences the piston corer was used. This system is operated via a platform-mounted tripod with three winches and steel ropes (Fig. 3.2). The first rope controls the release of the piston, the second the vertical position of the coring tube in the water column, and the third operates a hammer. A metal

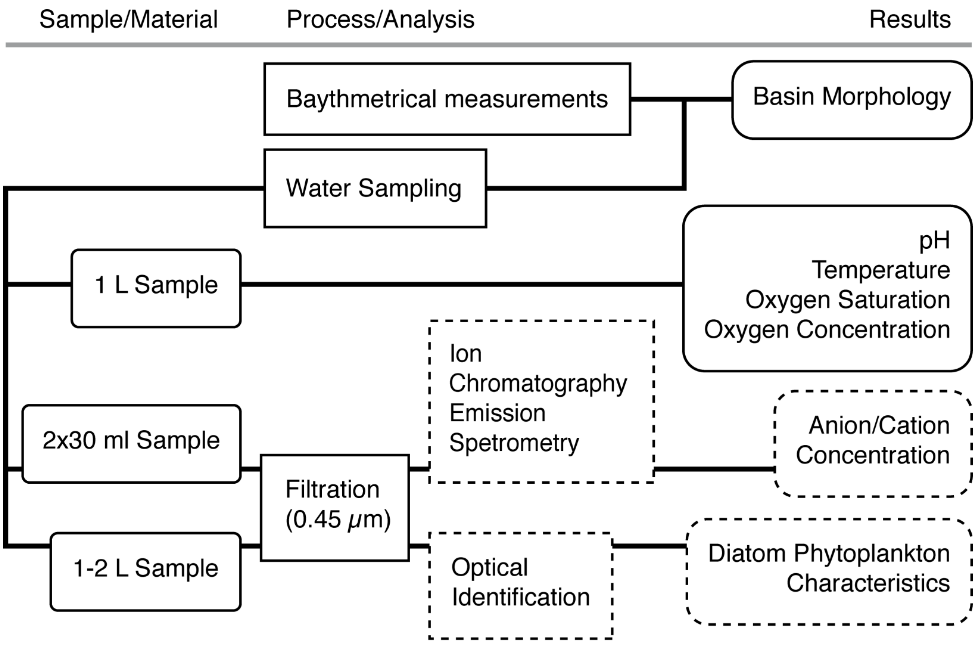

Fig. 3.1.: Overview of bathymetrical measurements, water sampling and related results carried out in the field (full frames) and in the laboratory (dotted frames).

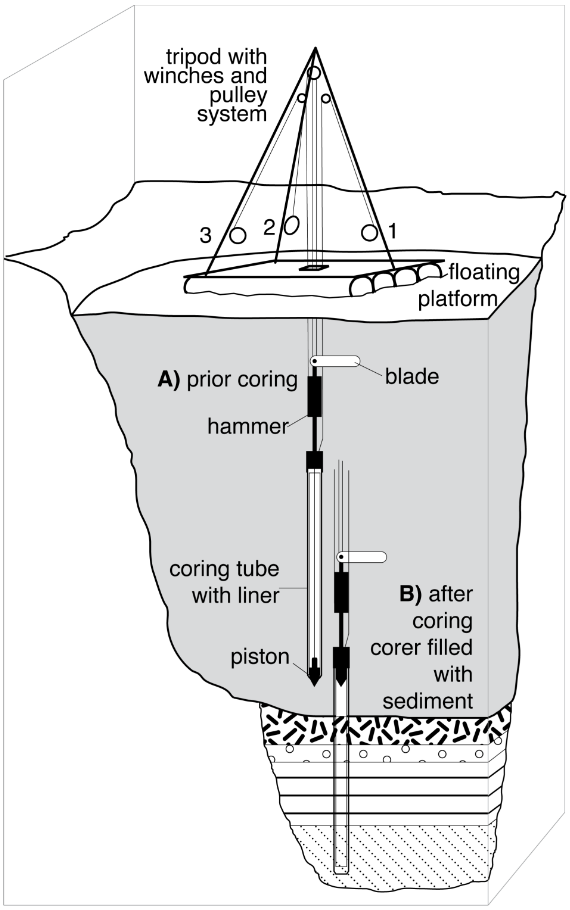

Fig. 3.2: Sketch of sediment coring procedure. Winches and steel ropes (1, 2, 3) give control in the coring process. For deeper sequences the coring tube has to be hammered to the intented sediment depth before the piston is released.

blade on the top of the coring unit prevents twisting of the coring equipment (Fig. 3.2). The piston corer consists of a 3.30 m metal tube, which contains a 3 m long plastic tube of 6 cm in diameter. For the uppermost sediment core, the piston has to be released c. 50 cm above the sediment surface and the corer is hammered into the sediment by a driving weight. The sediment then becomes gradually captured in the plastic tube and after successful coring the corer contains about 2.5 m of sediment (Fig. 3.2). For deeper sediments, the coring system has to be hammered into the sediment before the piston is released. A minimum overlap of 50 cm enables a correlation between the single 3 m segments. This procedure can be repeated until stiff sand and coarse layers and/or too consolidated sediments stop further penetration. A more detailed description of the coring procedure is given in Melles *et al.* (1994).

3.2 Laboratory Work

3.2.1 Water samples and surface sediments

Anion and cation concentrations were measured with an ion chromatograph (DX320, DIONEX Corp., USA) and an emission spectrometer (ICP-OES Optima 3000 XL, Perkin-Elmer Inc., USA), respectively. The hydrogen carbonate content was examined through HCl titration.

For diatom phytoplankton analysis 1.0 to 2.0 litres of lake water was filtered through cellulose acetate filters (pore size 0.45 µm). After pre-treatment with 30 % hydrogen peroxide and 65 % nitric acid at 100 °C, removal of excess acid and diminution of sample quantity to 50 ml slides were prepared according to methods by Battarbee (1973) with

Naphrax® as mounting media. Diatoms were identified and counted using an Olympus BX51 microscope equipped with a UPlanFl 100x/1.30 oil immersion objective with differential interference contrast (DIC). For diatom phytoplankton concentrations and relative species abundances 200–600 diatom valves were counted on each slide.

3.2.2 Sediment cores

Sediment sequences from three lakes were investigated and according to the focus of the study different analyses were applied. Geophysical properties as magnetic susceptibility (MS), wet bulk density (WBD, i.e. gamma-ray density (GRD)) were measured at 1 cm intervals on the whole-core segments with a Multi Sensor Core Logger (MSCL; Geotek, Corp., Germany) (Weber *et al.*, 1997). After core opening and sediment description sediment optical and micro-radiographic images and X-ray fluorescence spectrometry (XRF) elemental profiles were measured with an ITRAX core scanner (COX Analytical Systems, Sweden). The sediment sequences were sub-sampled into 0.5, 1 or 2 cm intervals and the water content was calculated after freeze-drying from the mass difference between the wet and the freeze-dried sample. In order to perform all analyses at the same sample horizon samples were split into aliquots.

For biogeochemical measurements samples were ground to homogenise the material. Total carbon (TC), total nitrogen (TN) and total sulphur (TS) content were quantified using a VARIO EL III analyser (ELEMENTAR Corp., Germany). Total organic carbon (TOC) was determined using a METALYT CS1000S analyser (ELTRA Corp., Germany) after removal of carbonate with 10% HCl at 80°C. Total inorganic carbon (TIC) was calculated from the difference between TC and TOC. TOC/TN and TOC/TS ratios were calculated using TOC, TN and TS mass percentages, respectively. Biogenic silica (BSi) content was measured with the SAN++ Automated Wet Chemistry Analyser (SKALAR Co., Netherlands) according to the method described by Müller and Schneider (1993) and the biogenic silica content was calculated from the extrapolated intercept concentration (DeMaster, 1981). For determination and enumeration of macrofossils sample material was wet sieved and the residue analysed using a dissecting microscope.

After removal of organic matter with H_2O_2, grain size analyses were performed on fractions <1mm or <1.25mm from individual sediment sequences with laser diffraction particle size analysers (Cilas 1180L, Success Scientific Corp., Taiwan; Fritsch Co., Germany; and LS 200, Beckmann Coulter Inc., USA). The coarser fraction (>1mm) was weighed and is given as weight percentage of the entire sample.

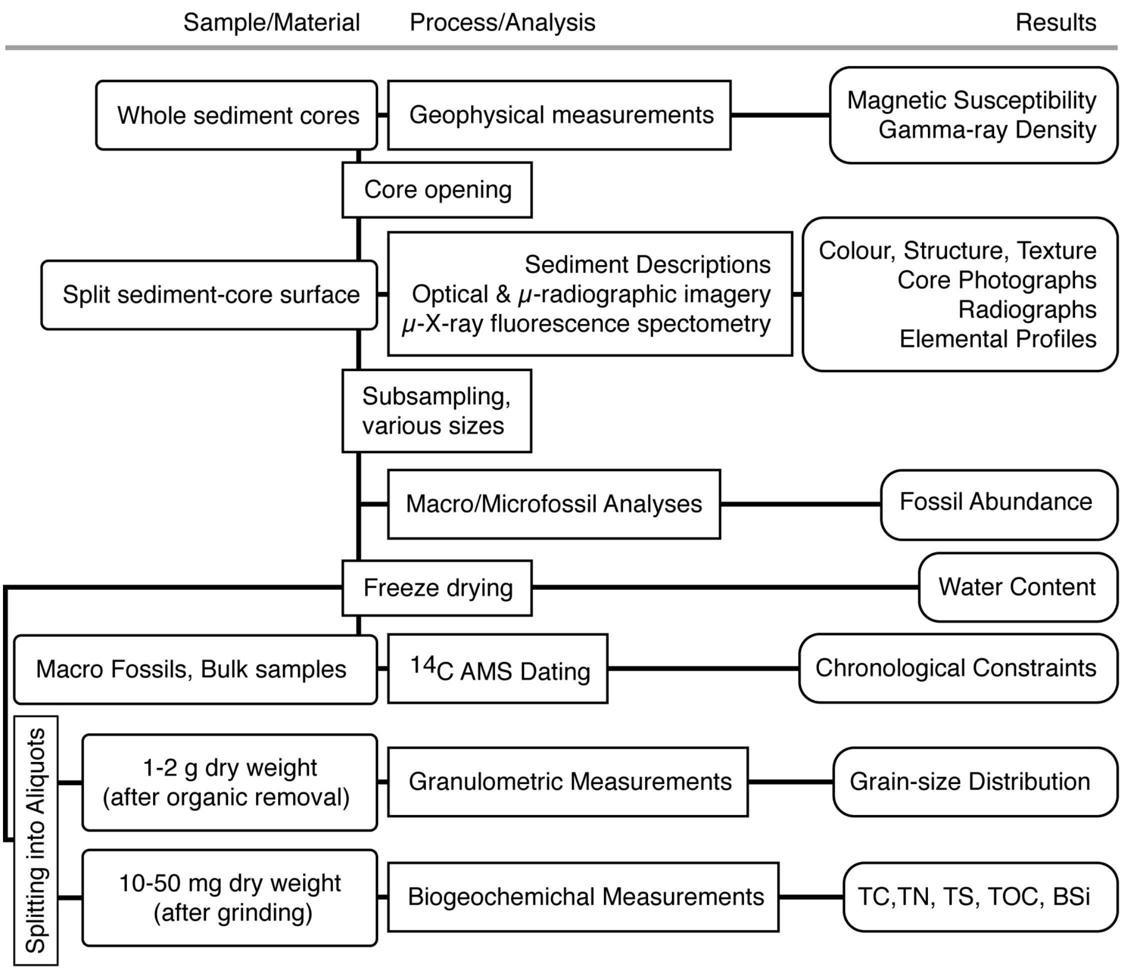

Fig. 3.3.: General Overview of samples and material, processes and analyses applied in the laboratory and related results conducted for palaeolimnological studies.

Radiocarbon dating of terrestrial and marine fossils and of bulk sediment samples was performed using ^{14}C Accelerator Mass Spectrometry (AMS) at the Leibniz-Laboratory for Radiometric Dating and Stable Isotope Research, University of Kiel and the Radiocarbon Dating Laboratory of the GeoBiosphere Science Centre at the University of Lund. ^{14}C ages of marine samples were calibrated into calendar years (cal. yr BP) after subtracting 550 yr (delta R=141; Rasmussen and Rahbek, 1996) for the local marine reservoir effect and using the CALIB5.0 program and the MARINE04 dataset (Stuiver and Reimer,1993; Hughen et al., 2004). ^{14}C ages of terrestrial fossils and bulk sediment samples were calibrated into calendar years before present (cal. yr BP) using the calibration program Calib version 5.0.2 (Stuiver and Reimer, 1993) based on the calibration data set IntCal04 (Reimer et al., 2004) and using CalPal (Danzeglocke *et al.*, 2008) and the calibration data CalPal-2007$_{Hulu}$ (Danzeglocke *et al.*, 2008, Weninger and Jöris, 2008).

4 Contribution of Martin Klug to the chapters 5 - 9

5 Hydrology and Diatom Phytoplankton of High Arctic Lakes and Ponds on Store Koldewey, Northeast Greenland

Holger Cremer wrote the text jointly with the co-authors. I assisted during fieldwork and sampling and contributed to the discussion and interpretation.

6 First indication of Storegga tsunami deposits from East Greenland

Within the framework of my diploma thesis in 2004 I conducted a pilot study of the entire sediment sequence from Loon Lake (results published in Klug and Wagner, 2008) and supplied a data set. Bernd Wagner wrote the text jointly with the co-authors; I contributed to the discussion and the interpretation.

7 Lake sediments from Store Koldewey, Northeast Greenland, as archive of Late Pleistocene and Holocene climatic and environmental changes

The study was planned jointly. I wrote the text jointly with Bernd Wagner and Ole Bennike. Steffi Schmidt and Oliver Heiri analysed and interpreted the chironomid assemblages.

8 Repeated short-term bioproductivity changes in a coastal lake on Store Koldewey, North-East Greenland, an indicator of varying sea-ice coverage?

I planned the study in cooperation with the supervisors and wrote the text jointly with Ole Bennike and Bernd Wagner.

9 Late Weichselian ice-front history on Store Koldewey, North-East Greenland, as illustrated by a lacustrine record

I planned the study in cooperation with the supervisors and wrote the text jointly with the Bernd Wagner.

5 Hydrology and Diatom Phytoplankton of High Arctic Lakes and Ponds on Store Koldewey, Northeast Greenland *

5.1 Introduction

Freshwater ecosystems in the High Arctic are poorly investigated and understood compared to the enormous coverage of the northern polar region with freshwater reservoirs. However, lakes, ponds and rivers are prominent features of high arctic landscapes with a simplified ecology, and, as their counterparts in mid and low latitudes, these ecosystems are subject to natural and human-induced climate and environmental change (Vincent and Hobbie, 2000). Polar ecosystems are generally limited by the harsh climatic conditions such as constantly low temperatures and extremes in irradiance, seasonality and freeze-thaw cycles. These conditions have a distinct effect on the habitat structure, species diversity and food web (Born and Böcher, 2001; Vincent and Hobbie, 2000).

The geographical coverage of freshwater ecosystem studies in the Arctic is poor. Whereas a fairly good data base of limnological surveys exists for arctic Canada (Antoniades *et al*., 2003, and references therein) and northern Alaska (Hobbie, 1984), modern limnological data are rare for arctic Siberia (e.g., Cremer and Wagner, 2003; Duff *et al*., 1999) and Greenland. In Greenland, limnological research is mainly restricted to West Greenland (e.g., Willemse, 2002, and references therein), although limnological research in the eastern part of Greenland has also a long history, being pioneered during the Denmark Expedition in 1906-08 (Trolle, 1913). Much work has been done on the flora and fauna of lakes and ponds in the region (e.g., Østrup, 1910; Røen, 1962). In 1999 a limnological monitoring programme was initiated at the Zackenberg Station (Fig. 5.1; Caning and Rasch, 2000).

Diatom floras in high arctic lakes and ponds are likewise poorly documented (see Douglas and Smol, 1999 and Cremer *et al*., 2001a for detailed reviews of available literature) and particularly, lake phytoplankton data for East Greenland are scarcely published (e.g., Rasch and Caning, 2003). However, based on their siliceous cell walls and good preservation potential in the fossil record, diatoms are a powerful and, meanwhile, routinely applied indicator of long-term environmental and climate change in polar freshwater ecosystems (Douglas and Smol, 1999; Spaulding and McKnight, 1999; Smol and Cumming, 2000).

*This chapter, including figures and tables, is part of the publication: Cremer H, Bennike O, Håkansson L, Hultzsch N, Klug M, Kobabe S and Wagner B. 2005. Hydrology and diatom phytoplankton of high arctic lakes and ponds on Store Koldewey, Northeast Greenland. *International Review of Hydrobiology* **90**, 84-99. It is available at: http://www3.interscience.wiley.com/

In this study, we provide hydrological and phytoplankton characteristics for nine lakes and two ponds on Store Koldewey, a previously poorly investigated island in Northeast Greenland. This research is part of an interdisciplinary, multiproxy study of Pleistocene and Holocene climate and environmental change in Northeast Greenland (Bennike *et al.*, 2004). The island was visited during a fieldwork campaign in August and September 2003.

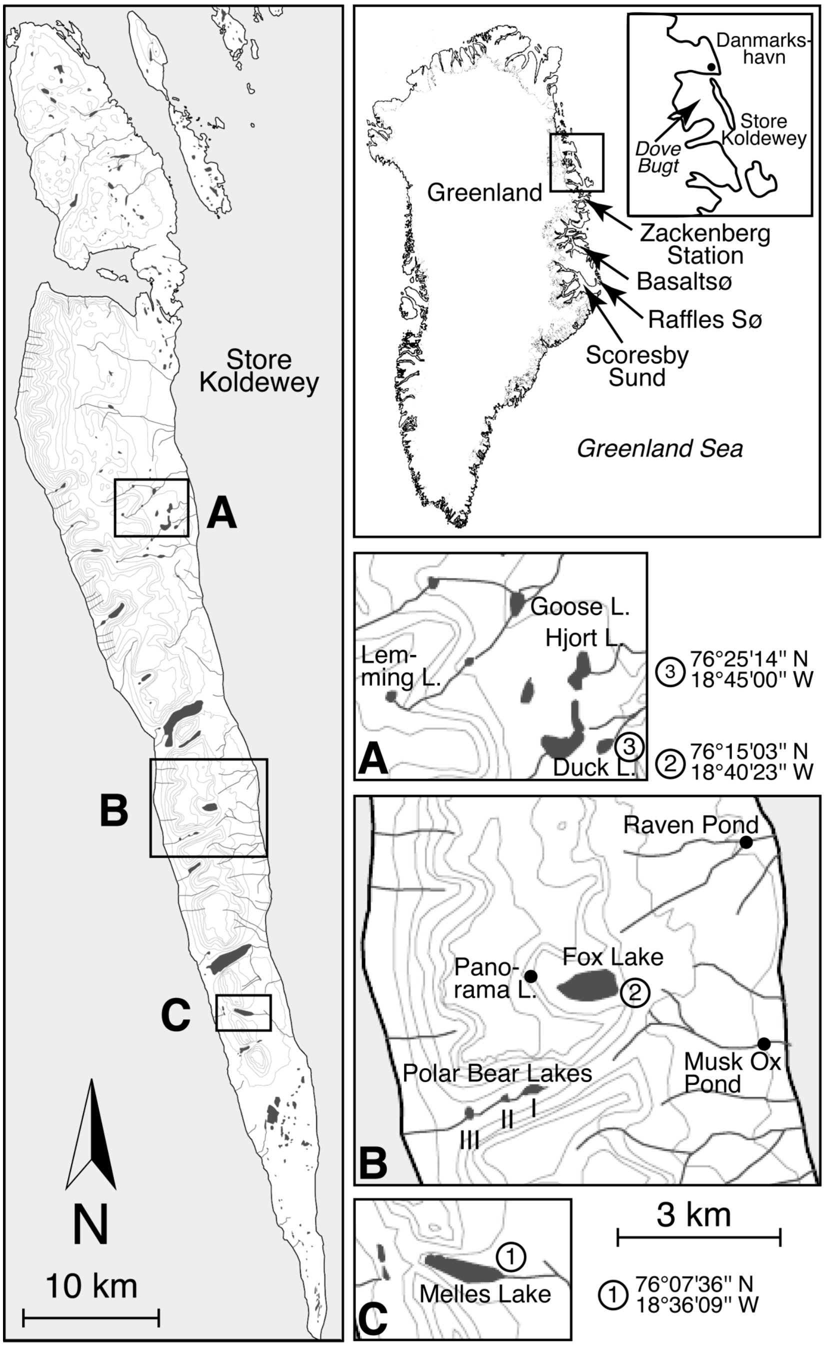

Fig. 5.1: Geographical location of the studied lakes and ponds on Store Koldewey, Northeast Greenland.

5.2 Material and Methods

5.2.1 Study Area

Store Koldewey is an elongated island off Northeast Greenland at 75°55'–76°45' N and 18°27'–19°10' W (Fig. 5.1). The island is *ca.* 80 km long and has a maximum width of *ca.* 10 km, and is separated from the Greenland mainland by Dove Bugt. Geologically, Store Koldewey is part of the East Greenland Caledonian Fold Belt and mainly composed of Precambrian metamorphic rocks (gneisses) older than 1800 Ma (Escher and Pulvertaft, 1995). Jurassic and Cretaceous marine sediments are exposed at the east coast of the island (Escher and Pulvertaft, 1995) and Quaternary deposits are widespread above these Mesozoic sediments (Bennike *et al.*, 2004).

Topographically, Store Koldewey is dominated by a north-south oriented, flat-topped mountain range up to 900 m a.s.l. (Figs. 5.1 and 5.2). The west coast of the island is characterized by steep slopes close to the shoreline (Fig. 5.2), whereas plains with a maximum elevation of *ca.* 150 m a.s.l. extend along the east coast (Fig. 5.3). Shallow lakes and ponds occur in geomorphological depressions on these plains (Figs. 5.3A and 5.3E–H). The mountain range is repeatedly cut by west-east oriented, U-shaped valleys with steep slopes in which lakes of various water depths are found. The investigated nine lakes (two deep and seven shallow lakes) and two ponds on Store Koldewey are located in three areas (Fig. 5.1, Table 5.1) and are formed as kettle lakes, cirque lakes (Fox Lake and Melles Lake;

Fig. 5.2: North-ward view along the mountain range at the west coast of Store Koldewey.

Figs. 5.3A and 5.3B) or so-called paternoster lakes (Polar Bear Lakes; Fig. 5.3C).

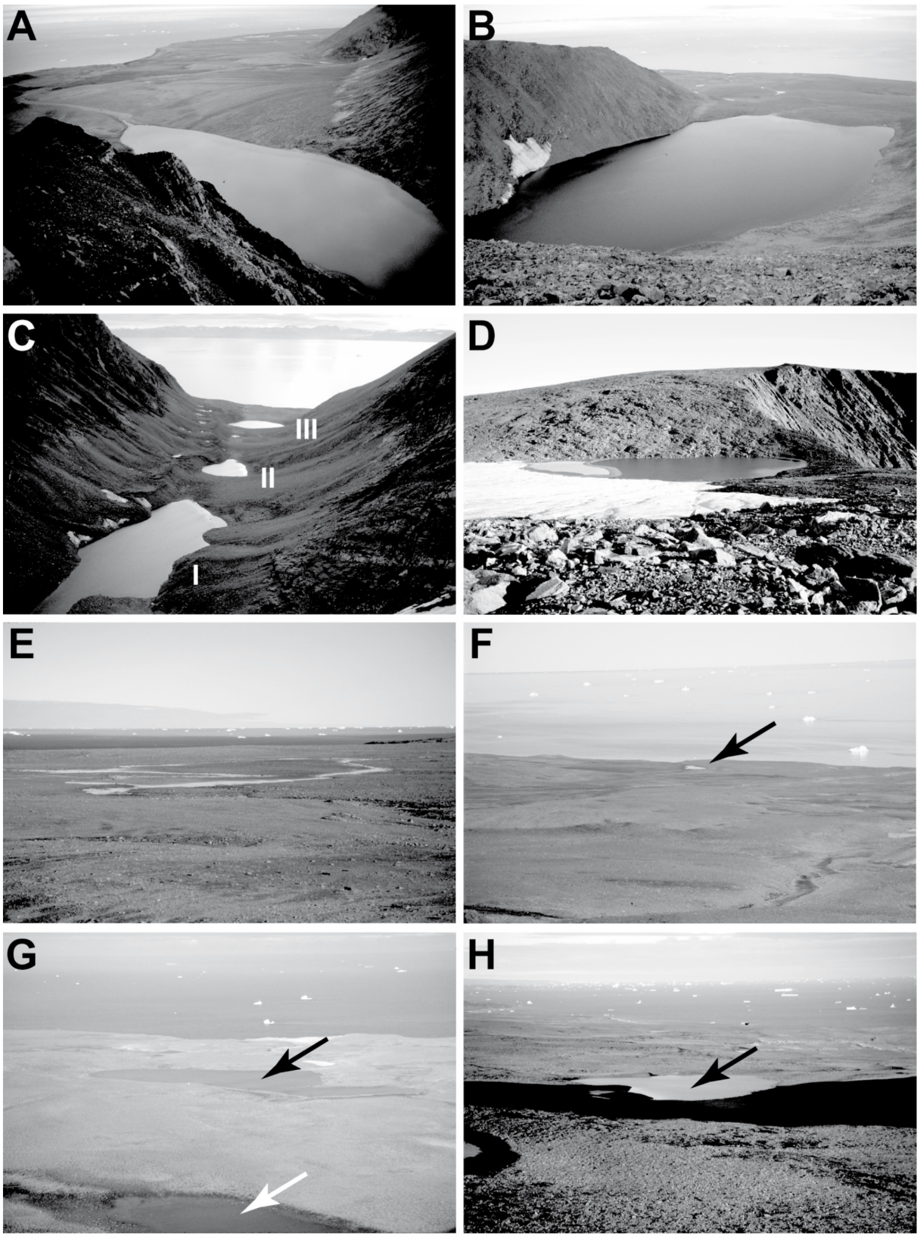

Fig. 5.3: Photographs of lake and pond catchments on Store Koldewey. A. Southeast-ward view onto Melles Lake, the eastern plain and the Greenland Sea. B. East-ward view onto Fox Lake and the Greenland Sea. C. West-ward view onto the Polar Bear Lakes I, II and III, the Dove Bugt and the mainland. D. North-ward view onto Panorama Lake high in the mountain range. E. East-ward view onto Musk Ox Pond and the Greenland Sea. F. East-ward view onto the eastern plain with Raven Pond (arrow). G. East-ward view onto Lemming Lake (foreground, white arrow) and Hjort Lake (background, black arrow). H. East-ward view onto Duck Lake (arrow) and the Greenland Sea. All photos taken by H. Cremer.

Modern climatic conditions in Northeast Greenland are generally characterized by low temperatures and low, mainly solid precipitation. The coastal zones of Greenland are influenced by maritime conditions with fog often covering the lowermost 100 to 200 m a.s.l. At Danmarkshavn weather station (Fig. 5.1) the long-term (1961–1990) mean temperature for the coldest and warmest months of the year is –24.3°C and 3.7°C, respectively, and the mean annual precipitation is 141 mm (Cappelen *et al.*, 2001). All studied lakes and ponds were ice-free during the fieldwork campaign. The vegetation on Store Koldewey is dominated by herbaceous plants but the dwarf-shrubs Arctic Willow (*Salix arctica*), Mountain Aven (*Dryas octopetala*) and Arctic White Heather (*Cassiope tetragona*) are also common. Due to the cold and dry climate, soils are generally poorly developed and are mostly permafrost-affected polar desert soils (gelic gleysols) with a very thin humus layer (Jakobsen, 2001; Bennike *et al.*, 2004).

5.2.2 On Site Measurements and Water Sampling

Hydrological measurements were carried out in nine lakes and two ponds (Table 5.1). In the shallow ponds the measurements were performed on one surface sample at the deepest site location. In Lemming Lake and Panorama Lake measurement sites were located at the shoreline. In the deeper lakes, the bathymetry was recorded with a handheld echosounder from a floating platform, before water sampling was carried out along vertical profiles using a 5 L water sampler (UWITEC Corp., Austria). Following sampling, one litre of this quantity was immediately used to record temperature (WTW Oxi196 probe; WTW GmbH, Weilheim, Germany), oxygen saturation and concentration (WTW Oxi196 probe), conductivity (WTW LF197 probe) and pH (pHScan WP2 tester, Eutech Instruments Ltd., Singapore). In the deeper lakes the transparency was recorded using a Secchi disc.

At selected water depths samples were taken in order to measure anion and cation concentrations. For this purpose two 30 ml samples were filtered through a cellulose acetate filter with a pore size of 0.45 µm and subsequently filled into Nalgene bottles. The water samples for cation analyses were additionally preserved with 0.2 ml nitric acid. All anion and cation samples were stored cool and dark until laboratory analysis. Anion and cation concentrations were measured with an ion chromatograph (DX320, DIONEX Corp., USA) and an emission spectrometer (ICP-OES Optima 3000 XL, Perkin-Elmer Inc., USA), respectively. The hydrogen carbonate content was examined by determining the acid neutralizing capacity of the water samples through HCl titration.

For the analysis of diatom phytoplankton 1.0 to 2.0 litres of lake water was filtered through cellulose acetate filters (pore size: 0.45 µm). The filters were stored in the dark at cool temperatures. In the laboratory the cellulose acetate filters were treated consecutively with 30 % hydrogen peroxide and 65 % nitric acid at 100 °C for the dissolution of filters, removal of organic material and disaggregation of diatom frustules. Following the removal of excess acid and the diminution of the sample quantity in the beaker to 50 ml, slides were prepared employing the sedimentation tray method of Battarbee (1973). Naphrax® was used as a mounting media. Identification and counting of diatoms was carried out using an Olympus BX51 microscope equipped with an UPlanFl 100x/1.30 oil immersion objective with differential interference contrast (DIC). On each slide 200–600 diatom valves were counted in order to calculate diatom phytoplankton concentrations and relative species abundances in the water column. Identification of planktonic diatoms was based on Krammer and Lange-Bertalot (1991) and various specialized and updated literature (e.g., Håkansson, 2002).

5.3 Results and Discussion

5.3.1 Hydrology

Altitude, Bathymetry, and Temperature. - Altitude and maximum water depth of the nine lakes and two ponds are summarized in Table 5.1. Whereas the lakes located on the eastern plain have a moderate maximum water depth between 2.0 and 7.4 m, the lakes on the mountain range are generally deeper and have relatively steep basin topographies (Figs. 5.1, 5.3A–C).

Lake water temperatures generally were low ranging from 7.2 °C to 0.8 °C (Table 5.2). The temperature is influenced by the topography and size of the sites, maximum water depth, solar radiation, cloud cover, and thermal capacity. For example, the Polar Bear Lakes which are located in a narrow west-east oriented valley in the mountain range (Fig. 5.3C), had coldest surface temperatures. These lakes receive less solar radiation than the lakes on the eastern plain (Duck Lake, Hjort Lake, Goose Lake), Fox Lake and Melles Lake. Panorama Lake, located high in the mountain range on 600 m a.s.l., was still partly ice covered in August/September 2003 (Fig. 5.3D) and had the lowest temperature of 0.8 °C. The shallow Raven Pond and Musk Ox Pond, both located on the eastern coast (Figs. 5.1, 5.3E and 5.3F), also had relatively low water temperatures at around 2.5 °C, possibly due to their low thermal capacity and persistently blowing cool winds.

Table 5.1. Geographical location, altitude (Alt.), maximum water depth (Mwd) and Secchi disc transparency (SeDe) of lakes and ponds on Store Koldewey, Northeast Greenland.

Lake	**Latitude** [N]	**Longitude** [W]	**Alt.** [m a.s.l.]	**Mwd** [m]	**SeDe** [m]
Melles Lake	76°07’40’’	18°37’57’’	166	72.0	5.4
Duck Lake	76°25’15’’	18°45’00’’	118	6.4	5.8
Hjort Lake	76°25’59’’	18°45’41’’	122	6.1	
Goose Lake	76°26’36’’	18°48’14’’	115	7.4	4.8
Lemming Lake	76°25’42’’	18°47’44’’	121	*ca.* 2.0	
Fox Lake	76°15’06’’	18°41’32’’	302	57.7	8.1
Raven Pond	76°16’34’’	18°36’18’’	104	0.4	
Polar Bear Lake I	76°14’10’’	18°43’48’’	175	15.9	7.3
Polar Bear Lake III	76°14’01’’	18°46’12’’	129	12.3	3.3
Musk Ox Pond	76°13’57’’	18°35’55’’	115	0.2	
Panorama Lake	76°14’51’’	18°45’34’’	602	*ca.* 5.0	

Vertical hydrological profiles measured in the deeper lakes showed only minor variations in lake-water temperature (Table 5.2) indicating complete, wind-induced thermal mixing of the lakes during the ice-free season. A weak, probably time-limited thermal stratification was observed in Melles Lake (Table 5.2) likely due to a positive impact of a period with relatively calm and sunny weather. A regularly re-occurrence of thermal stratification lasting days or weeks during the summer period would classify Melles Lake and other deeper lakes on Store Koldewey as discontinuous polymictic (Wetzel, 2001a).

Conductivity and Major Ions. - Conductivity in the water bodies on Store Koldewey was extremely low and ranged between 8 and 41 µS cm^{-1} (Table 5.2), which was confirmed by comparably low concentrations of major ions (Table 5.3). Generally, these low values reflect the low degree of soil formation and vegetation cover (Bennike *et al*., 2004) and the bedrock geology on the island consisting mainly of metamorphic gneisses that are relatively resistant to weathering. Lowest conductivity and ion concentrations were measured in Fox Lake confirming its clearness and moderate Secchi depth transparency of 8 m (Table 5.1). The conductivities are among the lowest hitherto measured in arctic and subarctic freshwater environments (e.g., Antoniades *et al*., 2003; Blake *et al*., 1992; Brodersen and Anderson, 2002; Douglas and Smol, 1994; Duff *et al*., 1999; Sorvari *et al*., 2002; Willemse, 2002) and demonstrate that the lakes and ponds on Store Koldewey are mainly fed by ion-depleted meltwaters. Conductivities in lakes on Store Koldewey were in the same range as values reported from other sites in Northeast Greenland, including the shallow lakes (6 to

77 µS cm^{-1}) at Zackenberg Station (Fig. 5.1; Jeppesen *et al*., 2001; Meltofte and Rasch, 1998; Rasch and Caning, 2003).

Table 5.2. Hydrological data along vertical profiles in lakes and ponds on Store Koldewey, Northeast Greenland.

Lake	Water depth	Temperature	Oxygen saturation	Oxygen concentration	Conductivity	pH
	[m]	[°C]	[%]	[mg l^{-1}]	[µS cm^{-1}]	
Melles Lake	1	6.8	102	12.2	38	8.0
	2	6.7	105	11.8	37	7.9
	5	6.8	104	12.5	38	7.9
	10	6.5	107	12.2	37	8.1
	15	6.3	105	12.7	37	8.0
	20	5.9	108	12.5	37	7.9
	25	5.4	107	13.0	37	7.9
	30	5.3	110	13.4	37	7.9
	35	4.4	112	13.8	37	7.9
	40	4.4	109	13.8	37	7.9
	45	4.4	108	13.3	37	7.9
	50	4.4	109	13.7	37	7.9
	55	4.4	111	13.8	37	7.9
	60	4.2	110	13.8	37	7.9
	65	4.1	111	13.8	37	7.9
	68	4.2	109	13.6	37	7.9
	70	4.0	112	14.0	37	7.9
Fox Lake	1	4.8	112	13.6	9.0	8.4
	3	4.8	112	13.9	8.8	8.2
	5	4.7	115	14.2	8.8	8.1
	10	4.7	116	14.5	8.8	8.1
	15	4.7	116	14.2	8.8	8.0
	20	4.7	115	14.5	8.8	7.9
	25	4.7	111	13.7	8.9	7.9
	30	4.7	111	13.9	9.0	7.9
	35	4.7	114	14.1	8.7	7.8
	40	4.7	116	14.4	8.7	7.8
	45	4.7	114	14.1	8.8	7.6
	50	4.7	115	14.3	8.7	7.5
	55	4.7	111	13.7	9.3	7.5
Polar Bear Lake I	1	3.3	125	15.7	35.1	8.5
	3	3.3	125	15.6	31.0	8.4
	5	3.3	125	15.7	30.9	8.4
	10	3.3	124	15.4	30.8	8.4
	12	3.3	125	15.5	30.9	8.3
	15	3.3	121	15.2	30.8	8.2

Table 5.2. (continued)

Lake	Water depth [m]	Temperature [°C]	Oxygen saturation [%]	Oxygen concentration [mg l-1]	Conductivity [µS cm-1]	pH
Polar Bear Lake III	1	3.5	123	15.7	41.5	8.4
	3	3.4	124	15.6	41.4	8.4
	5	3.4	124	15.5	41.3	8.3
	10	3.4	126	15.8	41.3	8.3
	13	3.4	126	16.0	41.3	8.3
Panorama Lake	0.2	0.8	120	15.9	8.0	8.5
Raven Pond	0.2	2.4	126	16.0	21.0	8.2
Musk Ox Pond	0.1	2.6	120	15.9	26.9	8.4
Duck Lake	1	7.1	130	15.4	11.2	9.0
	2	7.2	130	15.5	11.1	9.0
	3	7.1	135	15.7	11.1	8.8
	4	7.2	128	15.4	11.0	8.6
	5	7.2	129	15.1	11.0	8.6
	6	7.2	127	15.2	11.1	8.6
	6.4	6.7	126	14.9	11.9	8.8
Hjort Lake	1	5.3	128	15.4	16.3	9.3
	2	5.3	124	15.3	16.1	9.3
	3	5.3	119	15.4	16.2	9.2
	4	5.3	121	14.6	15.8	9.1
	5	5.4	115	15.3	15.7	9.2
	6	5.3	125	14.9	15.9	9.2
Goose Lake	1	4.9	118	14.5	12.2	9.1
	2	4.5	118	15.3	11.1	9.0
	3	4.4	121	15.2	11.1	9.0
	4	4.5	120	15.0	11.2	9.0
	5	4.4	122	15.1	11.0	9.0
	6	4.5	122	15.2	11.0	8.9
	7	4.4	122	15.3	11.1	9.0
Lemming Lake	0.2	5.8	125	15.3	23.7	8.3

However, the concentration of major ions was extremely low which is atypical even for lakes in the High Arctic. The main recorded cations Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, and Si^{4+} had mean concentrations in the surface water layers (1 m) of 2.23, 0.48, 1.34, 0.34, 0.34 mg l^{-1}, respectively. The concentrations of the major anions, Cl^{-}, SO_4^{2-}, and HCO_3^{-} ranged from 1.53 to 3.92 mg l^{-1} (mean: 2.18 mg l^{-1}), from 0.53 to 3.25 mg l^{-1} (mean: 2.14 mg l^{-1}), and from 1.37 to 13.8 mg l^{-1} (mean: 7.86 mg l^{-1}), respectively (Table 5.3). The observed cation (Ca > Na > Mg > K) and anion (HCO_3 > Cl ≥ SO_4) proportions are typical for lakes and ponds formed on igneous bedrock (Wetzel, 2001a) and reflect the composition of metamorphic gneisses on Store Koldewey. The measured ion concentrations were a magnitude lower than many records from lakes and ponds in West Greenland (Willemse, 2002), Arctic Canada (Antoniades *et al*., 2003; Douglas and Smol, 1994) and Siberia (Duff *et al*., 1999), which likely is the result of the harsher conditions in Northeast Greenland and, hence, lower degree of soil formation, weathering and bioproduction.

The lack of variation in the vertical profiles of conductivity and ion concentration (Tables 5.2 and 5.3) confirms the complete mixing of the lake water column during the ice-free period.

Table 5.3. Cation and anion data for lakes on Store Koldewey, Northeast Greenland.

Lake	**Water depth [m]**	**Cation and anion concentrations[mg l^{-1}]**							
		Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	Si^{4+}	Cl^{-}	SO_4^{2-}	HCO_3^{-}
Melles Lake	1	3.40	0.96	2.64	0.59	0.42	3.92	3.25	13.7
	2	3.35	0.96	2.32	0.55	0.42	4.15	3.44	13.8
	5	3.35	0.95	2.33	0.55	0.42	3.86	3.34	13.1
	10	3.34	0.95	2.31	0.53	0.42	3.82	3.47	13.1
	20	3.32	0.95	2.34	0.56	0.41	3.91	3.35	13.2
	30	3.30	0.94	2.30	0.54	0.41	3.97	3.39	13.1
	40	3.31	0.94	2.31	0.53	0.42	3.77	3.39	13.2
	50	3.32	0.95	2.32	0.54	0.42	3.86	3.31	13.0
	60	3.29	0.93	2.28	0.58	0.42	3.81	3.36	12.8
	70	3.28	0.94	2.32	0.56	0.43	3.86	3.18	13.0
Fox Lake	1	0.73[1]	0.79[1]	1.51[1]	0.58[1]	2.96[1]	1.69	0.53	1.86
	3	0.36	0.17	0.95	< 0.20	0.12	1.79	0.55	1.86
	5	0.36	0.17	0.95	< 0.20	0.12	1.84	0.57	1.74
	10	0.36	0.17	0.96	< 0.20	0.12	1.77	0.53	1.49
	20	0.36	0.17	0.96	< 0.20	0.12	1.77	0.53	1.74
	30	0.39	0.24	—	< 0.20	0.14	1.74	0.61	1.37
	40	0.36	0.17	0.94	< 0.20	0.12	1.73	0.54	1.74
	50	0.38	0.20	0.96	< 0.20	0.27	1.80	0.56	1.74
Polar Bear Lake I	1	3.93	0.45	1.20	0.44	0.39	2.04	3.43	11.5
	5	3.94	0.44	1.20	0.43	0.38	2.03	3.65	11.6
	15	3.55	0.43	1.45	0.59	0.44	2.45	3.75	10.5
Polar Bear Lake III	1	5.41	0.63	1.42	0.54	0.47	2.48	5.48	15.2
	5	5.44	0.63	1.44	0.55	0.47	2.50	5.32	14.8
	12	5.45	0.63	1.43	0.56	0.48	2.45	5.19	15.4
Duck Lake	1	0.66	0.34	0.94	< 0.20	0.11	1.65	0.65	3.32
	3	0.65	0.34	0.92	< 0.20	0.11	1.63	0.69	3.44
	6	0.64	0.33	0.94	< 0.20	0.11	1.68	0.64	3.44
Hjort Lake	1	1.10	0.55	1.18	< 0.20	0.17	1.97	0.69	6.37
	3	1.08	0.54	1.19	< 0.20	0.16	1.90	0.69	6.25
	6	1.10	0.55	1.23	< 0.20	0.16	1.96	0.71	6.25
Goose Lake	1	0.75	0.24	1.08	< 0.20	0.70	1.53	0.97	3.08
	3	0.75	0.25	0.97	< 0.20	0.71	1.49	1.02	2.83
	7	0.78	0.24	0.98	< 0.20	0.75	1.53	0.94	2.96

[1]The high cation concentrations at 1 meter depth in Fox Lake result from the failure of water filtration after sampling.

pH. - The pH of surface lake and pond waters was circumneutral to alkaline with values ranging between 7.5 and 9.3 (Table 5.2). This agrees with other limnological sites in West Greenland (Willemse, 2002), Canada (Douglas and Smol, 1994; Vézina and Vincent, 1997), Alaska (Hobbie, 1984) and Siberia (Duff et al., 1999). On the other hand, at Zackenberg Station (Fig. 5.1) most shallow lakes display a neutral to weakly acidic pH ranging between 5.5 and 7.8 (Meltofte and Rasch, 1998; Rasch and Caning, 2003). The lake-water pH is influenced by the bedrock geology, the presence and composition of soils, and the influx of weakly acidic meltwaters following thaw during spring and summer. The snow in the catchment of Fox Lake, for example, had a pH of 5.6, whereas the water in the main inflow already had an alkaline pH of 8.6, despite a relatively short travel distance, emphasizing the importance of the bedrock geology for the pH of lakes in the soil poor mountain range. The distinctly higher pH in the larger lakes located on the eastern plain on Store Koldewey (Duck Lake, Hjort Lake, and Goose Lake; Table 5.2) was possibly due to a higher biogenic productivity which would have led to a removal of carbon dioxide from the water column, thus shifting the inorganic carbon equilibrium towards more alkaline conditions (Wetzel, 2001a). Like the other hydrological parameters, the vertical distribution of pH did not display any considerable variation within the water column (Table 5.2), which is a typical feature for oligotrophic lakes (Wetzel, 2001a).

Oxygen. - All studied lakes and ponds on Store Koldewey were characterized by high oxygen concentrations and supersaturation of the water with oxygen. Surface water oxygen concentrations ranged from 12.2 to 15.9 mg l^{-1} and oxygen saturations from 102 to 130 % (Table 5.2). These values are in the range of the few published oxygen regimes for arctic lakes and ponds (e.g., Anderson *et al.*, 1999; Howard and Prescott, 1973; Willemse, 2002). The availability of dissolved oxygen in lake water which is essential for most aquatic animals and bacteria is balanced by inputs (atmosphere, photosynthesis) and losses (biological and chemical oxidations) which are a function of temperature, pressure (altitude), salt content, biological activity, and wind (Wetzel, 2001a). The high oxygen levels recorded in the lakes and ponds on Store Koldewey were likely a combined result of low water temperature, permanently blowing wind, reduced salt content, and low biotic respiratoric activity. As algal primary productivity is the only process that can cause oxygen supersaturation, the elevated oxygen concentrations in the studied lakes indicate considerable algal productivity. There was no variation of the oxygen concentration within the water column (Table 5.2). Only Melles Lake displayed a distinct increase of the oxygen concentration with water depth, which most likely was a result of the downwards decreasing water temperature. The thermal warming in the upper layers of Melles Lake during summer and the hence reduced oxygen solubility combined with minor hypolimnetic and benthic activity resulted in the formation of the observed orthograde oxygen depth-profile which is typical for thermally stratified oligotrophic lakes (Wetzel, 2001a).

Nutrients. - Although phosphate, nitrate and chlorophyll concentrations were not determined within this study, the lakes and ponds on Store Koldewey were likely oligotrophic. This is for example indicated by the low concentrations of silica and sulphur (Table 5.3), the uniform vertical oxygen profiles (Table 5.2), and the moderate Secchi disc transparencies. The lake-water concentrations of the main metallic micronutrients (iron, manganese) were generally below 20 µg l^{-1}. Furthermore, comparable high arctic, culturally undisturbed sites in Canada and at Zackenberg Station (Fig. 5.1) also display extremely low chlorophyll, phosphorus, and nitrogen concentrations (Antoniades *et al*., 2003; Rasch and Caning, 2003), being distinctly lower than the mean values given in the general trophic classification of lakes (Wetzel, 2001a).

5.3.2 Diatom Phytoplankton

Species Composition. - Apart from the documentation of the basic hydrological characteristics in the lakes on Store Koldewey, the diatom flora in the water column was surveyed. A total of eight planktonic diatom species were identified (see Cremer and Wagner, in press, for a detailed taxonomical documentation). Table 5.4 presents a species list and Table 5.5 documents the inter-lake distribution and abundance of planktonic diatom taxa in the water column. The diatom communities were generally dominated by one or two species only, usually *Cyclotella pseudostelligera* HUSTEDT and/or *C. rossii* HÅKANSSON and/or *Fragilaria tenera* (W. SMITH) LANGE-BERTALOT (Table 5.5). Low phytoplankton diversity and the dominance of small diatom species is a typical feature of arctic and subarctic lakes (e.g., Hobbie, 1984; Cremer and Wagner, 2003). However, only few data on diatom phytoplankton from arctic lakes have been published and in most cases rely on surface sediment surveys. For example, diatom surface sediment assemblages in several lakes from Finnish Lapland were dominated by small *Cyclotella* and *Aulacoseira* species including *C. rossii* (Sorvari *et al*., 2002). Many lake sediment assemblages in the Canadian and Siberian Arctic were also co-dominated by small *Cyclotella* species apart from the usually dominant small benthic fragilarioid species (see Michelutti *et al*., 2001; Perren *et al*., 2003 and Rühland *et al*., 2003, and references therein). *Cyclotella*- and *Fragilaria*-dominated diatom sediment assemblages from Greenland were earlier described by Cremer *et al.* (2001a, 2001b) in long sediment cores recovered in Raffles Sø and Basaltsø (Fig. 5.1). Diatom phytoplankton, however, was not observed in three of the studied lakes, among them the shallow Panorama Lake and Lemming Lake and the deep Fox Lake (Table 5.5). In particular, the lack of diatom phytoplankton in 57 m deep Fox Lake was surprising and might have been triggered by the extremely low ionic concentrations and conductivity (≤ 9 µS cm^{-1}), which were the lowest among all studied lakes (Tables 5.2 and 5.3). These values are possibly lower than the minimum values required by planktonic diatoms. Moreover, the extremely low silica concentration in Fox Lake (< 0.20 mg l^{-1}, Table 5.3) might be advantageous for chrysophytes and non-siliceous algae (Wetzel, 2001a). Algal communities in arctic lakes dominated by classes other than diatoms were described, for example, from Zackenberg Station (chrysophytes; Rasch and Caning, 2003) and from Meretta Lake, northern Canada (chrysophytes and cryptophytes; Douglas and Smol, 2000).

Table 5.4. Planktonic diatom taxa identified in water samples from lakes on Store Koldewey, Northeast Greenland. 1Single findings. The morphology and morphometric data of the taxa are documented in Cremer and Wagner (in press).

Diatom taxon
Aulacoseira islandica (O. MÜLLER) SIMONSEN[1]
Aulacoseira tethera HAWORTH
Aulacoseira sp. 1[1]
Cyclotella antiqua W. SMITH
Cyclotella pseudostelligera HUSTEDT
Cyclotella rossii HÅKANSSON
Fragilaria tenera (W. SMITH) LANGE-BERTALOT
Stephanodiscus minutulus (KÜTZING) CLEVE et MÖLLER[1]

Diatom Valve Concentration. - The concentration of diatoms in the water column clearly fluctuated in and among the lakes (Table 5.5). Highest valve numbers were recorded in Melles Lake (956,000 valves l^{-1}) and Hjort Lake (1,282,000 valves l^{-1}) indicating that the valve concentration apparently was unrelated to water depth. There is also no positive correlation between the diatom valve concentration and the hydrological characteristics given in Tables 5.2 and 5.3. The vertical distribution of diatom valves in the water column also displayed clear within-lake and inter-lake variations. Melles Lake, Polar Bear Lake I, and Hjort Lake had a valve maximum in the subsurface layers of the water column (Table 5.5), which is a typical phenomenon in oligotrophic lakes with large transparency and most probably reflects the harmfulness of intensive solar radiation (UV) in the uppermost layers of such aquatic environments (Wetzel, 2001a). In the other lakes valve numbers either increase (Polar Bear Lake III, Goose Lake) or decrease (Duck Lake) with water depth indicating that other factors (sinking, grazing, mixing) might have been also an important trigger of the phytoplankton concentration at various water depths.

Table 5.5. Diatom phytoplankton characteristics in lakes on Store Koldewey, Northeast Greenland. Abbreviations: Dp, Diatom phytoplankton; Cc, chrysophyte cysts; Wd, water depth; DVC, Diatom valve concentration; Frust, frustules; P, present; NP, not present; Atet, *Aulacoseira tethera*; Cpse, *Cyclotella pseudostelligera*; Cros, *Cyclotella rossii*; Ften, *Fragilaria tenera*; oPlank, other planktonic diatoms; Benth, benthic diatoms.

Lake	Dp	Cc	Wd	DVC	Relative abundance of diatom taxa [%]					
			[m]	[valves·1000 l^{-1}]	Atet	Cpse	Cros	Ften	oPlank	Benth
Melles Lake	P	P	2	761.8	0.0	76.2	17.2	0.2	0.0	6.4
	P	P	5	956.6	0.0	73.1	20.4	0.4	0.0	6.1
	P	P	10	873.2	0.0	76.8	16.6	1.5	0.0	5.1
	P	P	20	905.6	0.0	68.5	22.0	0.8	0.3	8.4
	P	P	30	584.3	0.0	75.7	17.7	1.1	0.0	5.5
	P	P	40	473.9	0.0	80.1	14.7	2.4	0.0	2.8
	P	P	50	518.4	0.0	67.6	13.0	1.2	0.0	18.2
	P	P	60	440.7	0.0	79.0	14.1	1.0	0.0	5.9
Fox Lake	NP	P	1-40	no planktonic diatoms present						few
Polar Bear Lake I	P	P	1	198.9	0.0	38.5	2.2	45.9	0.0	13.4
	P	P	5	230.7	0.0	37.6	2.1	50.0	0.4	9.9
	P	P	10	136.5	0.0	41.2	1.6	43.3	0.0	13.9
Polar Bear Lake III	P	P	1	53.8	0.0	2.6	0.0	84.5	0.6	12.3
	P	P	5	81.7	0.0	5.6	1.0	83.1	2.1	8.2
	P	P	10	162.7	0.0	7.7	0.4	80.7	0.9	10.3
Panorama Lake	NP	P	0.4	no planktonic diatoms present						few
Duck Lake	P	P	1	245.0	3.8	6.4	2.1	0.0	0.4	87.3
	P	P	3	71.9	0.5	0.0	1.1	1.1	0.0	97.2
	P	P	5	113.6	3.2	4.6	2.8	0.5	0.0	88.9
Hjort Lake	P	P	1	880.4	3.5	3.7	0.2	63.6	0.4	28.6
	P	P	3	1282.8	0.8	0.0	0.0	69.9	0.2	29.1
	P	P	5	837.2	4.1	0.0	0.0	68.5	0.2	27.2
Goose Lake	P	P	1	607.4	8.9	0.6	0.0	5.5	0.0	85.0
	P	P	3	433.4	13.1	1.3	0.0	7.4	0.0	78.2
	P	P	5	589.1	8.1	0.7	0.7	8.5	0.0	82.0
Lemming Lake	NP	P	0.3	no planktonic diatoms present						few

However, the observed pattern of diatom distribution and the real algal standing stock in the lakes must be confirmed in future studies by measurements of the chlorophyll concentration and by in-situ examination of other algae classes, which was not possible within the framework of this study. The few high arctic lakes for which diatom concentration data are available include, for example, Meretta Lake, a culturally affected lake in Arctic Canada. The diatom cell concentrations in Meretta Lake in the mid 1990's ranged from 121 to 1600 cells per litre (Douglas and Smol, 2000), which is a relatively small number compared to the lakes on Store Koldewey. These numbers represented between 0.7 and 18.9 % of the total phytoplankton community which was dominated by cryptophytes, chrysophytes, and dinoflagellates (Douglas and Smol, 2000). Because the examination of other phytoplankton groups was not included in this study, the proportion of diatoms of the total phytoplankton community has to remain undetermined.

The phytoplanktonic diatom communities in the relatively shallow Duck Lake, Goose Lake and, to a minor degree, Hjort Lake were dominated by primarily benthic diatoms (Table 5.5). The proportion of benthic diatom taxa in these lakes varied between 27.2 and 97.2 %. Compared to the deeper lakes of the rocky mountain range, the lake bottom and water column of these three lakes were stronger affected by wind-induced mixing and sediment resuspension. This and also the observation that the catchments and shorelines of these lakes had a denser vegetation cover and hence a higher substrate availability for benthic diatom taxa, resulted in enhanced presence of benthic diatoms in the water column.

5.4 Conclusions

Based on the hydrological results, the lakes and ponds on Store Koldewey can be characterized as cold monomictic, thermally unstratified, alkaline, likely oligotrophic water bodies with low diatom phytoplankton diversity. Most lakes on Store Koldewey were ice-free in summer 2003 except for one studied high altitude lake, Panorama Lake, which remained partly ice-covered. Water column mixing likely is continuous resulting in negligible vertical variation of hydrological characteristics. Weak thermal stratification might occur in some lakes (e.g., Melles Lake) for the duration of a few days which then would classify those lakes as discontinuous polymictic lakes. The continuously high oxygen concentration throughout the water column in all studied lakes indicates both relatively high botanical productivity and only limited decomposition activities.

Diatom phytoplankton communities were present in all but one lake and were dominated by small *Cyclotella* and/or elongated *Fragilaria* species. The deep Fox Lake entirely lacked a diatom phytoplankton community during the ice-free season in 2003, presumably as a result of its extremely low ion concentration which hindered planktonic diatoms to compete efficaciously with chrysophytes and non-siliceous algae. However, a sound limnological and ecological interpretation of the here presented data has to remain incomplete at this stage. The study of more than the here reported limnological and floral characteristics (e.g., the study of nutrients, chlorophyll, other algae classes) would allow to describe the state of these high arctic lakes in more detail.

Lakes and ponds in the High Arctic still belong to the least investigated freshwater ecosystems on earth although they offer the unique opportunity to study the limnology of freshwater ecosystems under extreme climatic conditions and in regions that are not or hardly directly affected by human activity. Particularly, in high arctic Greenland there is a lack of limnological studies. This study is a first mosaic stone to fill this gap.

6 First indication of Storegga tsunami deposits from East Greenland *

6.1 Introduction

The most prominent and largest submarine landslide during the Holocene in the North Atlantic, the so-called second Storegga Slide, occurred ca. 8100 cal. yr B.P. offshore western Norway and displaced between 2400 and 3200 km^3 sediment (Haflidason *et al.*, 2005; Fig. 6.1). The slide generated a series of major tsunami waves, which reached the adjacent coastal regions within minutes to hours (Dawson *et al.*, 1988; Bondevik *et al.*, 1997a, 2005a). When the tsunami waves reached shallow waters near the coast, wave heights increased dramatically, and large volumes of sediments in the shallower water of the near-coastal regions became eroded and redeposited. In addition, onshore low elevation areas were inundated, and in some funnel shaped fjords the runup exceeded 10 m (Bondevik *et al.*, 2003, 2005a). Deposits related to the Storegga tsunami have been recorded from coastal lakes, raised marine sediments, and peat sequences from eastern Scotland (Dawson *et al.*, 1988; Long *et al.*, 1989; Dawson and Smith, 2000), western Norway (Bondevik *et al.*, 1997b; Bondevik, 2003; Bondevik *et al.*, 2005a), the Faroe Islands (Grauert *et al.*, 2001), and the Shetland Islands (Bondevik *et al.*, 2005b). In Iceland and Greenland, however, such deposits have not been detected hitherto (cf. Noa Sø; Fig. 6.1; Wagner and Melles, 2002, pers. comm. Antony Long), although modelling has shown that a 3 m high wave reached the coast of eastern Greenland only 2.5–3 hours after the release of the slide (Bondevik *et al.*, 2005a).

An unnamed lake, here referred to as Loon Lake (N 72°53.2', W022°08.3'), on outer Geographical Society Ø, East Greenland, was visited during an expedition in the summer 2003 (Bennike *et al.*, 2004; Fig. 6.1). Loon Lake is today located at 18 m a.s.l. in a low land area that drains into Kejser Franz Joseph Fjord and the Greenland Sea to the north. According to relative sea-level curves (Funder, 1989; Wagner and Melles, 2002) and unpublished biostratigraphical work of the region, the basin was inundated by the sea after the last deglaciation in the early Holocene (Bennike and Björck, 2002; Fig. 6.1). The maximum water depth of Loon Lake today is 11.7 m in its centre, where a 10.25 m long sediment sequence was recovered using gravity and piston corers (both from UWITEC Co., Austria).

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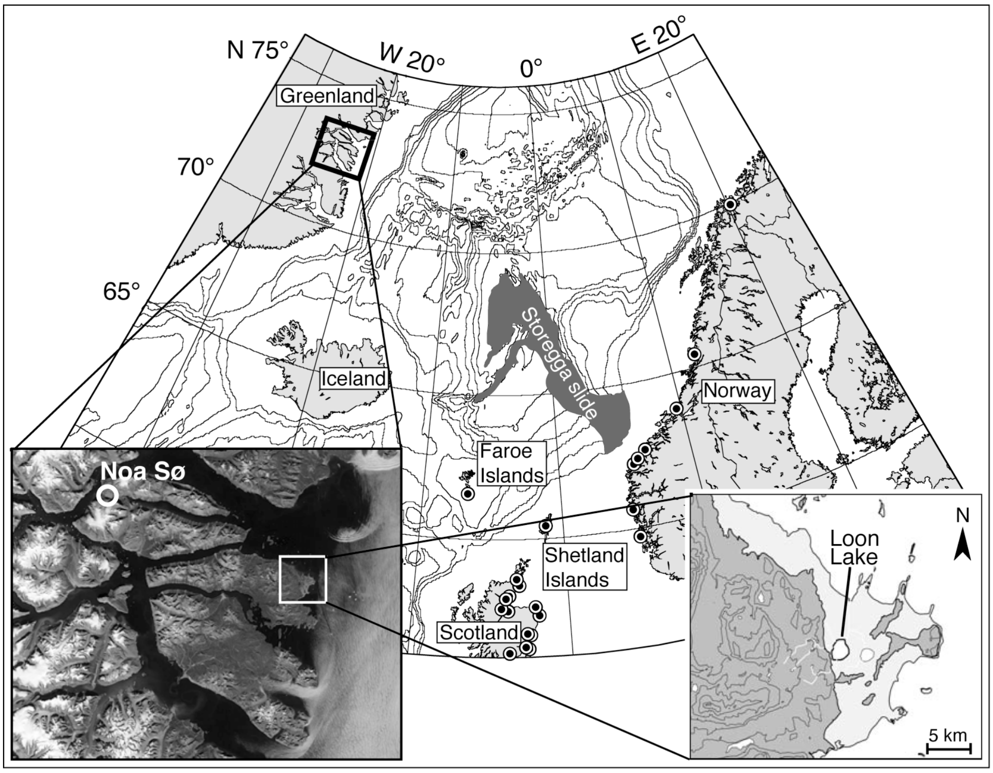

Fig. 6.1: Map of the North Atlantic region showing the extent of the Storegga slide and locations where deposits of the Storegga tsunami have been recorded (references are mentioned in the text). The insert to the left shows a satellite photography with the location of Noa Sø mentioned in the text and the vicinity of Loon Lake on outer Geographical Society Ø, East Greenland. The insert to the right shows the present coastline (light grey) of outer Geographical Society Ø and the presumed coastline at ca. 8000 cal. yr BP (dark grey), which is based on relative sea-level reconstructions (Funder, 1989; Wagner and Melles, 2002) at about 30m above the present sea level.

A compact diamicton at the basis of the sequence prevented further penetration into the sediments. Relatively homogenous, fine grained sediments form the major part of the sediments above the diamicton. However, one of the piston core segments, covering the sediment depth between 2.51 and 5.24 m, contains a 0.72 m thick horizon that differs markedly in facies from those above and below. A range of geophysical, granulometric, biogeochemical, biological, and chronological analyses were carried out in order to evaluate the genesis of this unique horizon.

6.2 Laboratory work

Prior to core splitting, magnetic susceptibility and Gamma Ray density were measured in one-centimetre intervals with a Multi Sensor Core Logger (MSCL, Geotek Co.; Weber *et al.*, 1997). Following core opening, description, and photographic documentation, one core half was continuously subsampled in 2 cm intervals. Subsamples were split into two sets of aliquots, one of which was kept cool and dark for macrofossil analyses, whilst the other set was freeze-dried and the water content determined by the loss of weight.

Biogeochemical and grain-size analyses were conducted in intervals of 2 to 16 cm of the freeze-dried aliquots. For the biogeochemical measurements sediment was ground to < 63 µm and homogenized. Total carbon (TC) content was measured on an Elementar III (VARIO Co.) analyser. Total organic carbon (TOC) content was measured on a Metalyt CS 1000S (ELTRA Co.) analyser, after treating the sediment with 10% HCl at 80°C in order to remove carbonates. Total inorganic carbon (TIC) was calculated from the difference between TC and TOC. Grain-size analyses in the range < 1250 µm were conducted with a laser particle analyser (Fritsch Co.), after organic matter had been removed with 3% H_2O_2.

Diatom analyses, which were carried out according to Cremer *et al*. (2001a), reveal an absence of diatoms throughout the sequence investigated. For enumeration of macrofossils each subsample was wet sieved on 0.1, 0.2, and 0.4 mm sieves and the residue left on the sieves analysed using a dissecting microscope.

Radiocarbon dating of shells or shell fragments of marine bivalves (Table 6.1) was conducted by accelerator mass spectrometry (AMS) at the Leibniz Laboratory for Radiometric Dating and Isotope Research in Kiel, Germany. The obtained ^{14}C ages were calibrated into calendar years (cal. yr B.P.) after subtracting 550 yr (delta R = 141; Rasmussen and Rahbek, 1996) for the local marine reservoir effect and using the CALIB 5.0 programme and the MARINE04 data set (Stuiver and Reimer, 1993; Hughen *et al*., 2004).

Table 6.1. Radiocarbon dates of shells and shell fragments from core segment Lz1116-7 from Loon Lake.

Sample number	Corr. depth (cm)	Weight (mg)	Material	C (mg)	$\partial\,^{13}C$ (%)	^{14}C Age (^{14}C yr BP)	Corr. age (corr. yr BP)	Cal. age (cal. yr BP)
KIA-24754[a]	259–260	40.0	shell, not identified	1.1	0.41	7285±40	6735±40	7535–7700
KIA-27660	387–388	44.4	*Portlandia arctica*	1.4	-0.77	8270±45	7720±45	8465–8760
KIA-27661	391–392	20.7	*Portlandia arctica*	1.4	0.85	8475±45	7925±45	8725–9025
KIA-27662	443–454	31.3	shell, not identified	1.1	1.68	8190±45	7640±45	8395–8605
KIA-27663[a]	482	9.2	*Portlandia arctica*	1.1	1.54	8065±45	7515±45	8300–8500
KIA-27664	481–482	8.0	*Portlandia arctica*	0.9	0.37	8370±45	7820±45	8595–8935
KIA-27665[a]	491–492	100.8	*Astarte borealis*	1.5	0.56	8105±50	7555±50	8330–8540
KIA-24755[a]	509–510	16.0	shell, not identified	0.9	-5.95	8175±60	7625±60	8370–8630

[a] Samples are believed to be not re-deposited.

6.3 Stratigraphy, interpretation and conclusion

Visual core description, and physical properties, grain size, and biogeochemical analyses reveal that two facies can be distinguished in the core sequence studied.

Facies 1 occurs below 4.77 m and above 4.05 m. It is dominated by silt and clay, with a grain-size median below 12 µm, a water content of 28 to 35%, a magnetic susceptibility of less than 70 10^{-5} SI, a GRD of 1.55 to 1.80 g/ccm, and TIC and TOC contents of 0.3 to 0.7 and 0.5 to 0.7%, respectively (Fig. 6.2). The radiocarbon ages obtained from facies 1 indicate that the sediment sequence was deposited between 8630 cal. yr B.P. (sample KIA24755 at the basis) and 7535 cal. yr B.P. (sample KIA24754 at the top; Fig. 6.2, Table 6.1). According to samples KIA24755 (8370-8630 cal. yr B.P.), KIA27665 (8330-8540 cal. yr B.P.), and KIA27663 (8300-8500 cal. yr B.P.), and neglecting sample KIA27664 (8595-8935 cal. yr B.P.), which is supposed to be redeposited, the sedimentation rate at the basis of the sequence can be calculated by the means of the radiocarbon ages to ca. 3 mm/yr. The lack of lamination and the widespread occurrence of marine fossils throughout this facies indicate that the sediments are bioturbated and were deposited in a marine setting. This confirms relative sea-level curves of the region (Funder, 1989; Wagner and Melles, 2002), which suggest that the Loon Lake basin was inundated by the sea at the time of deposition of facies 1. Taking the modern altitude of Loon Lake at 18 m a.s.l., the relative sea-level curve for outer Greenland (Wagner and Melles, 2002), the depth of the sequence studied at around 4 m below the present sediment surface, and unpublished biostratigraphical work, the local sea level at this time was about 30 ± 10 m higher than today, making Loon Lake a 15–35 m deep marine basin, which was somewhat sheltered from the North Atlantic towards the east by a few islands (Fig. 6.1).

Facies 2 is found between 4.77 and 4.05 m and is the only such in the complete core sequence after deglaciation. It is separated from the underlying facies 1 by an erosive basis and is characterized by several layers of graded sand and silt, which partly indicate cross-bed layering (Fig. 6.3A). Lower water contents of 15 to 25%, higher but varying magnetic susceptibility and of more than 60 10^{-5} SI, and significantly higher GRD decreasing upward from 2.1 to 1.8 g/ccm prevail. TIC and TOC contents are in general lower, particularly at the basis of this facies, but also show distinct fluctuations (Fig. 6.2). Marine fossils occur, but are less common and less well preserved than in facies 1. The highest concentration of *Sphacelaria* (marine brown algae) correlates with low magnetic susceptibility, small grain-size and high TIC and TOC values. Sample KIA27663 (8300–8500 cal. yr B.P.) from 5 cm below facies 2 provides a maximum age for the deposition of this facies (Fig. 6.2; Table 6.1). The ages of the *Portlandia arctica* remains (samples KIA27660 and KIA27661) from shortly

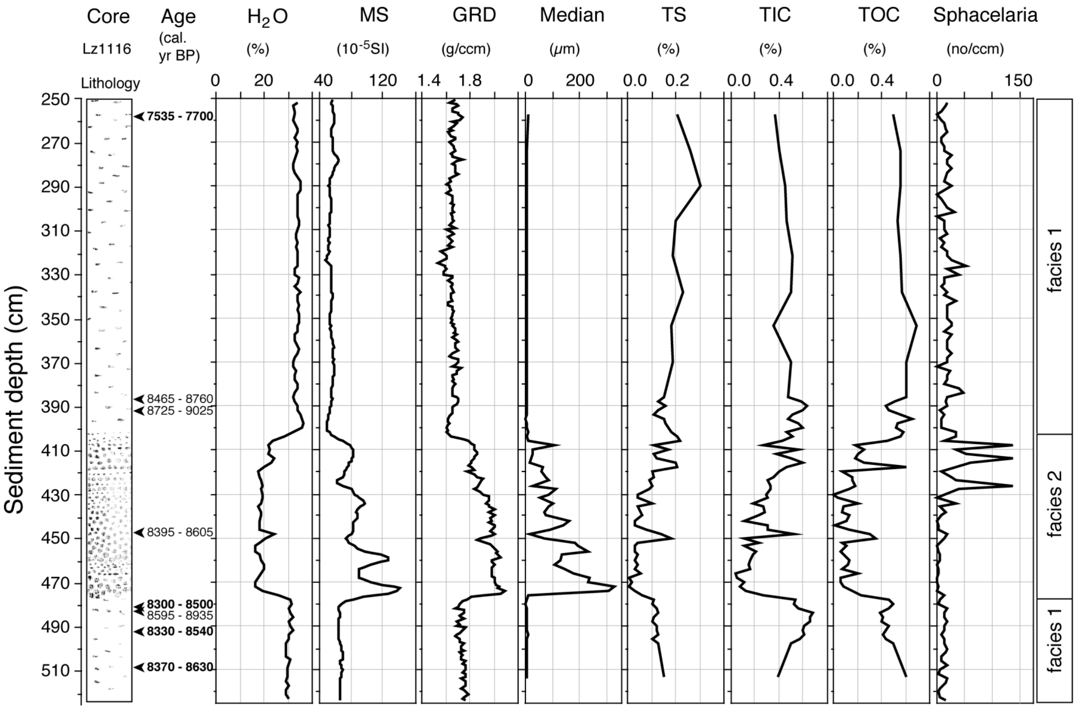

Fig.6.2: Lithology, calibrated radiocarbon dates, water content, magnetic susceptibility, gamma-ray density, grain-size median, total sulphur, total inorganic carbon, total organic carbon and Sphacelaria sp. of core Lz1116 between 251 and 524cm sediment depth. Radiocarbon ages are given as mean values and uncertainties (Table 1). Calibrated ages in bold are presumed to be not re-deposited (Table 1).

above facies 2 are reversals and, thus, supposed to be erroneous and not applicable for the exact determination of the duration of the deposition of facies 2. However, assuming that the sedimentation rate was stable during deposition of facies 1 and amounted to ca. 3 mm/yr, the upper boundary of facies 2 can be extrapolated from the age of sample KIA24754 (7535–7700 cal. yr B.P.) to ca. 8010–8175 cal. yr B.P. The inferred high sedimentation rate, along with the sedimentological characteristics and particularly the erosive and sharp lower boundary (Fig. 6.3), imply that facies 2 was deposited during a single, short-term event. Such an event could be a turbidity current or another mass movement process. The enclosure of redeposited material is indicated by the radiocarbon age of shell remains from within facies 2 (sample KIA27665), which is slightly older (8330–8540 cal. yr B.P.) than those from the underlying facies 1. However, the repeated fluctuations in grain size, GRD, magnetic susceptibility, TIC and TOC contents, and macrofossil remains argue against a single turbidite, and a series of turbidites without intervening pelagic sediments is considered unlikely.

One alternative explanation for the formation of the sand horizon could be that the 8.2 kyr cooling event led to distinct changes in sediment supply. However, it is unlikely that such a process created a series of mass movement processes as indicated in the single graded

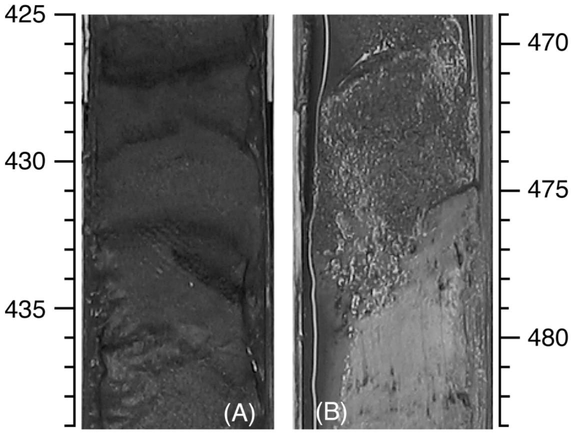

Fig. 6.3: Photographs of the erosive basis of the presumed Storegga tsunami deposits in core Lz1116. (A) is from 425 to 439 cm sediment depth and shows several layers with an upward-fining grain size from sand to silt. (B) shows the erosive basis at 475 cm sediment depth.

sand beds. Other explanations, such as a major storm event or deposition as a consequence of isostatic uplift (cf., Grauert *et al.*, 2001) can also widely be discarded, because the sandy horizon is the only such in the complete Holocene sequence and marine macrofossils in the sediments above document ongoing marine conditions.

The most likely explanation for the sandy horizon is a tsunami event. This would explain all characteristics of facies 2, including the erosive basis, the occurrence of several beds of upward fining sand layers, and the occurrence of interspersed marine fossils. In the Storegga tsunami deposits in basins in western Norway the deposition of repeated fining upward beds has been attributed to a series of waves that inundated these shallow marine basins or coastal lakes during the tsunami event (Bondevik *et al.*, 1997b). Although facies 2 from Loon Lake is generally finer grained than Storegga tsunami deposits elsewhere, the radiocarbon ages obtained from marine shells below, within, and above this facies suggest that the sandy horizon has also been formed by the Storegga tsunami. The comparatively finer grain sizes are probably due to the loss of transport energy, when the waves were partially blocked by the emerged areas towards the east of Loon Lake basin (Fig. 6.1). It should be mentioned that most of the dated samples from Loon Lake were performed on *Portlandia arctica* shells (Table 6.1). This is a deposit feeding species, and dating of shells of this species can give dates that are some hundred years older than ages obtained on other marine molluscs (Forman and Polyak, 1997). No *Portlandia arctica* shells were used to establish the 550 years marine reservoir effect in eastern Greenland. The use of *Portlandia arctica* shells in this study may, in addition to the unknown thickness of eroded sediments and/or shifts in the reservoir effect during the early Holocene, explain why the date for the basis of facies 2 is slightly older (8300–8500 cal. yr B.P.; sample KIA27663) than expected for Storegga tsunami deposits (around 8100 cal. yr B.P.).

As shown by the grain-size composition, physical properties, biogeochemical, and macrofossil data at least four waves inundated Loon Lake basin. Similar evidence for several repeated Storegga waves have been recorded elsewhere (Bondevik *et al.*, 1997b). The presence of sea ice during the tsunami, which would have produced a very different tsunami signature in the Loon Lake basin, can apparently be excluded. Dawson and Smith (2000) suggested that the tsunami likely took place during autumn. This would be confirmed by our results, taken that the early Holocene was part of the climatic optimum in East Greenland (e.g., Wagner *et al.*, 2000) and winter sea ice formed late in the autumn. Deposition of reworked marine fossils on top of the tsunami sequence, such as indicated in samples KIA27660 and KIA27661, is a common phenomenon and has been explained by a period with deposition of older organic-rich matter (e.g., Bondevik *et al.*, 1997a, b; Grauert *et al.*, 2001).

Astonishingly, deposits of the Storegga tsunami have not been detected in other lakes from East Greenland so far. Partly, this may be due to the fact that the sediment records from these lakes were not extensively studied for such deposits (pers. comm. Antony Long). The absence of the Storegga tsunami deposits in Noa Sø in the inner fjord region of East Greenland (Fig. 6.1; Wagner and Melles, 2002) can probably be explained by its relatively sheltered location at the very end of the Dusens Fjord.

6.4 Conclusions

We conclude that the sandy horizon in the sediment sequence from Loon Lake most likely represents the first indication of Storegga tsunami deposits in eastern Greenland. The reconstruction of runup heights, as done in other investigations, is not possible at this stage, because Loon Lake was the only lake sampled. Nevertheless, the discovery of Storegga tsunami deposits in East Greenland demonstrates the large impact of this event in the North Atlantic region.

7 Lake sediments from Store Koldewey, North-East Greenland, as archive of late Pleistocene and Holocene climatic and environmental changes *

7.1 Introduction

Reconstruction of the glacial history of the coastal region in North and East Greenland is based on numerous studies of onshore and offshore archives. During the Last Glacial Maximum (LGM), North Greenland is suggested to have been completely covered by the Greenland Ice Sheet and a local ice cap north of it (Zreda *et al.* 1999; Bennike 2002; Bennike & Björck 2002; Bennike 2004). Exposure dating using cosmogenic ^{36}Cl and radiocarbon dating of raised marine deposits indicate that the outer coastal areas became ice-free between 10.2 and 7.9 thousand calibrated radiocarbon years BP (cal. kyr BP) in western North Greenland and between 11.2 and 9.3 cal. kyr BP in central North Greenland (Zreda *et al.* 1999; Bennike 2002; Bennike & Björck 2002; Bennike 2004). Models of the ice extent in North-East and East Greenland and studies of off-shore regions suggest that the Greenland Ice Sheet reached onto the shelf during the LGM, but was mostly restricted to transverse channels and fjords (Nam *et al.* 1995; Funder & Hansen 1996; Elverhøi *et al.* 1998; Funder *et al.* 1998; Bennike & Weidick 2001; Evans *et al.* 2002). It is commonly accepted that some uplands at the outer coast remained ice-free (Mangerud & Funder 1994; Funder *et al.* 1998; Denton *et al.* 2005) and it has also been suggested that some low elevation areas remained ice-free (Hjort 1979, 1981; Adrielsson & Alexanderson 2005). For example, Hjort (1981) suggested that eastern Store Koldewey remained ice-free after the Kap Mackenzie stadial, which was originally thought to be of early Weichselian age but it was re-dated to the Saalian or older by Hjort & Björck (1984).

Climatic and environmental reconstructions for the coastal region between North and East Greenland are partly hampered by the scarcity of suitable archives. In North Greenland, in particular, detailed information about climatic or environmental evolution is lacking. Most information comes from raised marine and littoral deposits, from lacustrine sediments, and from glacial landforms (Fredskild 1995; Funder & Hansen 1996; Hjort 1997; Bennike 2002, 2004). Pollen and macrofossil studies of a sediment sequence from a lake on Amdrup Land, North Greenland, indicate an early Holocene thermal maximum (Fredskild 1995).

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However, due to sediment slumping, this lake does not contain a continuous Holocene record. The climatic and environmental history of North-East and East Greenland after the last deglaciation has been reconstructed in more detail. Ice-core studies and modelling of past temperatures reveal a distinct temperature increase at 11.7-11.5 kyr BP (Dansgaard *et al.* 1989; Johnsen *et al.* 1992b; Dansgaard *et al.* 1993; Dahl-Jensen *et al.* 1998; Andersen *et al.* 2004). The thermal maximum occurs in the early to mid Holocene until *c.* 6 kyr BP, when temperatures gradually decreased to values below present during the Little Ice Age. This Holocene temperature pattern is also recorded in deltaic sedimentary records and several lake sediment sequences (Wagner *et al.* 2000; Christiansen *et al.* 2002; Wagner & Melles 2002; Wagner *et al.* 2005). Most of the former studies on lake sediments used pollen or biogeochemical proxies (Funder 1978, 1979; Björck *et al.* 1994a, b; Fredskild 1995; Wagner *et al.* 2000; Wagner & Melles 2001, 2002). Chironomid larvae are the most abundant insects found in fresh water lakes (Cranston 1995) and preserve well as fossils in lake sediments. Chironomid assemblages are known to respond sensitively to factors such as salinity, pH, lake depth, summer temperatures, dissolved organic carbon, and oxygen concentrations (Walker & MacDonald 1995; Walker 2001; Brodersen & Anderson 2002; Porinchu & MacDonald 2003). Nevertheless, chironomid larvae have only rarely been used to reconstruct Holocene palaeoenvironmental changes in Greenland. A pioneering study from central East Greenland suggests that changes in fossil chironomid assemblages can be used to reconstruct palaeoenvironmental changes (Wagner *et al.* 2005). Other studies have documented significant changes in the concentration of head capsules of chironomid larvae during the Holocene, although chironomid remains have not been identified to a higher taxonomic level (Bennike & Funder 1997; Bennike & Weidick 2001; Bennike *et al.* 2002). Palaeoenvironmental investigations in the ice-free part of North-East Greenland form an important link between investigations of ice cores from the Inland Ice and marine sediment cores from the Greenland Sea. From the ice-free areas, information can be obtained on the past long-term history of the ice sheet, and climatic changes after the last deglaciation. A reconstruction of the ice-sheet evolution during the late Quaternary is a precondition for a thorough understanding of the stability of the Greenland ice-sheet during future climate changes.

In this study, a 2.9 m long sedimentary record from a small lake on Store Koldewey, North-East Greenland, was investigated for its physical and biogeochemical properties and, for the first time in such a high arctic region in Greenland, for its fossil chironomid assemblages. The aims of this project are to provide information about i) the possible existence of local ice-free regions during the LGM, ii) the local climatic and environmental evolution during Late Pleistocene and Holocene times, iii) the usefulness of chironomid larvae in North-East Greenland as a tool for palaeoenvironmental reconstructions and iv) the climatic and environmental characteristics in North-East Greenland compared to those of other coastal regions of Greenland.

7.2 Study site

Store Koldewey is a north to south orientated island located in North-East Greenland between 76°45' and 75°55' N and 19°10' and 18°27' W (Fig. 7.1). The island is *c.* 90 km long and up to 11 km wide. It is separated from the Greenland mainland by the Dove Bugt in the west. A 3 to 5 km wide mountain range with an elevation of up to 900 m a.s.l. dominates the island along the north-south axis. The mountains are flat-topped, with steep slopes down to the shoreline on the western part of the island. The eastern part of the island is characterized by a 2 to 5 km wide plain at *c.* 150 m a.s.l. (Fig. 7.1). Several steep-sided, east-west orientated, U-shaped valleys cut through the mountain range.

Store Koldewey is part of the East Greenland Caledonian Fold Belt. The bedrock of the mountain range is Precambrian metamorphic rock, mainly gneiss. Middle to late Mesozoic and supposedly Pliocene marine sediments occur on the eastern plain (Escher & Pulvertaft 1995; Bennike *et al.* 2004). Quaternary sediments, such as till and raised marine and littoral deposits, are widespread (Bennike *et al.* 2004). Some of the U-shaped valleys and some of the depressions on the eastern plain are filled with lakes and ponds. One of the lakes, unofficially called Duck Lake, is located at 118 m a.s.l. on the eastern plain. This is well above the marine limit, which is around 50 m a.s.l. (Hjort 1979). Duck Lake is *c.* 3.2 ha in size and has a catchment area of about 30 ha. The lake is mainly fed by melt water from surface runoff in the catchment area and drains towards the northeast into the Greenland Sea.

The surroundings of the lake are characterized by sparse vegetation, with the dwarf shrubs *Salix arctica*, *Dryas octopetala*, *Cassiope tetragona*, grasses, sedges, mosses and lichens (Bennike *et al.* 2004). Various breeding and migratory birds and land mammals, such as arctic fox, arctic hare and lemming, represent the local vertebrate fauna. The fairly low faunal diversity is related to the low summer temperatures and the low degree of soil evolution, which is reflected in the occurrence of a very thin humus layer (Bennike *et al.* 2004; Cremer *et al.* 2005). The mean monthly temperatures recorded at Danmarkshavn weather station a few kilometres north of Store Koldewey range between -23°C in February and 4°C in July. The mean annual precipitation amounts to 141 mm (Cappelen *et al.* 2001).

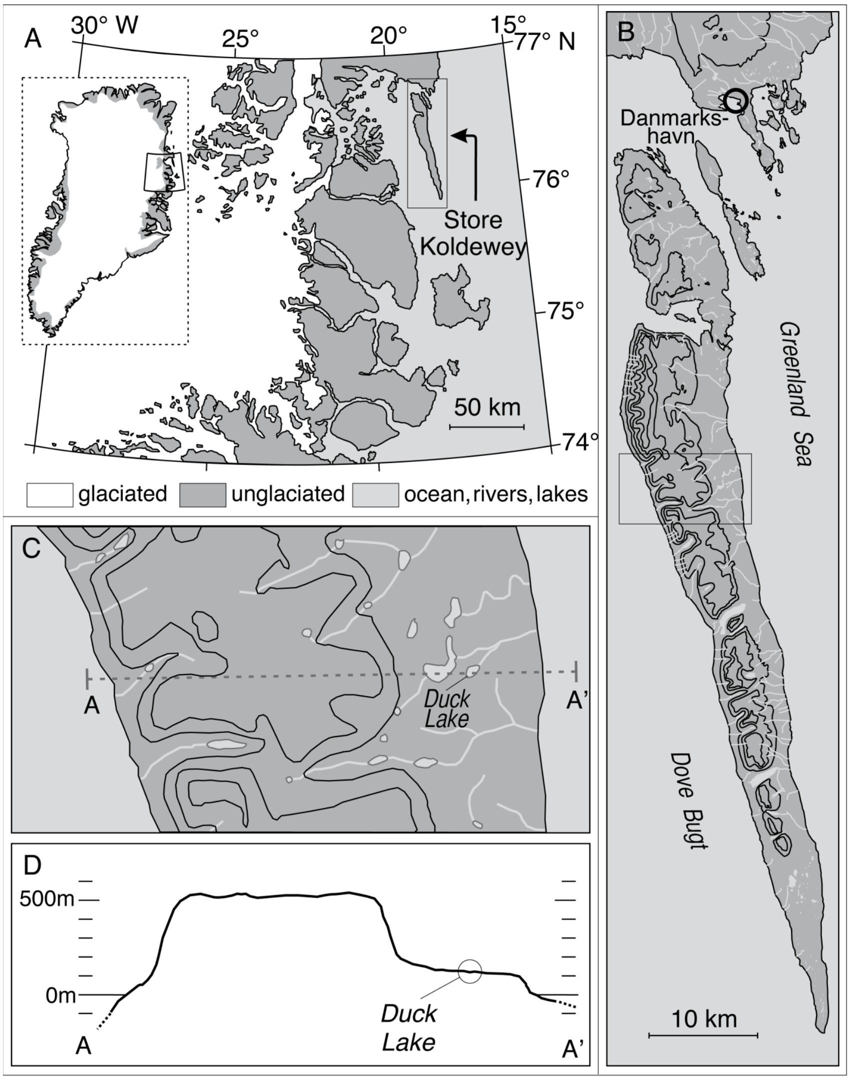

Fig. 7.1: A) Modern ice margin and ice-free regions of the Northeast Greenland Fjord Zone with Store Koldewey. B) Map of the Store Koldewey region with 200 m contour lines and place names mentioned in the text. C) Detailed map of the surrounding of Duck Lake located on the plateau east of the mountain chain. D) Vertical east–west profile (A–A'), vertical exaggeration (VE) is about 4.

7.3 Material and methods

7.3.1 Fieldwork

The fieldwork was carried out in the August 2003. Bathymetrical measurements along crossing profiles revealed a simple basin morphology with a maximum water depth of 6.4 m in the centre of the lake (Fig. 7.2).

The Secchi disc transparency of Duck Lake was 5.8 m (Bennike *et al.* 2004; Cremer *et al.* 2005). Water sampling was performed along a vertical profile in the centre of the lake using a 5 l water sampler (UWITEC Co., Austria) in order to measure oxygen content and oxygen saturation, temperature, pH (WTW Oxi196 probe), and specific conductivity (WTW LF 197; both WTW Corp., Germany).

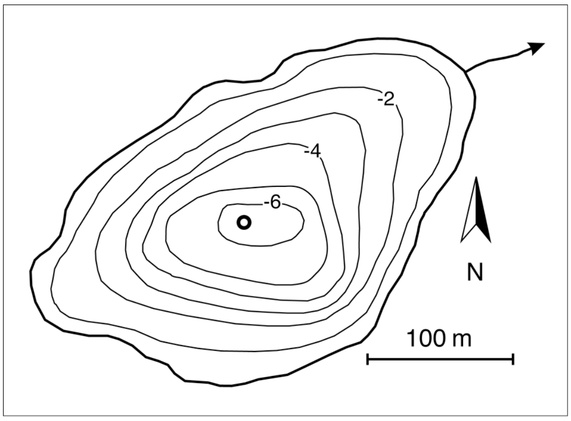

Fig. 7.2: Bathymetry of Duck Lake with 1 m contour lines based on echo-sounder measurements along four crossing profiles. The coring position is shown by an open circle and the outlet with an arrow.

Sediment cores were recovered from a floating platform. A gravity corer (UWITEC Corp., Austria) was used to retrieve undisturbed surface sediments (core Lz1103-2, cf. Fig.7.4). A piston corer (UWITEC Corp., Austria) was used to recover longer and deeper sediment sequences (cores Lz1103-5/6, cf. Fig. 7.4). The sediment cores were cut into segments up to 1 m in length in the field and stored at a temperature of 4°C until further processing.

7.3.2 Laboratory work

Magnetic susceptibility (MS) and wet bulk density (WBD) were logged in one-centimetre intervals using the MultiSensorCoreLogger MSCL 14 (GEOTEK Corp., Germany). The sensors and the processing routine have been described by Dittmers & Niessen (2002) and by Weber *et al*. (1997), respectively. After splitting the cores lengthwise, the individual core sequences were correlated by matching distinct horizons and by the results of the MS and WBD measurements. One half of the complete sequence was contiguously sectioned into 0.5 cm intervals (0 - 152 cm) or into 2 cm intervals (200 - 290 cm). The water content was calculated from the mass difference between wet and freeze-dried samples. Subsequently, the samples were split into aliquots in order to perform all analyses on the same sample horizons.

For grain-size analysis about 100 mg of sediment of each sample was wet sieved to separate the fraction >1 mm. The fraction <1 mm was treated with H_2O_2 to remove organic matter and with 1 ml sodium polyphosphate to disaggregate the sample. After a 5 s ultrasonic bath, grain-size analysis was performed with a laser diffraction particle size analyser (Cilas 1180L, SUCCESS SCIENTIFIC Corp., Taiwan). Mean and sorting were calculated according to Folk & Ward (1957).

For biogeochemical measurements, about 100 mg sediment of each sample was ground for 30 s at 1500 rpm in zirconium oxide grinding jars by the mixing mill MM200 (RETSCH

Corp., Germany). Total carbon (TC), total nitrogen (TN) and total sulphur (TS) content were determined using a VARIO EL III analyser (ELEMENTAR Corp., Germany). Total organic carbon (TOC) was quantified using a METALYT CS1000S analyser (ELTRA Corp., Germany) after removal of carbonate with 10% HCl at 80°C. Total inorganic carbon (TIC) was calculated from the difference between TC and TOC. TOC/TN and TOC/TS ratios were calculated using TOC, TN and TS mass percentages respectively.

The other half of the sediment cores was used for analyses of head capsules of chironomid larvae. 2 g of wet sediment were taken at 4 cm intervals throughout cores Lz1103-2 and Lz1103-5. For sorting of chironomid remains, the samples were deflocculated for 12 min in 10% KOH at a temperature of 100°C. The sediment was then sieved using a 90 µm sieve and a maximum of 80 head capsules was picked out from the residue left on the sieve using a dissecting microscope (25x magnification) and a Bogorov sorting tray. Finally, the head capsules were successively dehydrated in 80% and 100% ethanol and mounted in Euparal®. All chironomid head capsules in the residue were counted in order to determine the total chironomid concentration.

Head capsules were mainly identified with reference to Hofmann (1971), Saether (1975), Cranston (1982), Oliver & Roussel (1983), Wiederholm (1983), and Brooks *et al.* (2007). *Micropsectra* species were separated into *Micropsectra insignilobus*-type and *Micropsectra radialis*-type, with *M. radialis*-type being characterised by a shorter antennal pedestal, a more roundish ventral postoccipital margin and a strongly reduced postoccipital plate (Heiri & Lotter 2003; Heiri *et al.* 2004). *Corynocera oliveri*-type and *Tanytarsus lugens*-type are difficult to differentiate, in particular in their young instars. *Corynocera oliveri*-type has a mandible with a large surface tooth obscuring three inner teeth and a lateral mentum consisting of darkish compressed teeth. Specimens identified as *Tanytarsus lugens*-type had a mandible with two outer teeth, three inner teeth and a small surface tooth, and a lateral mentum consisting of clearly separated teeth. *Sergentia coracina*-type is differentiated from other *Sergentia* species by less strongly developed striae on the ventromental plates. *Heterotrissocladius* species were separated in *H. maeaeri*-type and a morphotype resembling *H. oliveri* and *H. subpilosus* as described in Saether (1975). However, the median mentum teeth were distinctly longer in this morphotype than in *H. subpilosus*-type, as described in Brooks *et al.* (2007) and we therefore refer to this taxon as *Heterotrissocladius* type A. *H. maeaeri*-type has a larger and more distinct gap between the median teeth and the first lateral mentum tooth than *H. subpilosus*-type or *Heterotrissocladius* type A. Two *Orthocladius* morphotypes were found, i.e. *O. consobrinus*-type, and *Orthocladius* type I (Brooks *et al.* 2007). The mediantooth in *Orthocladius* type I is distinctly longer than the *O. consobrinus*-type. The Shannon-Wiener index H (S) and

Evenness (E) were used as a measure of the diversity of the chironomid assemblages (Murray 1991).

Radiocarbon dating was conducted on one bulk sediment sample, five samples of aquatic mosses, and two samples of terrestrial plant remains. Measurements were performed at the Radiocarbon Dating Laboratory of the GeoBiosphere Science Centre at the University of Lund. Dating of remains of aquatic mosses and terrestrial plants from one level revealed that the aquatic mosses were 520 years older than the terrestrial plants. Since the terrestrial plant remains are regarded to represent the more reliable ages, 520 years were subtracted from all moss dates. All finite dates were converted into calibrated radiocarbon years before present (cal. yr BP) using the program Calib version 5.0.1 (Stuiver & Reimer 1993), and based on the calibration data set IntCal 04.14c (Reimer *et al.* 2004). Means were calculated from the lowest and highest dates at the calibrated 2σ probability distribution with 95% confidence interval (Table 7.2). The age-depth correlation is based on a polynomial function with the best fit to the calibrated dates. $\delta^{13}C$ values were measured by conventional mass spectrometry or in AMS, which, however, due to possible higher uncertainties from the AMS, are not comparable.

7.4 Results and discussion

7.4.1 Modern limnology

A detailed description of the modern hydrology of lakes on Store Koldewey is published by Cremer *et al*. (2005). Duck Lake was completely mixed at the time of sampling (Fig. 7.3). The relatively high oxygen content and oxygen saturation throughout the water column indicate low biotic respiratory activity, which is typical for oligotrophic to ultra-oligotrophic lakes in arctic regions. The extremely low conductivity and the fairly high pH value in Duck Lake are a result of sparse vegetation, little soil formation, gneissic bedrock and the Quaternary sediments in the catchment (Bennike *et al.* 2004; Cremer *et al.* 2005) (Table 7.1, Fig. 7.3).

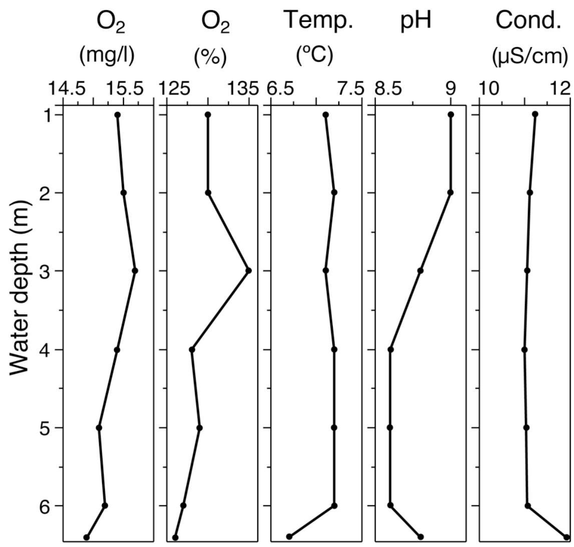

Fig. 7.3: Hydrological profile of Duck Lake in the summer 2003 at the coring position.

Table 7.1. Hydrological parameters of the coring position in Duck Lake in the summer 2003.

Water depth (m)	O_2 content (mg/l)	O_2 saturation (%)	Temperature (°C)	pH	Conductivity (µS/cm)
1	15.4	130	7.1	9.0	11.24
2	15.5	130	7.2	9.0	11.11
3	15.7	135	7.1	8.8	11.06
4	15.4	128	7.2	8.6	11.00
5	15.1	129	7.2	8.6	11.03
6	15.2	127	7.2	8.6	11.06
6.4	14.9	126	6.7	8.8	11.92

7.4.2 Stratigraphy

According to the correlation of gravity and piston cores, the total length of the sedimentary sequence of Duck Lake is 290 cm. This includes a gap between 152 and 200 cm (Fig. 7.4), which was not recovered as a result of coring process. Based on sediment description as well as whole-core and sedimentological analyses, three major stratigraphic units can be distinguished.

Unit A. - Unit A, from 290 to 268 cm, consists of stiff, grey clay and silt. Towards the top of the unit the grain size increases slightly and the sorting decreases. The sediment is homogeneous with sporadically faint laminations. Minor MS changes are seen, with a slight increase towards the top. The water content is low and matches the low content of organic matter and a relatively high WBD of around 2.3 g/cm^3 (Fig. 7.4). Chironomids are absent.

The almost massive structure and the fine grain size suggest high sedimentation rates under relatively calm conditions. The slight coarsening of the grain size towards the top indicates gradually increasing transport energy. The low content of organic matter suggests a low productive environment, as found in proglacial lakes. The stiff consistency and the low water content may have been caused by post-sedimentary compaction.

Unit B. - Unit B comprises the section between 268 and 123 cm. The sediments are grey to dark grey. At the base of the unit there is a layer of gravel with sub-angular pebbles that measure up to 60 mm in diameter. Above the gravel, laminated horizons of clay, silt, sand and fine gravel occur. The grain size decreases between 264 and 140 cm, and increases again towards the top of the unit (Fig. 7.4). Alternating layers of granules and sand, most with a thickness of less than 2 cm, occur between 245 and 208 cm. These varying grain sizes are reflected by strong variations in mean and sorting. Overall low water content, high MS and WBD, with significant changes are most likely the result of coarser grain sizes. The predominance of minerogenic sediment particles is reflected by very low TOC and TN values. Slightly higher content of TS at the top of this unit might be attributed to changes in

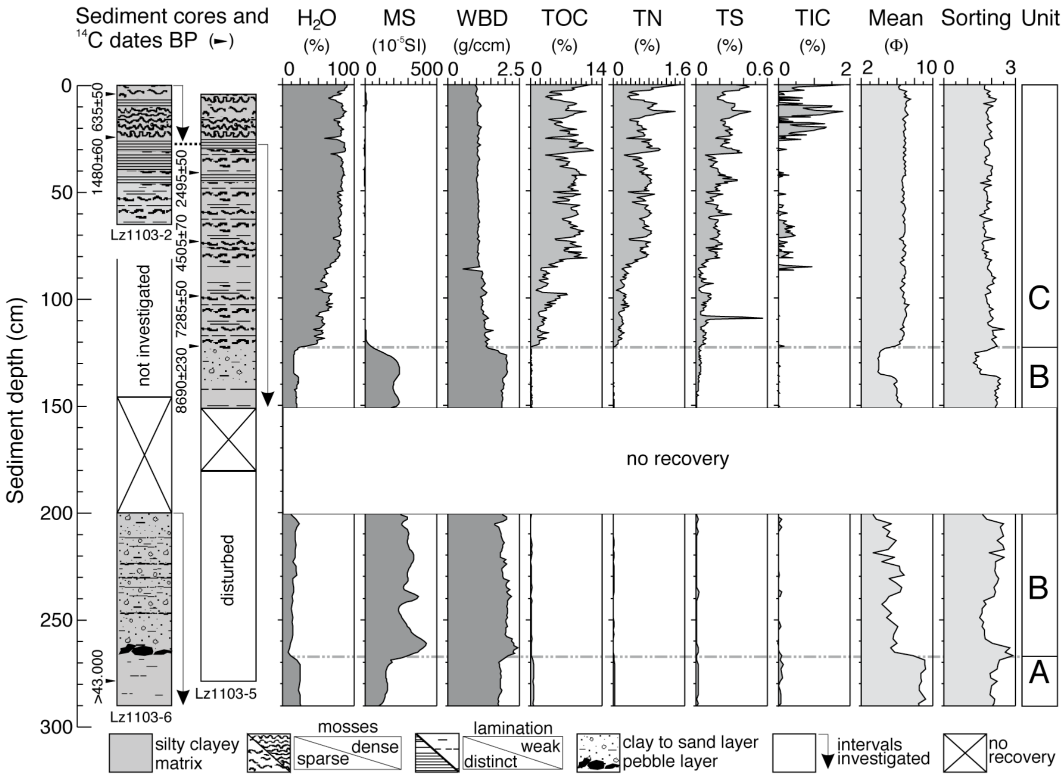

Fig. 7.4: Lithology and sedimentary structures of investigated cores Lz1103-2/5/6 versus sediment depth. Water content (H_2O), total organic carbon (TOC), total nitrogen (TN), total sulphur (TS), mean and sorting are measured at 0.5 cm intervals; magnetic susceptibility (MS) and wet bulk density (WBD) are measured on whole cores at 1 cm intervals; horizontal arrows indicate levels of ^{14}C dated horizons (cf. Table 7.2). Units A, B and C refer to intervals of similar sediment properties; between 152 and 200 cm, no sediment was recovered.

sediment source or to post-depositional migration of sulphur from the overlaying unit. Chironomids are absent in this unit.

The gravel layer at the base of unit B was probably deposited close to the ice margin just after the last deglaciation. The alternating layers of clay, silt, sand and granules were probably deposited in a glaciolacustrine environment, with the coarser grained layers deposited by turbidity currents or other types of density currents. Low temperatures due to ice proximity, reduced light availability owing to high suspension load during the melting season, and the lack of nutrients impeded biogenic production, which is reflected in low organic matter.

Unit C. - Unit C reaches from 123 cm to the top of the sequence and consists mainly of silt and clay of dark grey to olive grey colour. The consistency differs remarkably from unit A and B with a distinct change to soft sediment. Mean and sorting remain on constant levels throughout this unit and reflect silt- to clay-dominated inorganic sedimentation with single coarse sand grains. The abrupt increase of water content at the bottom of the unit, corresponding with a decrease of MS and WBD (Fig. 7.4), is in line with the change from almost pure minerogenic to significant biogenic sedimentation. This is also evident from higher TOC and TN contents. Layers of aquatic moss occur throughout the unit, but are less frequent between 100 and 80 cm. Horizons with little or no moss are characterized by laminations, with laminae thickness reaching up to 1 cm. The aquatic moss is presumably Warnstorfia exannulata, which is by far the most common moss at relatively deep water in lakes in NE Greenland.

The onset of biogenic production and the dominance of fine grained sediments in unit C is a clear indication of the establishment of lacustrine conditions in the Duck Lake basin. Slowly increasing TOC and TN values indicate enhanced biogenic production, probably in combination with increased decomposition. Frequent and pronounced fluctuations of TOC and TN in the upper 80 cm can be due to the irregular occurrence of moss layers, probably in combination with distinct changes of biogenic production due to climatic variations. The content of TS does not match the organic matter represented by TOC and TN. Hence, the TS peak at 110 cm depth is likely formed by minerogenic particles and could originate from a shift towards anoxic conditions in the bottom waters. The first distinct accumulation of TIC occurs at a depth of 85 cm. This peak could be due to carbonate precipitation as consequence of increased biogenic production, such as indicated in TOC and TN maxima, or due to the formation of siderite under strongly reducing conditions and enhanced organic matter decomposition, such as found in another lake in East Greenland (Wagner *et al.* 2000).

7.4.3 Chironomid succession

Chironomid head capsules are absent in unit A and B. In unit C, the concentration of head capsules in the sediment is relatively low, particularly in the upper 30 cm of the record, where the samples contained less than 50 chironomid head capsules. A total of 18 chironomid taxa were identified.

The first head capsules appear at a depth of 120 cm. *Micropsectra insignilobus*-type and *M. radialis*-type are amongst the first taxa to colonize the lake and they are found throughout the record in varying concentration. They are accompanied by *Corynocera oliveri*-type, *Pseudodiamesa*, and cf. *Oliveridia* (Fig. 7.5). These taxa and, with some delay, *O. consobrinus*-type dominate the chironomid assemblage between 120 and 90 cm. *Tanytarsus lugens*-type, and *Heterotrissocladius* type A appear sporadically.

At a depth of about 90 cm the chironomid assemblages change significantly. *O. consobrinus*-type, *Micropsectra insignilobus*-type, and *M. radialis*-type are still dominating, but *Pseudodiamesa*, and cf. *Oliveridia* disappear. They are replaced by *Paracladius*, *Corynoneura scutellata*-type and *Procladius*. *Sergentia coracina*-type and *Heterotrissocladius* type A occur at 80 cm depth. *C. oliveri*-type disappears between 70 and 55 cm depth. Other taxa such as *Psectrocladius sordidellus*-type, *Eukiefferiella* and *Diamesa* are represented by a few head capsules.

Another distinct shift in chironomid assemblages can be observed at about 30 cm depth (Fig. 7.5). *O. consobrinus*-type is still dominating, together with *M. insignilobus*-type and *M. radialis*-type. The latter has a slightly higher percentage value than in the underlying sediments. *Sergentia coracina*-type, *Heterotrissocladius maeaeri*-type and *Paracladius* disappear completely and are replaced by *Pseudodiamesa*.

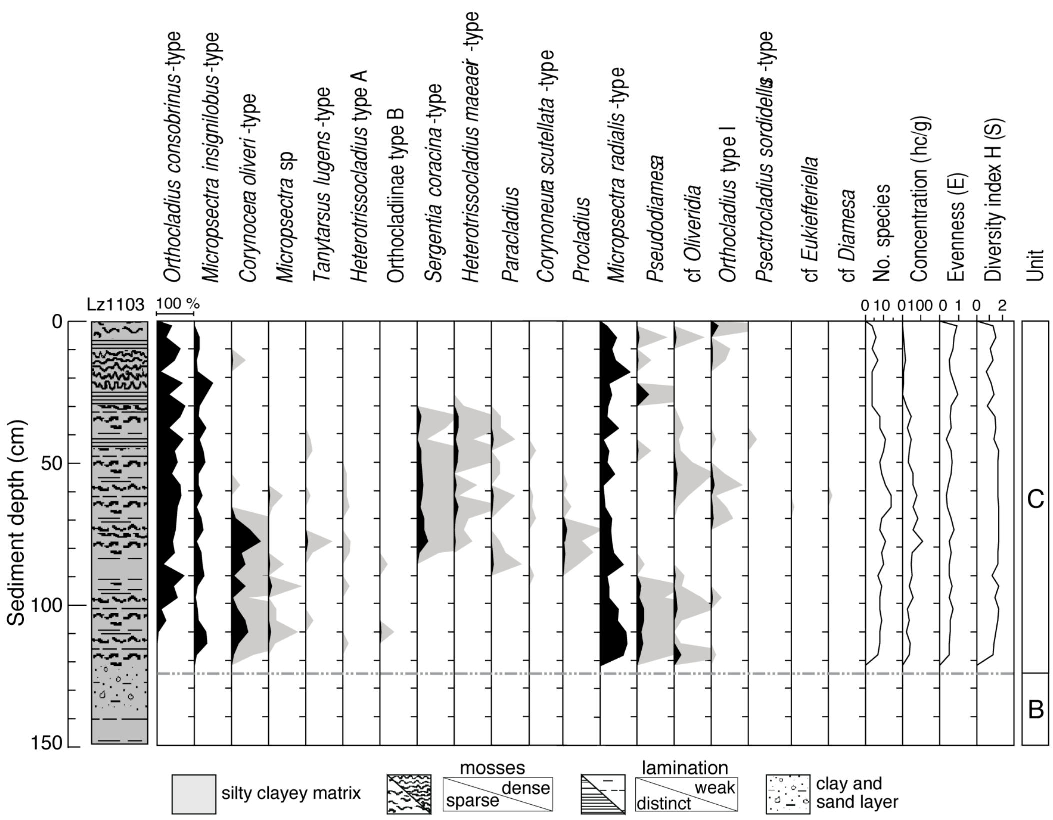

Fig. 7.5: Chironomid percentage diagram from unit C of core Lz1103-2/5 from Duck Lake. Grey colours denote an exaggeration of 10.

The diversity of the chironomid assemblages in the record from Duck Lake, as indicated by the Evenness and H (S) indices, remains relatively constant throughout the record. However, major shifts can be observed in the number of taxa and in the concentration of chironomid head capsules (Fig. 7.5). Values are relatively low in the lower part, higher between *c.* 80 and 50 cm and lower in the upper 30 cm of the record. With a total of 18 identified chironomid taxa, the record corresponds well with the number of taxa reported from other high arctic lake ecosystems, for example on Svalbard (Brooks & Birks 2004) and central East Greenland (Wagner *et al.* 2005). However, considering the harsher environmental conditions on Store Koldewey compared to that in central East Greenland, the taxonomic richness in the Duck Lake record is surprising.

7.4.4 Chronology

One bulk sediment sample was dated (LuS-6526), from 272-274 cm sediment depth (Table 7.2), 4 cm below the upper boundary of unit A. This sample yielded an infinite age and could not be used for chronological constraints. The unit B sediments are inappropriate for ^{14}C dating because of their extremely low contents of TOC and lack of macrofossils. Macrofossils, mostly aquatic mosses but also two unidentified terrestrial plant remains, occur in unit C, allowing us to establish a chronology for this unit by ^{14}C dating.

Table 7.2. Material for and results of AMS 14C-dating. Aquatic mosses were corrected by 520 yr according to parallel dating of aquatic moss and terrestrial plant remain at 25-25.5 cm depth, $\delta^{13}C$ values in italics are from AMS.

Sample number	Depth (cm)	Weight C_{org} (mg)	Material	Radiocarbon age (BP)	$\delta^{13}C$ (‰)	Age corrected	Calibrated age (yr BP)
LuS-6717	3.5-5.5	*c.* 15	aquatic moss	635 ± 50	-25.4	115	278 - 8
LuS-6522	25-25.5	*c.* 1	aquatic moss	2000 ± 60	*-26.2*	1480	1517 - 1294
LuS-6661	25-25.5	*c.* 0.5	terrestrial plant	1480 ± 60	*-31.5*		1517 - 1294
LuS-6716	42-43	>20	aquatic moss	2495 ± 50	-24.8	1975	2054 - 1820
LuS-6524	73.25	*c.* 0.9	terrestrial plant	4505 ± 70	*-27.3*		5320 - 4958
LuS-6715	100-101	*c.* 20	aquatic moss	7285 ± 50	-26.4	6765	7686 - 7564
LuS-6525	121-121.5	*c.* 0.6	aquatic moss	8690 ± 230	*-35.4*	8170	9547 - 8517
LuS-6526	272-274	>50	bulk	>43.000	-22.4	/	/

The 520 yr older ^{14}C age of aquatic moss in comparison with that of terrestrial plant remains at a depth of 25.25 cm (Table 7.2) could be due to redeposition of the moss induced, for instance, by lake level changes or wind. Another possible reason for the difference could be that the CO_2 in the lake water was not in equilibrium with that of the atmosphere, due to input of old carbon from soils surrounding the lake, ^{14}C-depleted meltwater from glacier remnants in the catchment, or the existence of a long-lasting lake ice cover (Björck & Wohlfarth 2001; Wolfe *et al.* 2004). The latter is a common phenomenon in arctic lakes and may also have affected Duck Lake. Any detailed reconstruction of changes in the reservoir effect in the past would require parallel dating of aquatic and terrestrial organic remains from several horizons, which was not possible in Duck Lake. Therefore, a constant reservoir correction of 520 yr was applied to all age determinations of aquatic mosses. This is considered the best possible estimate. Since the reservoir effect may have varied through time, the chronology of the Duck Lake sediment record has to be regarded as tentative.

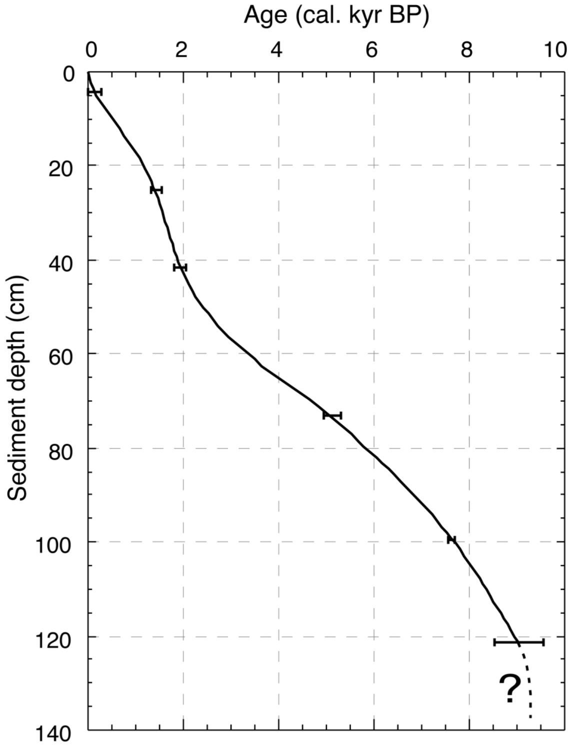

Fig. 7.6: Possible polynomial age–depth relationship of the Duck Lake record; for correction of aquatic mosses, see Table 7.2. Ages were calibrated with Calib version 5.0.1 and the calibration data set IntCal04.14c; error bars of ages account for 2 SD.

Aquatic mosses at a depth of 121.25 cm were dated to 9031 ± 515 cal. yr BP (sample LuS-6526), i.e. the minimum age of onset of biogenic production in the lake. Since mosses from a depth of 4.25 cm indicate an age of 115 ± 50 yr BP (sample LuS-6717) after correction of 520 yr and hence are likely not affected by ^{14}C originating from nuclear bomb testing, the sediment surface is assumed to represent the present. All dated levels show increasing ages with increasing depth (Fig. 7.6).

Unit C, with a thickness of *c.* 123 cm, therefore covers at least the past 9.1 kyr. According to the age-depth model, sedimentation rates are high from 9 to 6 cal. kyr BP and after 3 cal. kyr BP, and lower from 6 to 3 cal. kyr BP (Fig. 7.6). The changes in the sedimentation rates are probably related to the occurrence of the moss layers in the early and late Holocene sediments, since the mosses are relatively voluminous compared with remains of algae and clastic sediments.

7.5 Climatic and environmental history

7.5.1 Pre-Holocene and earliest Holocene (>9.2 cal. kyr BP)

The only pre-Holocene age in the sediment sequence from Duck Lake derives from unit A, 4 cm below the boundary to unit B (Fig. 7.4) and has an infinite age (>43 kyr BP). However, the organic content in unit A is extremely low, and it may also be that the sediments contain some reworked, 'old' organic material. Pre-Holocene sediments are common in North-East Greenland (Hjort 1979, 1981; Hjort & Björck 1984; Houmark-Nielsen *et al.* 1994; Bennike *et al.* 2002). On eastern Store Koldewey, Pliocene marine shell-bearing deposits occur at elevations between 110 and 130 m a.s.l. (Bennike *et al.* 2004) and are similar to unit A sediments from Duck Lake with respect to colour, grain-size composition and elevation above sea level. However, no shells or shell fragments were identified in unit A, and we consider it more likely that the unit A sediments represent a late Pleistocene glacio-lacustrine deposit. The stiff consistency of unit A could possibly be due to compaction by glacial overriding.

The wide range of grain sizes of Unit B, particularly the layer of pebbles at the base, the alternating laminated and homogenous layers with low water content and high MS values (Fig. 7.4) can be attributed to glaciolacustrine sedimentation. A mineralogical differentiation of the provenance of the sediments in unit B was not possible, so it has to remain open whether the glacial debris was supplied by the Greenland Ice Sheet or by a local ice cap. The mountain range on Store Koldewey forms a barrier for the Greenland Ice Sheet, which extended eastward onto the shelf during the LGM. Therefore, the Duck Lake basin to the east of the mountain range might have been affected only by an expanded local ice cap. With respect to the timing of the glaciation, Hjort (1981) suggested two glacial advances on Store Koldewey, the oldest (the Muschelberg stadial) of mid Weichselian age and the youngest (the Nanok stadial) dated at *c.* 13 to 9 kyr BP. The sediments in unit A and B suggest a succession from glacio-lacustrine conditions to glacier coverage associated with partial erosion of unit A. This was followed by a retreat of the ice margin with renewed glacio-lacustrine sedimentation. Since there is no evidence for subaerial or subaquatic sedimentation over a longer time period prior to the onset of biogenic sedimentation in the early Holocene, we suggest that the Duck Lake basin was glaciated during the LGM and became ice-free during the subsequent deglaciation. This interpretation is in line with data from E and NE Greenland, which indicate that the outer coastal areas became ice-free close to the Pleistocene-Holocene transition (Bennike & Björck 2002).

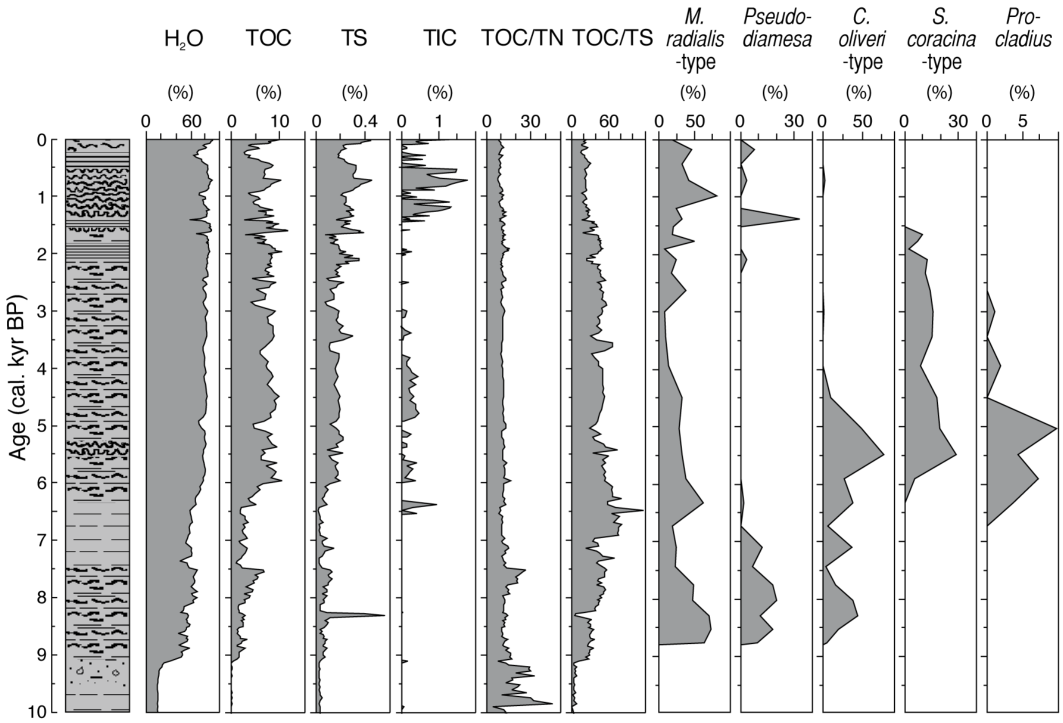

Fig. 7.7: Water content, major biogeochemical properties and the most common chironomids in core Lz1103 from Duck Lake plotted according to the polynomial function presented in Fig. 7.6.

7.5.2 Early Holocene (9.1–6 cal. kyr BP)

At *c.* 9.1 cal. kyr BP, slowly increasing TOC values are evidence of the first significant biogenic production in Duck Lake after the LGM. TOC/TN mass ratio (Fig. 7.7) balances at values around 10, indicating a predominantly autochthonous biogenic production (Meyers & Teranes 2001), presumably in an open basin comparable with recent conditions. Although the early Holocene is regarded as the Holocene thermal maximum in East Greenland (Wagner *et al.* 2000; Kaufman *et al.* 2004), the organic matter accumulation in Duck Lake remained low. Besides their sensitivity to climatic variations, lakes are also susceptible to other environmental changes. The early occurrence of aquatic mosses in the record are at odds with a restricted light availability due to meltwater input with high suspension load from decaying ice in the catchment. Another limiting factor could be lack of nutrients. Since present soil formation on Store Koldewey is limited, nutrient availability following the last deglaciation must have been even lower. This is supported by the occurrence of *Micropsectra radialis*-type and *Pseudodiamesa,* chironomids, which are both indicators of oligotrophic conditions.

The distinct sulphur peak at about 8.3 cal. kyr BP (Fig. 7.7), represented by increased sulphur content in three samples, can most likely be attributed to reduced oxygen saturation in the water column, such as caused by perennial lake ice cover. Tentatively, we suggest that this could be a result of the pronounced cooling seen in the Greenland ice cores at 8.2 cal. kyr BP (Alley *et al.* 1997; Johnsen *et al.* 2001). The deviation of 100 yrs in our chronology is then most likely a result of an inaccuracy in the reservoir correction of the aquatic mosses.

From 7.5 to 6 cal. kyr BP the TOC content was low and aquatic mosses were lacking. This change is preceded by a distinct peak in TOC/TN ratios (Fig. 7.7). TOC/TN ratios are used to distinguish between autochthonous and allochthonous organic matter with values of 10 indicating an origin from non-vascular aquatic plants, whereas values of about 20 and higher reflect a terrestrial origin (Meyers & Teranes 2001). The peak in TOC/TN ratio most likely points to a shift in sediment source. An amplified terrestrial input might be accompanied by increased sediment input, thus terminating favourable conditions for water mosses by light restriction or burial of the mosses. The increased input of minerogenic sediment could be due to increased melting of permafrost in the catchment of Duck Lake.

The TOC/TS ratio shows a general increase from 9 cal. kyr BP, and culminates at 6.5 cal. kyr BP (Fig. 7.7). High TOC/TS ratios may indicate oxygen saturated bottom water conditions (Cohen 2003), such as they are promoted by more open water conditions and enhanced mixis of the water column. The increase of the TOC/TS ratio topwards hence seems to correlate with reduced shelf-ice coverage that has been reported to occur between 7.7 and 4.5 cal. kyr BP in North-East Greenland (Bennike & Weidick 2001).

Overall nutrient-poor conditions during the early Holocene in Duck Lake are corroborated by low abundance and diversity of the chironomid assemblages dominated by *Corynocera oliveri*-type, *Pseudodiamesa* and *Micropsectra radialis*-type.

7.5.3 Middle Holocene (6–3 cal. kyr BP)

The re-appearance of aquatic mosses at about 6 cal. kyr BP may be related to reduced in-wash of minerogenic sediment. The higher productivity and increased presence of mosses are reflected in higher TOC and TS percentage values. With the increase of organic matter accumulation shortly before 6 cal. kyr BP, the chironomid taxa *Sergentia coracina*-type and *Procladius* appear in the record (Fig. 7.7). *S. coracina* is known to occur in cold waters (Walker 1991), but may also occur in shallow lakes in subalpine regions (Brooks & Birks 2001). The appearance of *S. coracina*-type and *Procladius* chironomids along with higher TOC values, seem to reflect improved living conditions in Duck Lake, possibly as a result of increased nutrient availability. This assumption is supported by the co-occurrence of inorganic carbon at the same time, which can be due to carbonate precipitation following increased productivity.

At about 4.5 cal. kyr BP, the chironomids *Procladius* and *C. oliveri*-type decline in abundance. Possibly, this is related to deteriorating living condition perhaps due to a slight long-term decrease in summer temperatures, as is also suggested by decreases in the TOC/TS ratio.

7.5.4 Late Holocene (3 cal. kyr BP to present)

Low TOC/TS values during the late Holocene, with a minimum shortly before 2 cal. kyr BP and a decrease after 1.5 cal. kyr BP, in the Duck Lake record suggest prolonged lake ice cover. TIC maxima at 1.3 and 0.8 cal. kyr BP, punctuated by a minimum at 1 kyr BP (Fig. 7.7), most likely represent phases of high productivity during periods of warmer summers. Two warm peaks separated by a cold interval around 1 kyr BP have been registered in other records from East and North-East Greenland (Wagner et al. 2000; Wagner & Melles 2001). Distinctly lower chironomid concentrations in the late Holocene might be influenced by frequent changes of temperatures. The first reappearance of Pseudodiamesa and an increase of Micropsectra radialis-type since the early Holocene possibly indicate high oxygen availability and overall cooler conditions, which again could have led to a scarcity of nutrients.

7.6 Conclusions

We conclude that the sediment record of Duck Lake represents a succession from non-glacial via glacial to non-glacial settings. The highly compacted and fine-grained sediments at the base of the Duck Lake record and the overlaying ice-proximal sediments indicate that the lake basin was overridden by glacier ice during the LGM. Thus, we do not find support for the presence of ice-free areas on the eastern part of Store Koldewey during the LGM, as suggested by Hjort (1981). Deglaciation of the lake basin is reflected by glacio-lacustrine sedimentation followed by a limnic setting similar to the present that commenced latest at 9.1 cal. kyr BP, as evidenced by a radiocarbon age of aquatic mosses. Biogenic production in Duck Lake remained low until 6 cal. kyr BP, which is at variance with more favourable conditions and higher lake bioproductivity seen in other North-East Greenland lakes during this time span (Wagner et al. 2000). This contrast is most likely a result of limited nutrient availability and in-wash of minerogenic sediment. However, the marked cold event at 8.2 kyr BP may be represented by a pronounced TS peak reflecting reduced oxygen supply to the bottom water due to long lasting lake ice coverage of Duck Lake. During the middle Holocene, bioproductivity in Duck Lake remained relatively constant. In contrast, the late Holocene biogenic production shows clear variability reflecting temperature changes, such as known from other records in central East Greenland.

In general, the Holocene sediment record from Duck Lake shows that the interpretation of biogeochemical data with respect to climatic and environmental evolution in high arctic regions can be difficult. Following deglaciation, the low bioproductivity in Duck Lake hampers a detailed climatic reconstruction. Once nutrient levels and the biogenic production are higher, climatic reconstructions are possible. Therefore climatic and environmental reconstructions based on sediments from high arctic lake sediments have to be interpreted with caution.

Chironomid assemblages seem to have potential as a tool for climate and environmental reconstructions in high arctic lakes. Concentration and diversity of chironomids coincides with the biogenic production of Duck Lake during the Holocene. The fluctuation in the diversity and the low concentration of chironomids in the late Holocene seem to reflect distinct temperature changes during this period. In addition to temperature and oxygen content, nutrient availability seems to be the limiting factor for the fossil chironomid assemblages in Duck Lake throughout the Holocene. More information on the ecology and distribution of chironomids in the arctic lakes in East Greenland is essential to produce more detailed reconstructions of past limnic environments based on fossil chironomid assemblages.

8 Repeated short-term bioproductivity changes in a coastal lake on Store Koldewey, North-East Greenland, an indicator of varying sea-ice coverage? *

8.1 Introduction

For a long time the Holocene has been considered a period of relatively stable conditions compared to the climatic variability during the last glacial stage. However, during the past years investigations increasingly focused on the recent interglacial and attention has been directed towards high-resolution records that reveal variability on shorter timescales. The North Atlantic region, with its essential role in the formation of North Atlantic Deep Water (NADW) and impact on northward heat transport, is an important study region for understanding Holocene climatic variability. Information about the Holocene climatic history can be obtained from on- and offshore archives. Ice core studies from Greenland reveal that the temperature after the abrupt increase of about 10°C at the Younger Dryas-Holocene transition (Alley, 2000) culminated during the Holocene Thermal maximum with temperatures around 2°C warmer than present (Dahl-Jensen *et al.*, 1998; Johnsen *et al.*, 2001). Besides the Holocene long-term temperature trend repeated short-term fluctuations occurred, of which the cooling event at 8.2 kyr BP with a temperature decrease up to 8°C (Alley *et al.*, 1997; Johnsen *et al.*, 2001; Thomas *et al.* 2007) is the most pronounced. Offshore records from the Denmark Strait region reveal repeated temperature changes. For example, sea-surface temperature reconstructions from alkenone ratios on the Northern Iceland shelf show millennial scale oscillations of about 2°C. They are closely related to glacier advances in Northern Iceland. But from the mid Holocene the oscillations are increasingly connected to the strength of NADW formation (Bendle and Rosell-Melé, 2007). Bond *et al.* (1997) documented cold conditions in the Denmark Strait region with a pervasive periodicity of 1.5 kyr changes that may be forced by changes in solar activity. On the East Greenland shelf sea-surface cooling periods recorded by peaks in the carbonate flux are seen after 4.7 cal. kyr and are interpreted as increased flux of polar water and sea ice in the EGC (Jennings *et al.*, 2002). However, low resolution due to low sedimentation rates and chronological problems due to varying reservoir ages can hamper the interpretation and exact timing of marine records.

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Lake sediments record climatic and environmental changes with higher resolution and can provide reliable chronologies. Therefore, lacustrine records from coastal areas in Greenland can form an important link between ice core records and offshore records. Investigations of coastal lake sediments from southern Greenland indicated that lacustrine bioproductivity might be partly influenced by solar activity (Andresen *et al.*, 2004). Additionally, during the late Holocene, cold periods with indications of diminished sea-salt spray were interpreted as the influence of increased sea-ice coverage of the EGC (Andresen *et al.*, 2004).

Sea ice extent along the East Greenland coast is known to have been variable throughout the Holocene. Investigations of microfossil assemblages in marine sediments (Koç et al., 1993) show that the sea-ice extent in the EGC retreated during the early to mid Holocene thermal maximum and increased during the late Holocene. At about 5 cal. kyr BP the summer sea-ice margin was in front of Scoresby Sund (Wagner and Melles, 2001). However, because of low resolution or different scope these investigations could not resolve short-term fluctuations of the sea-ice extent.

During the last decades Arctic sea-ice extent has declined increasingly and ice-free conditions during late summer in the Arctic Ocean may be realized from 2050 onwards (Stroeve *et al.*, 2007). Sensitivity experiments indicate that the retreat of sea-ice may also affect the growing season both in the North Atlantic region and in Europe (Smith *et al.*, 2003). Lakes in coastal high arctic areas seem to be susceptible to varying oceanic conditions and sea-ice coverage. Lake sediment records can therefore help to understand temporal and spatial variability that is important to predict future arctic climatic and environmental changes.

8.2 Regional setting

Store Koldewey is an elongated island at the coast of North-East Greenland between 76°45' and 75°55' N and 19°10' and 18°27' W. The island is separated from the Greenland mainland by the Dove Bugt embayment and is bordered eastwards by the Greenland Sea (Fig. 8.1A). The morphology of Store Koldewey is dominated by 500 to 900 m high, flat-topped mountains on the western part and a 100 to 200 m plateau on the eastern part (Fig. 8.1B).

The mountain range consists of Caledonian crystalline rocks, mainly gneisses, and the eastern plateau of Jurassic, Cretaceous and Pliocene marine sediments with widespread Quaternary till and raised marine and littoral sediments (Bennike *et al.*, 2004; Escher and Pulvertaft, 1995; Henriksen, 2003). Lakes of various dimensions are found on the eastern

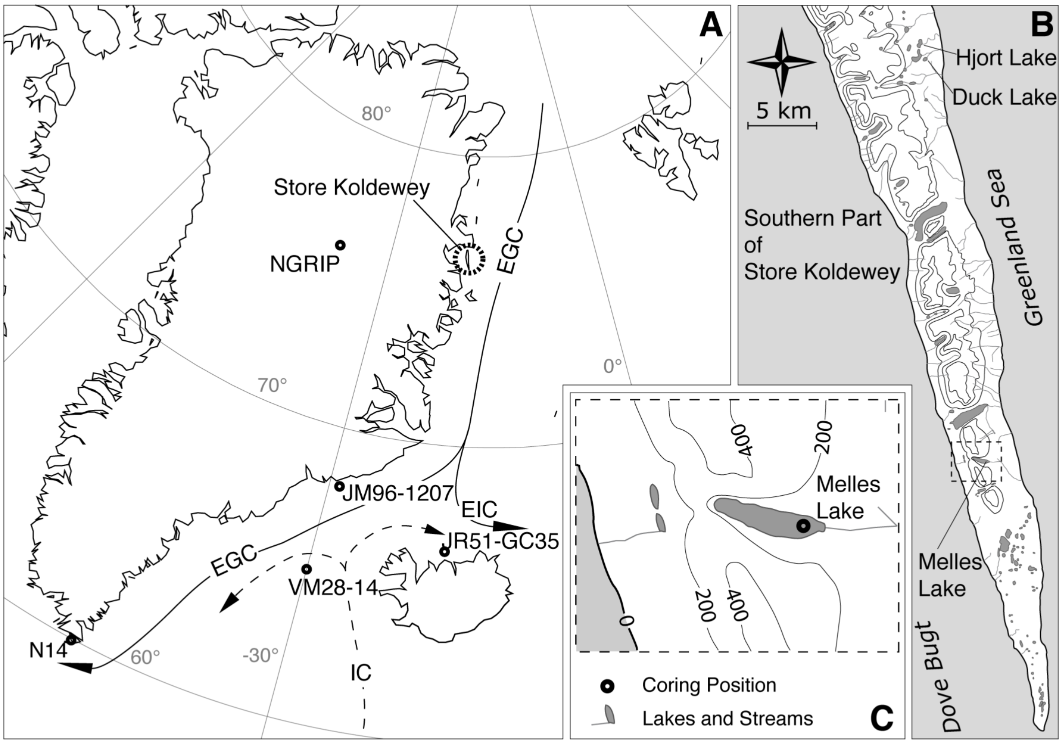

Fig. 8.1: A) Location of Store Koldewey on the coast of North-East Greenland, ice core and marine records mentioned in the text and simplified course of oceanic surface currents, (EGC=East Greenland Current, EIC=East Iceland Current, IC=Irminger Current, warm currents with dashed, cold currents with full line) that influence the mentioned proxy data in marine and coastal records; B) southern part of Store Koldewey with location of Melles Lake and other lakes mentioned in the text, C) enlargement with Melles Lake and surrounding area showing topographical features and coring position, for detail and bathymetry see Fig. 8.2.

plateau and in the valleys (Fig. 8.1B). Meteorological measurements at Danmarkshavn weather station north of Store Koldewey revealed mean monthly temperatures of -23°C in February and 4°C in July and a mean annual precipitation of 141 mm, coastal fog is very frequent in summer (1961–1990; Cappelen *et al.*, 2001).

The flora on Store Koldewey is high Arctic with a patchy distribution of dwarf shrubs (*Salix arctica*, *Dryas octopetala* and *Cassiope tetragona*), grasses, sedges, mosses and lichens resulting in only initial soil evolution (Bennike *et al.*, 2004; Cremer *et al.*, 2005). Various breeding and migratory birds and mammals, such as geese, polar bear, arctic fox, arctic hare and lemming, represent the local vertebrate fauna.

The unofficially named Melles Lake fills an E-W elongated basin on southern Store Koldewey at an elevation of 166 m. Steep slopes are found close to the lake in the west whereas its eastern part borders to the lower plateau (Figs. 8.1C, 8.2, and 8.3). The western slopes surrounding Melles Lake are covered by blockfield, the eastern plateau by fine to coarse sediments forming raw to initial soils (Fig. 8.3). The lake is probably ice free for 1-2 months each year.

8.3 Material and methods

8.3.1 Fieldwork

Fieldwork was conducted in August 2003. Melles Lake is about 1500 m long and 350 m wide. Bathymetrical measurements along several profiles showed a bipartite basin morphology. A 72 m deep western basin with steep slopes is separated by a 4.2 m deep threshold from a 6.9 m deep basin in the eastern part of the lake (Fig. 8.2). Water samples were collected along a vertical profile at the maximum water depth with a 5 l water sampler (UWITEC Co., Austria) and the samples were analysed for oxygen content and oxygen saturation, temperature, pH (WTW Oxi196 probe), and specific conductivity (WTW LF 197; both WTW Corp., Germany).

Coring was performed from a floating platform in the eastern, shallow basin (water depth

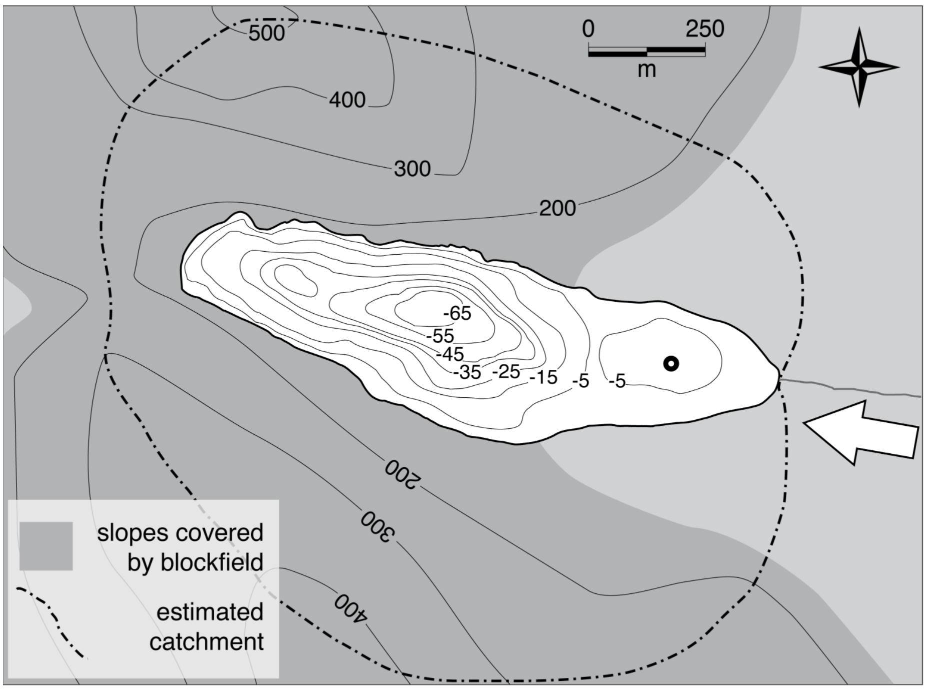

Fig. 8.2: Detail with Melles Lake and surrounding topography. Bathymetrical measurements were conducted along crossing profiles using hand-held GPS and echo sounder and show a bipartite basin morphology with the deepest part in the west. The contour interval is 5 m below lake surface. The bold circle in the shallower eastern basin marks the coring position where water depth was - 6.9 m. The dot-and-dashed line enclosures the estimated catchment area, the drainage is towards the east. The white arrow indicates the view direction in Fig. 8.3.

Fig. 8.3: View towards the west over Melles Lake with the outlet in the foreground. The slopes surrounding the western part are covered by blockfields. Parts of the summer 2003 field camp are in the middle right. View direction is as shown in Fig. 8.2 (Photograph courtesy of Daniel Winkelmann).

6.9 m) at 76°07'35''N and 018°36'39''W using a gravity corer to retrieve undisturbed near-surface sediments and a piston corer for deeper sediments (both UWITEC Co., Austria). Successive application of the piston corer enabled us to recover deeper sediments. A detailed description of the coring procedure is given by Melles *et al.* (1994).

8.3.2 Laboratory work

In the laboratory, the cores were split length-wise, photographed and described. One core half was stored for future work, and the other half of the complete core sequence was divided into one-centimetre interval samples excluding the disturbed material adjacent to the core liner. In order to measure all properties on the same horizon samples were split into aliquots.

Total carbon (TC), total sulphur (TS) and total nitrogen (TN) content were measured using a VARIO EL III analyser (ELEMENTAR Corp., Germany); total organic carbon (TOC) of carbonate-free samples was determined with a METALYT CS1000S analyser (ELTRA Corp., Germany). Calculations of total carbon and total organic carbon revealed no significant differences, showing that no inorganic carbon is present in the sediment. Because TN values are strongly correlated with TOC changes TN values are not discussed. TOC and TS are given in percentage of dry weight. Biogenic silica (BSi) content in each sample from the upper 45 cm and in every second sample between 45 and 189 cm was incrementally measured with the SAN++ Automated Wet Chemistry Analyser (SKALAR Co., Netherlands) according to the method described by Müller and Schneider (1993). The biogenic silica content was calculated from the extrapolated intercept concentration (DeMaster, 1981). For discussions of the long-term trends, the TOC, TS, and BSi data have been smoothed by applying a Stineman function using the KaleidaGraph software package (SYNERGY SOFTWARE Inc., USA) with a geometric weight applied to the current point ± 10% of the data range.

A total of 1 to 2 g sediment from every second sample was wet sieved to separate the fraction >1 mm. The <1 mm fraction was treated with H_2O_2, disaggregated for 12 h using sodium diphosphate and analysed with a laser diffraction particle size analyser (LS 200, BECKMAN COULTER Inc., USA). Lower, median and upper quartiles are from the cumulative frequency curves.

The concentration of aquatic moss in the sediment was visually determined in one-centimetre intervals. 0 means sediment without moss; 1, scattered moss remains; 2, few mosses; 3, common moss remains; and 4, abundant mosses. Moss remains from distinct layers are tentatively identified as *Warnstorfia sarmentosum* and *Warnstorfia exannulatus*. The latter is one of the most common aquatic mosses in lakes in North-East Greenland whereas the semi-aquatic *Warnstorfia sarmentosum* occurs preferably in mires but also submerged in lakes (efloras Online, 2008).

Ten samples of moss remains were dated at the accelerator mass spectrometry (AMS) facility at the Leibniz-Laboratory for Radiometric Dating and Stable Isotope Research, University of Kiel. The dates were calibrated into calendar years before present (cal. yr BP) using the calibration program Calib version 5.0.2 (Stuiver and Reimer, 1993) based on the calibration data set IntCal04 (Reimer *et al.*, 2004; Table 8.1). The age-depth correlation is based on polynomial functions with the best fit to the calibrated dates. $\delta^{13}C$ values are from the AMS machine, these values are not comparable with $\delta^{13}C$ values measured with conventional mass spectrometry.

Table 8.1. Results of 14C AMS dating of moss remains, δ 13C values are from AMS.

Sample	Depth (cm)	Radiocarbon Age BP	δ 13C ‰	cal Age BP (±2σ)
KIA-34397	8-9	397 ± 23	-22.53±0.10	509 - 332
KIA-34398	25-26	1262 ± 28	-25.15±0.21	1281 - 1091
KIA-34399	45-46	2314 ± 25	-23.14±0.14	2357 - 2212
KIA-31909	63-64	3020 ± 27	-21.77±0.11	3333 - 3081
KIA-31310	91-92	4270 ± 29	-21.71±0.08	4870 - 4733
KIA-34400	112-113	5172 ± 28	-21.48±0.13	5991 - 5901
KIA-31911	145-146	6521 ± 30	-24.12±0.17	7552 - 7328
KIA-34401	161-162	7671 ± 33	-20.83±0.18	8541 - 8407
KIA-31912	170-171	8113 ± 47	-21.18±0.08	9262 - 8814
KIA-34796	187-188	9706 ± 49	-25.12±0.07	11235 - 10809

8.4 Results and Discussion

8.4.1 Modern Limnology

Water measurements from August 2003 show a low specific conductivity of 38-37 µS/cm, a high oxygen level (saturation 102-112 %; concentration 11.8-14.0 mg/l), slightly alkaline ph (7.9-8.1), low temperatures ranging from 6.8 to 4.0 °C, and a Secchi depth of 5.4 m (Cremer *et al.*, 2005), which are typical values for oligotrophic to ultra-oligotrophic lakes in arctic regions. The water column showed only a weak stratification indicated by a minor warming and a slight oxygen decrease of the upper 30 metres. The low specific conductivity is due to the low concentration of ions, which reflects the low vegetation cover, only initial soil formation and the low degree of weathering (Bennike *et al.*, 2004; Cremer *et al.*, 2005). Considerable algal growth might be the cause for the overall high oxygen level, whereas the minor decrease within the upper 30 m can be the result of either reduced oxygen solubility due to increased temperature (Cremer *et al.*, 2005) or an increased biotic respiratory activity.

8.4.2 Stratigraphy

The recovered cores were correlated visually by matching distinct horizons. The composite sequence consists of two piston cores. Usually gravity cores are used to get undisturbed surface sediments. It turned out that the ages from the gravity core are not reliable. In contrast to many previous experiences the age of the surface near sediments from the piston core seems to be reliable (cf. Table 8.1, Fig. 8.5). Therefore, only the sediments from the piston core where used.

Table 8.2. Stratigraphy and Units of the Melles Lake sediments.

Depth (cm)	Description	Unit
0-13	clay gyttja with few mosses	C
13-22	layered clay gyttja	
22-26	clay gyttja with mosses	
26-34	layered clay gyttja	
34-45	layered clay gyttja with few mosses	
45-56	clay gyttja with mosses, insulated clay layer	
56-63	layered clay gyttja	
63-67	clay gyttja with few mosses	
67-77	layered clay gyttja, insulated moss layer	
77-79	clay gyttja with few mosses	
79-89	laminated clay gyttja	
89-94	clay gyttja with mosses	
94-106	laminated gyttja with few mosses	
106-111	moss rich gyttja	
111-113	laminated clay	
113-115	gyttja with mosses	
115-118	layered clay gyttja	
118-127	clay gyttja with mosses	
127-129	layered clay gyttja	
129-141	layered gyttja with mosses	
141-144	layered clay gyttja	
144-150	moss rich gyttja	
150-171	layered gyttja with mosses	
171-173	moss rich gyttja	
173-178	layered clay	
178-182	homogenous fine sand	
182-187	layered clay	
187-191	clay with thin moss layers	
191-211	layered clay	B
211-267	homogenous sand with interspersed clay wedges and lenses	
267-271	pebbles with sand	
271-291	upward coarsening with fine sand at the base	
291-297	pebbles and sand	
297-305	clay-dominated, fine-grained homogenous sediments	
305-310	mixture of pebbles of various size and sand, silt, clay	
310-436	dominating clay with silt, partly laminated, stiff consistency	A

The sequence here discussed is 436 cm long and has been divided into three units (Table 8.1). Since this study focuses on Holocene changes of climate and environment the description of the stratigraphy concentrates on the upper 190 cm. Only a brief description of the lower sediment sequence is provided. The interpretation of the pre-Holocene sediment sequence of Melles Lake requires further investigations and the results will be discussed elsewhere.

The lower Units A (436-310 cm) and B (310-190 cm) consist of minerogenic material, and are mainly dark and light grey to brownish. Unit A is dominated by partly layered, fine-grained clayey to silty sediments. Unit B is composed of alternating sand and gravel layers with fine-grained intervals. Biogeochemical analyses show a low content of TOC (c. 0.5%) and biogenic silica (c. 1%) in both units. Unit C (190-0 cm) is fine grained, with randomly distributed single grains of sand. The overall fine-grained, partly layered sediments suggest accumulation under calm conditions such as in a lacustrine environment comparable to the present.

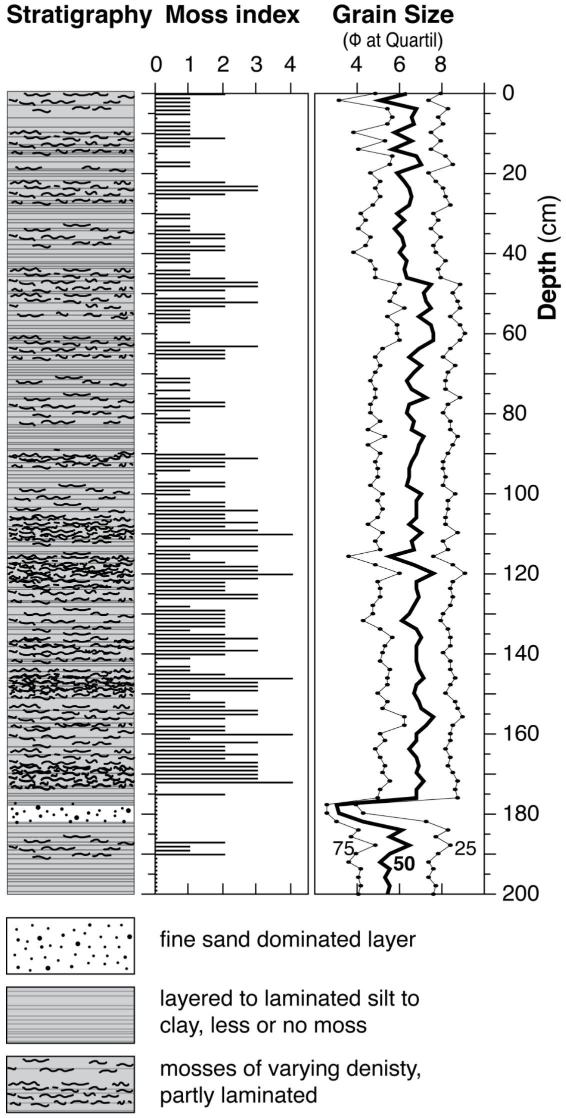

Fig. 8.4: Stratigraphy of the uppermost 200 cm with moss density in 1 cm interval (0 = no moss to 4 = abundant moss) and grain size in 2 cm interval versus depth (25, 50 and 75 percent are from the cumulative frequency curves).

The transition between Unit B and C is gradual, and the boundary is set at the first (rare) occurrence of moss remains, which are found in clay to silt rich layers between 190 and 186 cm depth (Fig. 8.4). The homogenous distribution of the mosses within and the lack of sharp transitions at the base of the layers make a redeposition of the moss remains unlikely. Because Unit A and B lack macroscopical plant remains and show low TOC values the moss remains mark the onset of significant biogenic production in Melles Lake. Following these first remains clay to fine sand dominate the following 14 cm until at 173 cm mosses reoccur.

Above 173 cm moss density increases and the colour of the sediment changes from greyish to olive greenish. The sediments are homogeneous or weakly laminated where mosses are rare. From 100 cm to the top mosses are less frequent, and layers with moss remains alternate with clay-dominated layers (Table 8.2, Fig. 8.4).

8.4.3 Chronology

The chronology of unit C is based on ten AMS ^{14}C age determinations of moss remains (Table 8.1, Fig. 8.5). Aquatic plants in high arctic lakes are considered to be appropriate for ^{14}C dating when dissolved carbon in lake water is in equilibrium with the atmosphere (Björck and Wohlfarth, 2001; Wolfe *et al.*, 2004). However, the exchange can be restricted and age discrepancies between different dated materials can occur (Miller *et al.*, 2005; Oswald *et al.*, 2005; Wagner *et al.*, 2008; Klug *et al.,* 2009). One reason can be that perennial lake ice cover prevents the ^{14}C exchange thus increasing the amount of “old” carbon in the lake, which in turn can affect the equilibrium and result in older ages of aquatic plants. Assuming that *Warnstorfia* spp. growth coincides with the open water season when light availability is highest and the water is mixed, these plants are considered suitable targets for ^{14}C dating. However, considerable reservoir ages on Store Koldewey were reported from two lakes located north of Melles Lake: Hjort Lake with 200 ^{14}C yr (Wagner *et al.*, 2008) and Duck Lake with 520 ^{14}C yr (Klug *et al.,* 2009) (Fig. 8.1). These lakes are surrounded by Jurassic and Cretaceous marine sediments and the possible carbon content in these sediments might have distorted the ^{14}C ages. Melles Lake is located in an area of crystalline bedrock without carbonates, and we suggest that the aquatic and semi-aquatic moss ages from Melles Lakes are free from hard-water effects. Terrestrial material for dating was not found in the sediment.

Several ^{14}C plateaus with near constant ^{14}C ages hampers calibration during the early Holocene. Two of these occurred from 10 to 9.9 ^{14}C kyr and from 9.6 to 9.5 ^{14}C kyr BP (Björck et al., 1997). The oldest dated moss, sample KIA-34796 from 187–188 cm dated to 9706 ± 49 radiocarbon yr BP, lies between these two plateaus and is therefore well suited for calibration. Sample KIA-34397 is the youngest dated sample; it gave an age of 397 ± 23 radiocarbon yr BP, and is unaffected by ^{14}C from atomic bomb testing. The surface sediment is supposed to represent the present. All samples are of Holocene age and the ages increase progressively with depth. Using a polynomial function including all calibrated ages an age-depth model (Fig. 8.5) was created. The sample resolution is 40 to 60 yr during the period from 0 to 7 cal. kyr BP and about 90 yr for the oldest 4.5 kyr, hence the resolution is on a multi-decadal to centennial scale.

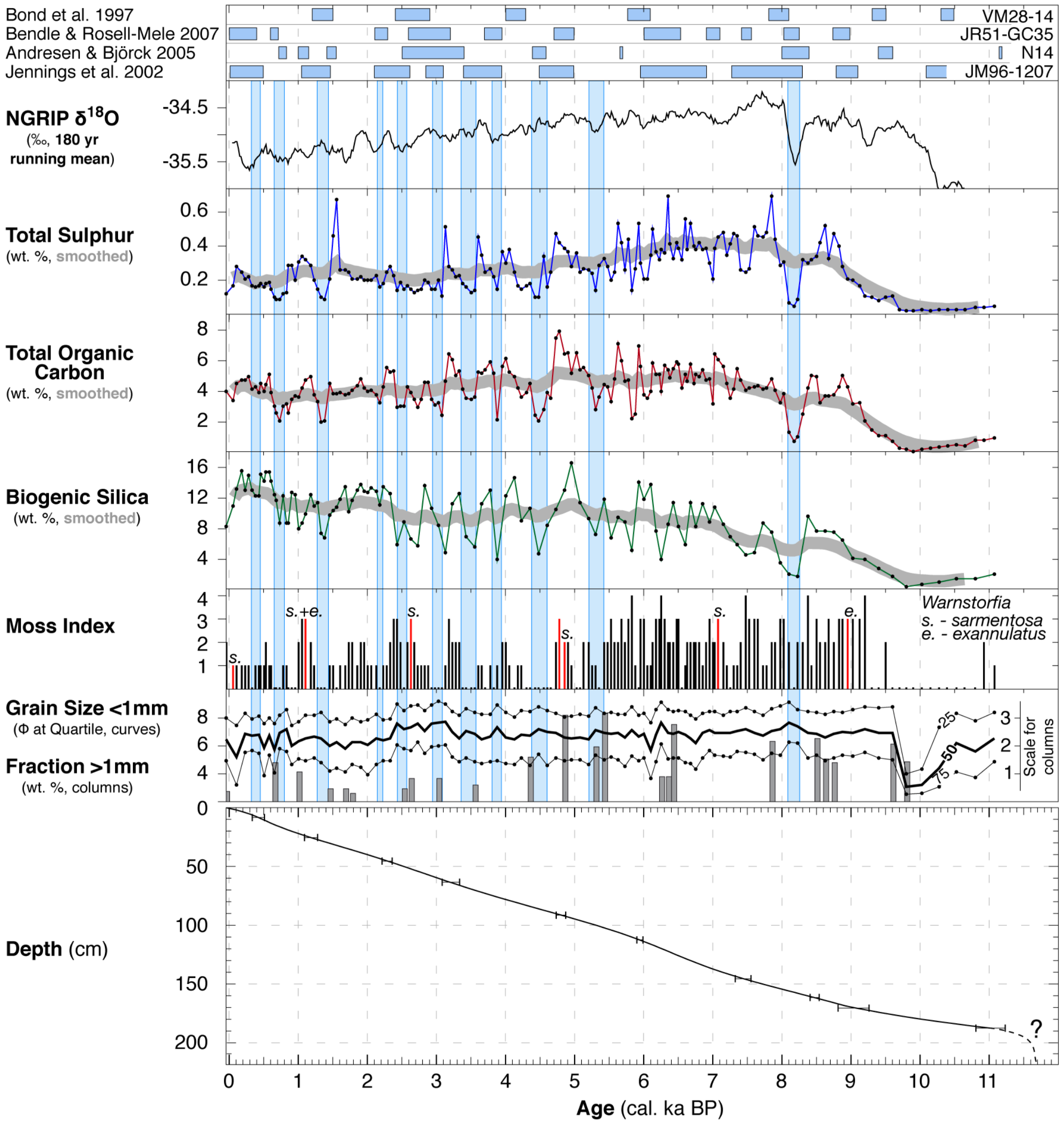

Fig. 8.5: Age-depth model of calibrated ^{14}C ages and selected parameter of the Melles Lake record (plotted versus age) Grain size data smaller 1 mm are plotted in curves at 25, 50 and 75 percent of the cumulative frequency curves. Columns show weight percentage of grains larger than 1 mm, scale for columns is in the right hand side. Moss index shows density of mosses with 0 = no mosses to 4 = abundant mosses. Red columns indicate levels with indentified mosses. Thin lines in biogenic silica, total organic carbon and total sulphur plots are weight percentages, for long-term trend comparison data were smoothed (thick grey line). For comparison, $\delta^{18}O$ values from NGRIP and cold periods in marine and lacustrine records are shown. Periods highlighted with blue bars in the background indicate deterioration of living conditions in the lake that can be related to either atmospheric and/or marine current temperature decreases.

8.4.4 Long-term climatic and environmental changes

Moss remains dated to c. 11 cal. kyr BP give a minimum age for the deglaciation of Melles Lake. The dated mosses are c. 2 kyr older than the oldest ages from two other lakes on Store Koldewey (Wagner *et al.*, 2008; Klug *et al.,* 2009), and older than basal sediments from other lakes in northern East Greenland (Cremer *et al.* 2008). Even though temperatures in coastal (Johnsen *et al.*, 1992a) and central Greenland (Johnsen *et al.*, 2001) increased abruptly after the Younger Dryas and the occurrence of aquatic moss in Melles Lake indicates the onset of biogenic production, organic matter accumulation remained low. One potential reason for this could be input of melt-water from decaying glaciers or from snow covering unstable raw soils in the catchment. Meltwater with a high load of suspended clay and silt could have led to reduced light conditions, which would have hampered photosynthesis. Concentrations of TOC, TS and BSi increase at about 9.5 cal. kyr (Fig. 8.5). This change may indicate that the environment, both in the catchment area and in Melles Lake, stabilised and improved growing conditions were established. The increase of the biogenic proxies matches the onset of biogenic productivity recorded in the two other lakes on Store Koldewey (Wagner *et al.*, 2008; Klug *et al.,* 2009) and in other lakes in East Greenland (Wagner *et al.*, 2000). The smoothed TS values culminate at about 7 cal. kyr BP, but high values are documented from about 8 to 5.5 cal. kyr BP. The high values coincide with the period when aquatic mosses are most common. In contrast to some other records from East Greenland (Wagner et al., 2000), a pronounced period with high smoothed TOC values is not seen. The highest values occur from about 7 to 4.7 cal. kyr BP (Fig. 8.5). Although the maximum TOC values are delayed in comparison with the TS values both proxies appear to reflect enhanced lake productivity due to the early to mid Holocene thermal maximum as it is seen in various archives in East Greenland (Johnsen *et al.*, 1992b; Fredskild, 1995; Wagner *et al.*, 2000; Christiansen *et al.*, 2002; Wagner and Melles, 2002; Kaufman *et al.*, 2004). One reason for the delay of TOC might be that during periods with more open lake-water conditions increased oxygen supply and enhanced decomposition changed the redox potential with the consequence of increased fixation of sulphide in the sediment (Wetzel, 2001b). Furthermore, sulphur-containing organic compounds tend to degrade more slowly than other organic compounds (Wetzel, 2001b) thus reducing the organic carbon with concomitant preservation of sulphur. Previous investigations on Store Koldewey showed that lakes located north of Melles Lake experienced only a slow increase of biogenic productivity during the warmer early to mid Holocene period. Wagner *et al.* (2008) and Klug *et al.* (2009) concluded that insufficient nutrient availability as a result of slow soil development and gradual increasing vegetation cover suppressed the biological productivity in the lakes. However, we do not find a similar development in Melles Lake.

After culminating during the early to mid Holocene TOC and TS concentration and moss density decreases towards the present (Fig. 8.5) and repeated phases with dominating minerogenic sedimentation occurred. These changes most likely mirror deterioration of living conditions that could be a result of decreasing temperatures as seen in coastal and central Greenland ice cores (Johnsen *et al.*, 1992a; Johnsen *et al.*, 2001).

BSi concentration in high arctic lake sediments corresponds to the concentration of diatoms and chrysophytes and is supposed to represent the palaeoproductivity of these algae (Conley and Schelske, 2002). Previous studies of lake sediments in East Greenland showed that measurements of biogenic silica could reflect both climatic change and fluctuations due to internal factors in a lake such as dissolution and nutrient availability (Wagner *et al.*, 2000; Wagner and Melles, 2001). In contrast to the long-term trend of TS and TOC with their more or less pronounced culmination during the early to mid Holocene, the smoothed BSi curve gradually increases during the Holocene (Fig. 8.5). Several factors control the formation and preservation of BSi in sediments (Conley and Schelske, 2002). Organisms that use silica to built skeletal elements are not only dependent on nutrients such as nitrate and phosphate but also on the availability of Si^{4+} ions (Egge and Aksnes, 1992). Measurements of ion concentrations in recent water samples from lakes on Store Koldewey showed that the Si^{4+}-concentration was generally low (Cremer and Wagner, 2004). Progressive weathering in the catchment and the reuse of silica might have increased the Si^{4+}-concentration during the Holocene. Hence it can be assumed that a lower concentration of dissolved silica during the early to mid Holocene could have hampered the growth of diatoms and chrysophytes. The recent diatom community is dominated by the planktonic species *Cyclotella pseudostelligera* HUSTEDT (Cremer *et al.*, 2005). Dissolution of biogenic silica during sinking of phytoplankton is most likely not a concern in Melles Lake but the amount of BSi retained in the sediment might be reduced due to low water retention rates. Moreover, during the Holocene the concentration of BSi increases with decreasing moss content. It could therefore be that the productivity in the early to mid Holocene was dominated by mosses, and by diatoms in the late Holocene. However, apart from some few short periods, when moss and BSi are coevally low, these parameters are negatively correlated, both on a long-term trend but also on the single sample level. Comparable findings were reported by Andresen *et al.* (2004) who argued that enhanced spring-melt surface runoff as a result of intensified winter snow fall provided more clastic and siliceous material to the lake thus lowering light availability for mosses growing submerged in shallow waters and on the lake bottom and in return enhancing the living conditions for diatoms in the water column. However, no evidence for increased precipitation as reflected by an increase in coarse grains or changes in fine grain accumulation is seen in the Melles Lake record. The grain size distribution for the fraction smaller than 1 mm is fairly uniform, and fine-grained sediments dominate throughout the Holocene (Fig. 8.5), which suggests calm conditions. Only a short period with more coarse grained sediments is registered between

10.5 and 9.5 cal. kyr BP, which may be associated with increased sediment input due to increased meltwater from glacier remnants in the catchment area. During the early to mid Holocene the amount of grains larger than 1 mm shows a slight maximum but such grains are sporadic and their occurrence is apparently not related to changes in bioproductivity. Since calm conditions prevailed in the lake the coarser grains most likely represent ice floe-attached sediment from the littoral zone or wind-blown sand deposited on the lake ice and released later during melting.

8.4.5 Short-term changes

Superimposed on the Holocene long-term trend of Melles Lake's biogeochemical and biological proxies repeated short-term fluctuations of bioproductivity are seen. Except for the period from 11 to c. 9.5 cal. kyr BP, when organic matter accumulation in Melles Lake was low and biological productivity was slowly adapting to the postglacial amelioration of living conditions, episodic variability of TS, TOC, BSi and moss content is seen (Fig. 8.5).

Short-term fluctuations of TS, TOC, BSi and moss concentration in Melles Lake occurred both simultaneously and non-simultaneously. Non-simultaneous fluctuations of productivity-related proxies may reflect a dominance of internal factors such as competition between organisms for nutrients and light or the complexity of decomposition and preservation of organic matter. In contrast, fluctuations where all proxies simultaneously decline may indicate a more general deterioration of the living conditions. This may result from an influence of external factors such as an increased lake ice cover following a temperature decline. A number of non-simultaneous fluctuations are seen during the early to mid Holocene between 8 and 5.5 cal. kyr BP. During this period moss concentration and BSi concentration show pronounced anticorrelation. As mentioned above, this anticorrelation might be the result of competition for light between mosses at the lake bottom and diatoms in the water column. From 5.5 cal. kyr BP onward fluctuations of biogenic related proxies seem to occur more synchronized. Here we discuss events during which all biological and biogeochemical proxies show low values, irrespective of intensity or duration. We suggest that such fluctuations are governed by external forcing and therefore reflect climatic variation.

The first indication of deteriorating living conditions is the small decline in TS and TOC concentration and moss content at 9.4 cal. kyr BP (Fig. 8.5). However, because biogenic production was initial and increased from 10 to 8.5 cal. kyr BP, the proxies only decrease weakly. At about 8.3 cal. kyr BP all biogeochemical and biological proxies in Melles Lake started to decrease. For a short period mosses disappear from the record, and at the same time a marked low in TS, TOC and BSi values point to a comprehensive decline in biogenic productivity. Since all biogenic proxies are simultaneously affected, this decline is most likely the result of deterioration of living conditions. According to the age-depth model the lowest values of TS, TOC and moss occurred at about 8200 cal. yr BP and the BSi minimum at about 8100 cal. yr BP but low values are recorded during a period of about 200 years. This pronounced low of biogenic productivity coincides with the most prominent cold climate event during the Holocene, known as the 8.2 cold event, as documented in various marine and terrestrial records (Dansgaard *et al.*, 1993; Alley *et al.*, 1997; Barber *et al.*, 1999; Johnsen *et al.*, 2001; Moros *et al.*, 2004; Andresen and Björck, 2005; Thomas *et al.* 2007). On Hochstetter Foreland a short but distinct cooling around 8300 cal. yr BP is documented in lacustrine sediments (Björck et al. 1994). It can therefore be argued that lake ice cover due to substantial cooling most likely persisted throughout the summer thus deteriorating the living conditions in Melles Lake and only little organic matter accumulated.

From 8.0 to c. 5.5 cal. kyr BP TOC, TS, BSi and moss concentrations show minor fluctuations only and point to fairly stable and favourable living conditions in Melles Lake. After 5.5 cal. kyr BP the short-term variability of bioproductivity-related proxies become more pronounced and Melles Lake experienced several concomitant low levels of bioproductivity. Short-term events occurred at about 4.5, 3.5, 3.0, 2.5, 2.2, 1.3, 0.7, and 0.4 kyr cal. BP. By analogy to the deterioration of living conditions during the 8.2 cold event these low levels of bioproductivity-related proxies may indicate similar conditions in the lake. However, these events are not equally pronounced for all proxies. Most of these events are characterised by fairly long moss-free periods and minerogenic-rich deposition dominated for several centuries, whereas low levels of TS, TOC and BSi were more short-lived. Smol and Douglas (2007) argued that during cooler periods biogenic productivity in deep arctic lakes could be restricted to a shallow open water moat in the littoral zone. Lake ice coverage may persist throughout the short summer thus preventing moss growth due to reduced light availability in the central part of the lake. Therefore the absence of mosses in the Melles Lake sediments may indicate overall cooler conditions with lowest temperatures coinciding with periods of lowest TS, TOC and BSi values.

8.4.6. Atmospheric and oceanic equivalents to the short-term cooling events in Melles Lake

Previous investigations of Greenlandic lakes have revealed that a variety of processes could result in long and short-term changes in lacustrine biogenic productivity (Fredskild, 1995; Wagner *et al.*, 2000; Wagner and Melles, 2001; Andresen *et al.*, 2004; Andresen and Björck, 2005; Wagner *et al.*, 2005; Klug *et al.*, 2009). Andresen and Björck (2005) documented changes of lacustrine biogenic productivity that were concomitant with changes in nuclide production rates. They concluded that at least some cold events recorded in lake sediments in southern Greenland might be related to changes in solar activity. Other investigations of lacustrine records from East Greenland revealed that biogenic productivity in lakes is related to temperature, showing that the insolation maximum during the early Holocene (Berger and Loutre, 1991) and the related temperature maximum play an important role for aquatic organisms and their productivity in high arctic regions (Wagner *et al.*, 2000; Wagner *et al.*, 2005). The long-term trend of biogenic productivity and especially the short-term events of deteriorations of living conditions in the Melles Lake record find, to some extent, counterparts in the $\delta^{18}O$ fluctuations in ice core records. Discrepancies in the timing between various records may be the result of chronological problems. Besides the concomitant changes of biological as well as biogeochemical proxy data with the $\delta^{18}O$ changes in the ice cores, corresponding fluctuations in marine cores of the North Atlantic can be seen (Fig. 8.5). In addition to the pronounced 8.2 cold event during which biogenic productivity in Melles Lake almost stopped, short-term events of deteriorations coincide partly with the cooling events recorded in marine sediment cores in the Denmark Strait region (Bond *et al.*, 1997) and north of Iceland (Bendle and Rosell-Melé, 2007). Cold periods in core VM28-14 (Figs. 8.1 and 8.4) are characterised by rich hematite-stained IRD that originate from Red Beds around the northern North Atlantic and volcanic glass from Iceland but magnitudes of the ocean surface temperature decreases probably did not exceed 2°C (Bond et al., 1997). Sea-surface temperatures (SST) changes on the northern Icelandic shelf (JR51-GC35, Figs. 8.1 and 8.4) are related to the strength of oceanographic influences from the EIC and the IC and variations of the North Atlantic Oscillation (NAO) (Bendle and Rosell-Melé, 2007). In contrast, Store Koldewey and the adjacent sea are not subject to the warm Irminger Current (IC). As it is more than 1000 km north of Iceland the island is also less or not influenced by changes of the NAO. It can therefore be argued that cold periods with only small temperature changes (Bond et al. 1997) or changes of the NAO (Bendle and Rosell-Melé 2007) are of minor impact on the biogenic productivity far north and find consequently only marginal expression in the Melles Lake record. Cold periods recorded in a lacustrine record (N14, Figs. 8.1and 8.4) from southern Greenland (Andresen et al. 2004) coincide more often with those recorded in Melles Lake. Although also influenced by reflected warm water masses from the IC lake N14 is affected by the East Greenland Current (EGC). This current drains about 70% of the Arctic Ocean (Moritz et al. 1990) and carries cold and low-saline water masses down along the East Greenland coast (Buch, 2007).

Cold periods recorded in Melles Lake correlate well with sea-surface cooling events that occurred on the East Greenland shelf (Jennings *et al.*, 2002; core JM96-1207, Figs. 8.1 and 8.4). Jennings et al. (2002) argued that the cooling of sea surface water as seen in increased amounts of carbonate in the sediment is the result of increased sea ice. The simultaneous cooling of the East Greenland shelf surface water and the deteriorations of living conditions in Melles Lake about 1000 km to the north require a mechanism that can connect the temperature history between these terrestrial and marine sites. Because of their location at the eastern Greenland coast both records seem to be susceptible to changes of the EGC. The SST of the EGC varied during the Holocene with a gradual warming of winter SSTs on the East Greenland shelf (Solignac *et al.*, 2006) but with a general decreasing trend towards the present (Andersen *et al.*, 2004a; Andersen *et al.*, 2004b). The temperature of the EGC is mostly affected by meteorological (Darby *et al.*, 2006; Vinje, 2001) and oceanographic conditions (Dyke *et al.*, 1997) in the Arctic Ocean and by the amount of sea ice that discharges from the Arctic Ocean through the Fram Strait. From 5 cal. kyr BP to the present low bioproductivity in Melles Lake coincides with increased sea-ice coverage on the East Greenland shelf (Jennings *et al.*, 2002). During periods with abundant sea-ice along the eastern Greenland coast the ocean-atmosphere temperature exchange was reduced with the result of prevailing low temperatures. Furthermore increased albedo due to the ice coverage itself or snow on ice may additionally reflected solar radiation and therefore hampered the warming of coastal areas. It might therefore be reasonable to suggest that persisting lake ice cover as a result of prevailing cold temperatures prevented biogenic productivity in Melles Lake. The Melles Lake data also suggest that the cold periods 3.9 to 3.5 and 2.6 to 2.2 kyr BP on the East Greenland shelf are composed of two short lasting periods each of which are not resolved in the marine record (Jennings *et al.*, 2002). Although not directly connected it can therefore be argued that the biogenic productivity of this coastal lake has been influenced by the EGC and information about short-term changes of sea-ice coverage can be obtained from the Melles Lake data.

8.5 Conclusions

Based on our chronology and the results of biological, biogeochemical and sedimentological investigations of Melles Lake's lacustrine record the following main conclusions about the Holocene climatic and environmental history of southern Store Koldewey and the causes for short-term deteriorations of living conditions can be drawn:

(1) Dating of aquatic mosses show that the lake basin was deglaciated prior to 11 cal. kyr BP. Although mosses indicate an early onset of biogenic productivity, organic matter accumulation remained low for c. 2 millennia. The early occurrence of mosses predates the onset of biogenic production in two other lakes on Store Koldewey by about 2 kyr. The growth of mosses in Melles Lake also suggests that at least parts of the southern island were already ice-free whereas deglaciation in the middle part of the island continued.

(2) Organic matter accumulation in Melles Lake started to increase at c. 10 cal. kyr BP. The highest concentrations of organic matter are seen in early- to mid-Holocene sediments. The ameliorated living conditions of aquatic organisms during the Holocene Thermal Maximum indicate higher temperatures than currently in North-East Greenland. The Holocene Thermal Maximum ended at c. 5.5 cal. kyr BP.

(3) At 8.2 cal. kyr cal. BP a marked low level in biogenic productivity is seen. It reflects a substantial deterioration in living conditions in Melles Lake and is the first pronounced record of the 8.2 cold event from a lacustrine record in East Greenland. Furthermore, comparison of biogenic productivity changes with temperature changes in NGRIP ice core reveal that lakes in the coastal area of North-East Greenland are also susceptible to atmospheric temperature perturbations on shorter time scales.

(4) Cold conditions on Store Koldewey that led to deteriorating living conditions in Melles Lake coincide with increased sea-ice extent on the East Greenland shelf around 4.5, 3.5, 3.0, 2.5, 1.3 and 0.4 kyr cal. BP. Owing to the resolution and the sensitivity of the Melles Lake record the periods with increased sea-ice extent around 3.5 and 2.5 kyr cal. BP can be divided into two cold periods interspersed with short periods of more favourable conditions.

(5) Lakes in coastal high arctic regions are subject to atmospheric as well as oceanic changes. They are susceptible to short-term climatic changes, which can interfere with a variety of internal processes of biogenic productivity in those lakes. The data therefore also demonstrate the difficulties associated with the interpretation of lacustrine records.

9 Late Weichselian ice-front history on Store Koldewey, North-East Greenland, as illustrated by a lacustrine record *

9.1 Introduction

Information about the spatial and temporal variations of past ice sheets are essential for understanding sea level and climate changes in the past and predict their evolution in the future (e.g. Clark *et al.*, 1999, Clark and Mix, 2002, Wild *et al.*, 2003, Parizek and Alley, 2004, Alley *et al.*, 2005, Overpeck *et al.*, 2006).

The extent of the late Quaternary glaciation and the geographic pattern of the ice margin in East Greenland have been of interest over the last decades and investigated using different archives. The Weichselian glacial chronology of North-East Greenland by Hjort (1981) who suggested a succession of three glaciations of gradually smaller extension has been re-evaluated by Hjort and Björck (1984). Based on geomorphological, lithostratigraphical, and palynological data they concluded that the two older glaciations, the Kap Mackenzie and the Muschelbjerg stadial are of Saalian age or older. Hjort and Björck (1984) also concluded that the youngest glaciation, the Nanok I stadial, occurred during the late Weichselian and was the least extensive one, with ice-free lowlands and some nunataks. Funder (1989) summarised the glacial chronology and the glacial extent during the late Quaternary for East Greenland and showed that the glaciations on central East Greenland correspond in time and extent with those in North-East Greenland. Within the framework of the PONAM (Polar North Atlantic Margins; Late Cenozoic Evolution) project the extent of the Greenland Ice Sheet during the Weichselian and the position of the ice margin during the Last Glacial Maximum (LGM) in North-East Greenland have been subject of numerous studies (e.g. Björck *et al.*, 1994b, Houmark-Nielsen *et al.*, 1994, Landvik, 1994). On Germania Land, evidences of two major glaciations were found. The older, undated glaciation covered the entire area and extended onto the continental shelf. Based on deglaciation dates and the marine limit the younger glaciation has been suggested to be of late Weichselian age with an margin beyond the present coast line and some ice-free low-relief mountains on eastern Germania Land (Landvik, 1994). About 200 km further to the south, marine shells from deltaic sediments associated with the Nanok I moraine indicated that the south-western rim of Hochstetter Foreland was glaciated until the beginning of the Holocene (Björck *et al.*, 1994b). Another 100 km to the south, at Kap Herschell, Wollaston Foreland, Houmark-Nielsen *et al.* (1994) documented a succession of two glacial advances interstratified by the deposition of glaciolacustrine sediments.

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Thermoluminescence dating of the glaciolacustrine unit indicate a maximum age of 43 kyr BP. Based on findings of marine shells older than 43 kyr and final positions of glacigenic geomorphological features Hjort (1981) concluded that the western mountain range of Store Koldewey obstructed and redirected the ice emanating from Greenland and that the lower eastern plateau on Store Koldewey remained unglaciated during the late Weichselian. Based on cosmogenic exposure dating of boulders and unscoured bedrock from the higher plateaus of Store Koldewey, Håkansson *et al.* (2007) suggested that the unscoured mountain plateaus on Store Koldewey were covered at least partly by cold-based ice during the LGM. However, the question whether the ice was dynamically connected to the Greenland Ice Sheet or if it originated from local ice caps remained open. Sedimentary records from lakes on central eastern Store Koldewey indicate glacigenic sedimentation at the Pleistocene/Holocene transition with subsequent lacustrine conditions during the Holocene (Wagner *et al.* (2008) Klug *et al.* (2009). This glacigenic sedimentation can tentatively be related to the late Weichselian glaciation; however, an unambiguous correlation with that time is hampered by dating uncertainties. Lacustrine sedimentary records (Cremer *et al.*, 2008) and macrofossil remains (Bennike and Björck 2002) from North-East Greenland show that the most areas were deglaciated within a period of about 1.5 ka at the beginning of the Holocene.

Here we use a 436 cm long lacustrine sedimentary record from southern Store Koldewey to reconcile the idea of ice-free areas on Store Koldewey during the LGM (Hjort 1981) with the cosmogenic exposure results of Håkansson *et al.* (2007) and the findings of Wagner *et al.* (2008) and Klug *et al.* (2009) from the central part of Store Koldewey.

9.2 Study Area

Store Koldewey is an island located at the coast of North-East Greenland between 75°55' and 76°45' N and 19°10' and 18°27' W. The north to south orientated island is c. 80 km long and up to 11 km wide. In the west the Dove Bugt embayment separates the island from the mainland; towards the east Store Koldewey is bordered by the Greenland Sea (Fig. 9.1). A 500 to 900 m high, flat-topped mountain range dominates the morphology of the western part of Store Koldewey. To the west of the mountain range, steep slopes descend to the present coastline, whilst the area to the east is characterized by a gently declined plateau, which is located at c. 100 - 200 m a.s.l. and ends in a coastal cliff. Several E-W orientated valleys of various dimensions repeatedly dissect the mountain range (Fig. 9.2).

The mountain range on Store Koldewey is formed by Caledonian crystalline rocks, mainly gneisses. The range and the steep western slopes are covered by variously matured

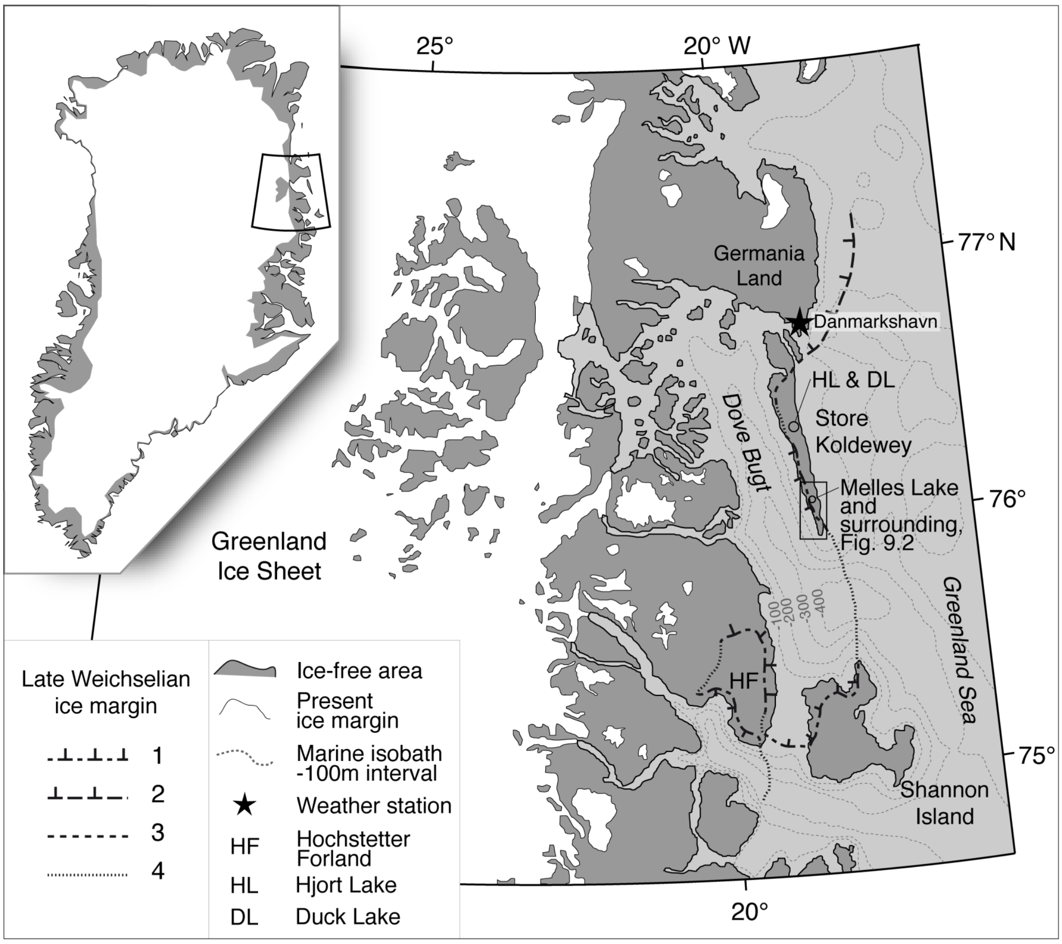

Fig. 9.1: Store Koldewey and adjacent areas with present coast, ice margins and ice-free areas. Melles Lake is located on southern Store Koldewey (box), for detail see Fig. 9.2. Late Weichselian ice margin according to 1 - Nanok Moraines from Hjort (1981) and Hjort & Björck (1984), 2 - Landvik (1994), 3 - Funder (1989), 4 - Correlations.

blockfields. Jurassic, Cretaceous, and Pliocene marine sediments are exposed at the eastern costal cliff. Quaternary till, raised marine and littoral sediments are found on the lower elevation of the eastern plateau (Escher and Pulvertaft, 1995, Henriksen, 2003, Bennike *et al.*, 2004).

Due to the harsh climatic conditions, the flora on Store Koldewey is high arctic. As recorded at Danmarkshavn weather station the monthly temperatures range between -23°C in February and 4°C in July and the mean annual precipitation amounts to 141 mm (Cappelen *et al.*, 2001). The vegetation is patchy and dominated by dwarf shrubs, grasses, sedges, mosses, and lichens. Initial soils partially cover the bedrock (Bennike *et al.*, 2004, Cremer *et al.*, 2005). The local vertebrate fauna is represented by various breeding and migratory birds and mammals, such as geese, polar bear, arctic fox, arctic hare and lemming.

Lakes of various dimensions occur in basins on the eastern plateau and in the valleys (Fig. 9.2). Melles Lake is located on southern Store Koldewey at 166 m a.s.l. The lake fills an E-W elongated, c. 1.5 km long and c. 0.35 km wide basin. Steep slopes surround the lake in the west whereas its eastern part borders to the lower plateau. The lake is exclusively fed by melt water from its estimated catchment area of about 1.6 km^2 and drains towards the east (Klug *et al.*, in press). During summer season the lake is probably ice free for 1-2 months each year. Since the local marine limit is reconstructed to about 53 m a.s.l. (Hjort 1981), the lake basin likely was not affected by a marine transgression.

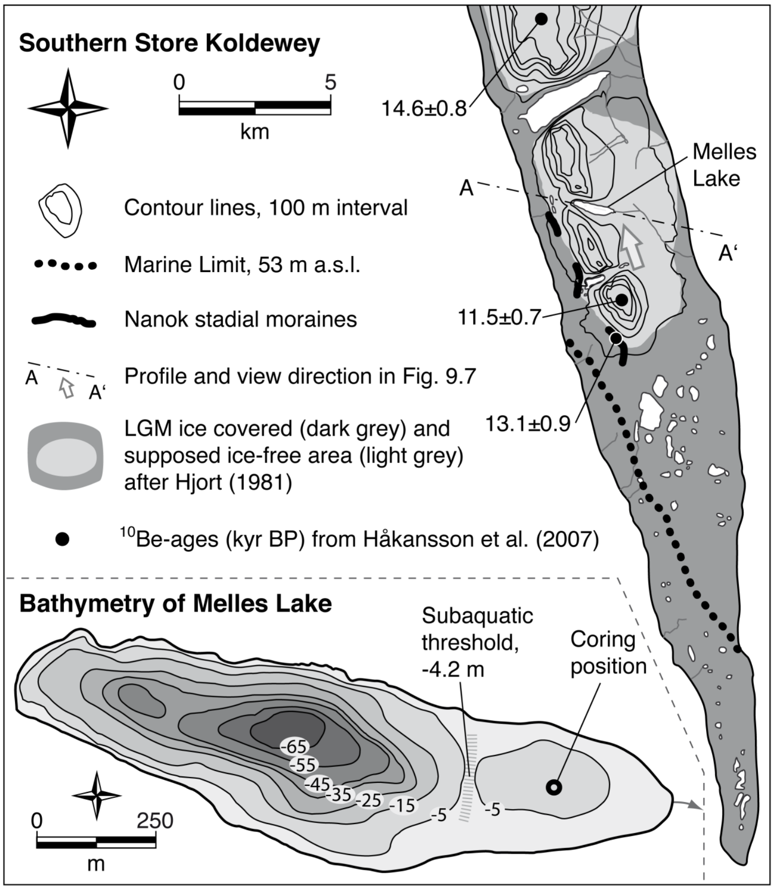

Fig. 9.2: Southern Store Koldewey with geomorphological settings and the location of Melles Lake. The insert shows the bathymetry of Melles Lake and the coring position in the eastern basin at -6.9 m water depth. The outlet of Melles Lake is indicated by the arrow to the east.

9.3 Material and Methods

9.3.1 Basin morphology and sediment coring

Bathymetrical measurements at Melles Lake were carried out in August 2003. Water depths were measured along several crossing profiles using a GPS device and a hand-held echo sounder. Melles Lake is separated by a 4.2 m deep subaquatic threshold into a western basin with a maximum water depth of 72 m and an eastern basin with a maximum water depth of 6.9 m. The lake morphology extends into the catchment, with the steep slopes in the west and a more gently setting in the east (Fig. 9.2).

A sediment core was recovered from the deepest part of the eastern, shallow basin (76° 7' 35" N and 18° 36' 38" W) using a 3 m long piston corer (UWITEC Co., Austria) from a floating platform. A detailed description of the coring procedure is given by Melles et al. (1994). For the uppermost sediment sequence, the piston was released c. 50 cm above the sediment surface. Two runs of the piston corer (cores Lz1101-6 and Lz1101-7) and an overlap of c. 25 cm between the two core segments enabled the recovery of a 436 cm long sediment sequence. After recovery the sediment cores were stored at 4°C until further processing.

9.3.2 Laboratory work

Prior to core opening magnetic susceptibility (MS) was logged in one-centimetre intervals using a MultiSensorCoreLogger (MSCL, GEOTEK Corp., Germany) with a loop sensor that integrates magnetic sediment characteristics 10 cm before and after the measurement point (Weber *et al.*, 1997). Subsequently, the cores were split length-wise, the split-surface was cleaned and the sediment described according to colour, texture, and structure. On one core half, optical and micro-radiographic images and X-ray fluorescence spectrometry (XRF) elemental profiles were measured with an ITRAX core scanner (COX Analytical Systems, Sweden). Radiographic images were taken with 1 s exposure time and 200 µm step size. XRF elemental profiles were measured with 1 mm step size and 20 s exposure time. A more detailed description of the measurement procedure of the ITRAX core scanner is given by (Croudace et al., 2006). Since surface roughness of split cores, sediment compositional variability, content of water and organic matter, textural and porosity changes, and degree of compaction can influence the detection (Croudace et al., 2006, Rothwell and Rack, 2006), elemental counts derived from XRF core logging should be regarded as semi-quantitative (Rothwell and Rack, 2006) and the elemental profiles have to be interpreted carefully.

Lithological description and core logging results were used to correlate the two piston core segments to a composite sequence. After correlation, one core half was sectioned into one-centimetre interval samples for subsequent laboratory analyses, whilst the other half was stored for future work.

All samples were freeze-dried and the water content was calculated from the mass difference between wet and dry samples.

After grinding and homogenizing of aliquots, total carbon (TC) content was measured using a VARIO EL III analyser (ELEMENTAR Corp., Germany). Total organic carbon (TOC) was determined in 1 cm intervals from 0-250 cm and in 4 cm intervals from 251-436 cm with a METALYT CS1000S analyser (ELTRA Corp., Germany) after pre-treatment with 10% HCl at 80°C to remove the carbonate. Since the differences between the amounts of TC and TOC are negligible, it can be assumed that TC is formed almost entirely by TOC. TOC is given in percentage of dry weight.

Grain-size analyses were carried out in 2 cm intervals from 0-234 cm and in 4 cm intervals from 235-436 cm. For sample preparation, about 1 to 2 g sediment were separated from particles > 1mm by wet sieving. The residue was dried and weighted, and then treated with H_2O_2 to remove organic compounds. Subsequently, the samples were disaggregated for 12 h using sodium diphosphate. After 10 sec of ultrasonic pre-treatment, the samples then were analysed with a laser diffraction particle size analyser (LS 200, BECKMAN COULTER Inc., USA). Lower, median and upper quartiles are calculated from the cumulative frequency curves.

For age determination of the entire sequence ages of ten samples of moss remains from unit C (Klug et al., in press) were supplemented by dating of four bulk samples from the lower part of unit C1 and from unit A using accelerator mass spectrometry (AMS) at the Leibniz-Laboratory for Radiometric Dating and Stable Isotope Research, University of Kiel. Sample pre-treatment and measurements were conducted according to Grootes et al. (2004). Of the bulk sediment samples the humic acid fraction (HA) and the alkali residue (humic acid free fraction, HAF) were individually measured and calibrated into calendar years before present (cal. yr BP) using CalPal (Danzeglocke *et al.*, 2008) and the calibration data CalPal-2007$_{Hulu}$ (Weninger and Jöris, 2008, Table 9.1).

Table 9.1. Radiocarbon ages, depth, material and calendar years BP of samples from Melles Lake. $\delta^{13}C$ values are from AMS and not comparable with conventional measured $\delta^{13}C$. Calendar ages in italics are in 1 SD.

Sample no.	**Depth** (cm)	**Material**	**Weight C** (mg)	**Radiocarbon Age BP**	**$\delta^{13}C$** ‰	**cal Age BP** (±2σ, *±1σ*)
KIA-34397	8-9	Moss remains	3.6	397 ± 23	-22.53 ± 0.10	509 - 332
KIA-34398	25-26	Moss remains	4.3	1262 ± 28	-25.15 ± 0.21	1281 - 1091
KIA-34399	45-46	Moss remains	3.7	2314 ± 25	-23.14 ± 0.14	2357 - 2212
KIA-31909	63-64	Moss remains	4.3	3020 ± 27	-21.77 ± 0.11	3333 - 3081
KIA-31910	91-92	Moss remains	3.4	4270 ± 29	-21.71 ± 0.08	4870 - 4733
KIA-34400	112-113	Moss remains	3.8	5172 ± 28	-21.48 ± 0.13	5991 - 5901
KIA-31911	145-146	Moss remains	4.0	6521 ± 30	-24.12 ± 0.17	7552 - 7328
KIA-34401	161-162	Moss remains	4.1	7671 ± 33	-20.83 ± 0.18	8541 - 8407
KIA-31912	170-171	Moss remains	1.3	8113 ± 47	-21.18 ± 0.08	9262 - 8814
KIA-34796	187-188	Moss remains	1.3	9706 ± 49	-25.12 ± 0.07	11235 - 10809
KIA-34402	201-202	Bulk Sediment-HA	3.2	19340 ± 90	-26.19 ± 0.24	*22830 - 23371*
	201-202	Bulk Sediment-HAF	0.4	22200 +510/-480	-21.82 ± 0.09	*25888 - 27520*
KIA-34797	331-332	Bulk Sediment-HA	2.1	37020 +730/-670	-25.42 ± 0.81	*41378 - 42349*
	331-332	Bulk Sediment-HAF	0.75	31200 +900/-810	-25.08 ± 0.25	*34620 - 36639*
KIA-34403	367-368	Bulk Sediment-HA	0.8	30620 +770/-700	-26.94 ± 0.21	*34204 - 35609*
	367-368	Bulk Sediment-HAF	1.3	34770 +800/-730	-25.70 ± 0.08	*38789 - 40832*
KIA-34798	431-432	Bulk Sediment-HA	1.3	37220 +1130/-990	-24.85 ± 0.12	*41190 - 42763*
	431-432	Bulk Sediment-AR	0.8	29970 +740/-680	-28.71 ± 0.64	*33467 - 34868*

9.4 Results and Discussion

9.4.1 Stratigraphy

The stratigraphy of core Lz1101 from Melles Lake has been recently described by Klug *et al.* (in press). The description, however, focused on the biogenic sediments in the uppermost 190 cm of the record (Klug et al. in press), and therefore is not useful for a detailed characterization of the more clastic sediments in the lower part of the sequence.

Unit A spans from the base of the core at 436 cm to 311 cm and consists of minerogenic material with a dark grey colour and a stiff consistency (Fig. 9.3). The sediment is almost homogenous and exhibits only faint laminations (Fig. 9.4). Unit A is dominated by clayey silt. The proportion of fine sand is small, but increases towards the top of the unit.

Unit B spans from 311 to 212 cm). It is also composed of minerogenic material and has mainly dark grey colour with dark brownish intersections. Main characterization of unit B is a significant increase in grain-size composition, with alternating sand and gravel layers and some interspersed fine-grained intervals. Occasionally, subangular rocks, which considerably exceed the size of the gravel, are observed. Overall, coarsest sediments were deposited at 270 cm, with an upward coarsening trend below and a gradual grain-size decrease above, resulting in a mixture of coarse to fine sand at the top of unit B (Figs. 9.3 and 9.4).

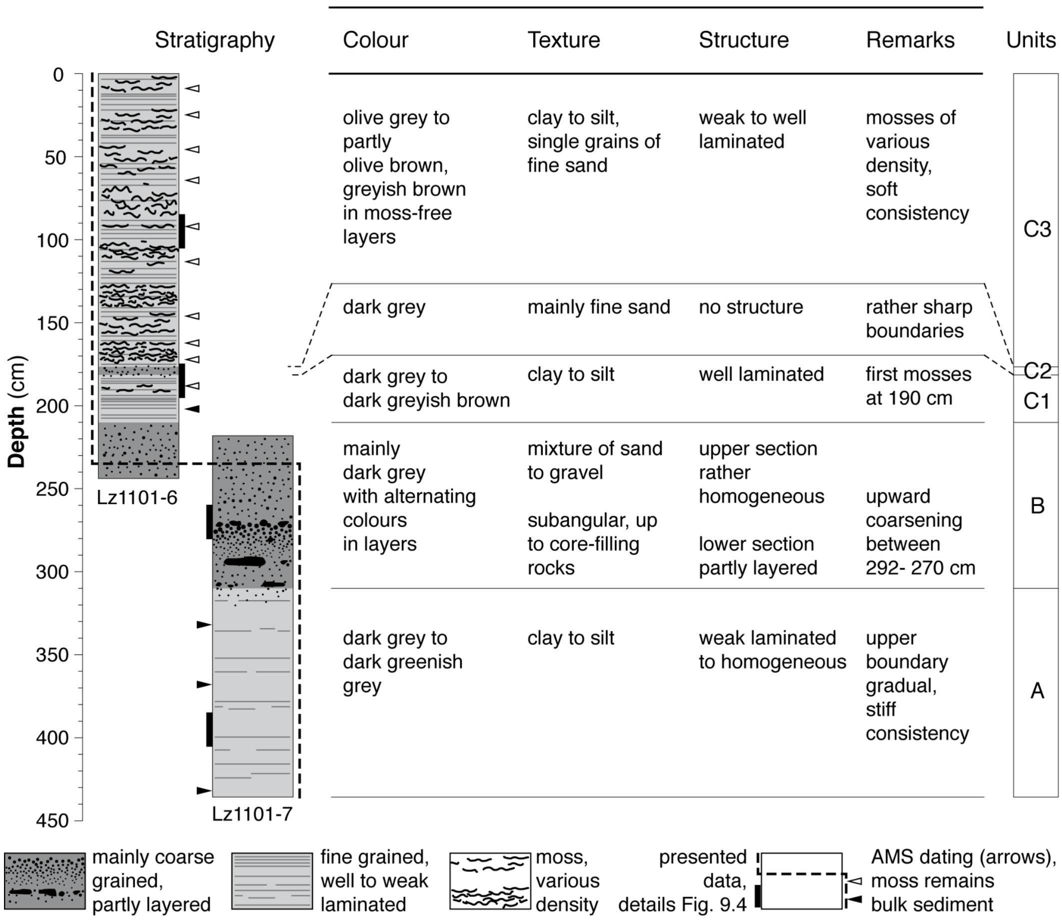

Fig. 9.3: Stratigraphy and main characteristics of cores Lz1101-6 and Lz1101-7 from the eastern basin of Melles Lake.

Unit C extends from 212 cm to the top of the sequence, but can be separated into three sub-units. Unit C1 spans from 212 to 181 cm and consists of minerogenic sediment that is finely laminated (< 1 cm; Fig. 9.4) with a dark greyish brown to dark grey colour. Occasional moss remains are found at 190 cm depth (Fig. 9.3). The sediment is silt dominated, but some grains of > 1 mm occur. A layer of medium to coarse sand at the depth of 181-177 cm forms unit C2. Bending of the laminated sections above and below the layer (Fig. 9.4, second column left hand side) likely point to dispersion due to collapse during coring. From 177 cm to the top, again laminated clayey silt with occasionally interspersed sand grains occurs and forms unit C3. Mosses reoccur from about 173 cm topwards (Fig. 9.3) and the minerogenic dominated sedimentation shifts into a biogenic-minerogenic sedimentation. The mosses, identified as the aquatic species *Warnstorfia exannulatus* and *Warnstorfia sarmentosum* (Klug *et al.*, in press), occur in layers of varying thickness (Fig. 9.4), with an overall decrease topwards. The sediment colour depends on the occurrence of moss layers and changes from dark and olive grey in moss layers to olive brown and greyish brown in clastic layers.

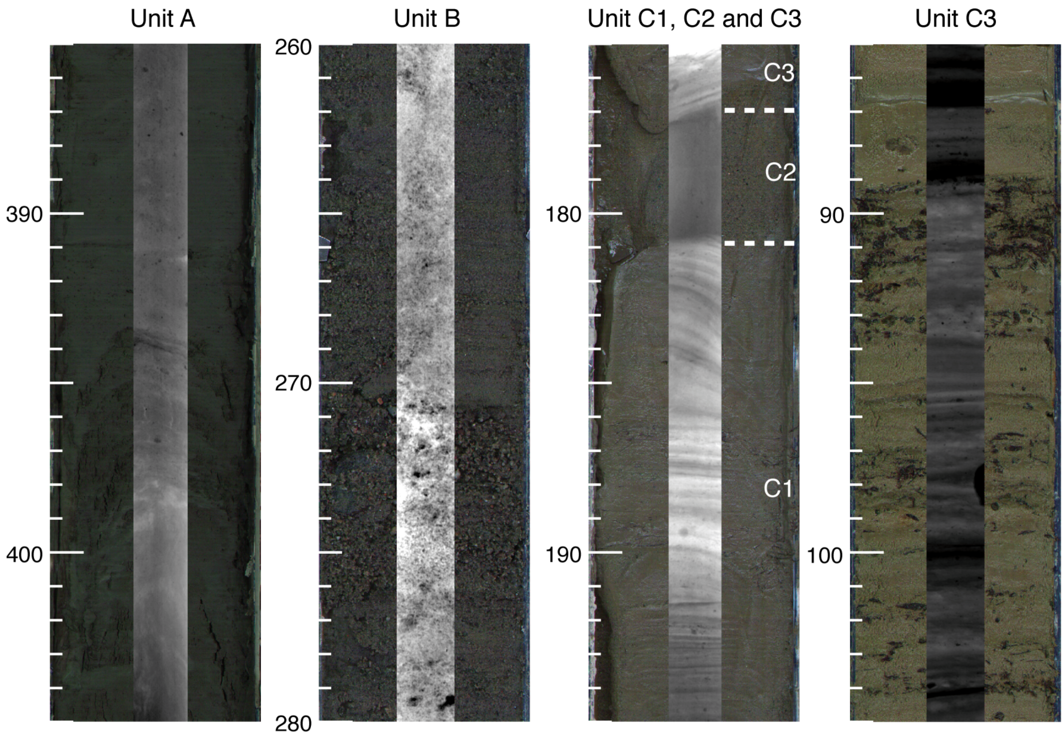

Fig. 9.4: Sections of optical and radiographic images of units A, B, C1, C2, and C3; scale in cm below sediment surface. Units are according to Fig. 9.3. Optical and radiographic images were taken with an ITRAX XRF core scanner (Cox Analytical System, Sweden).

9.4.2 Radiocarbon dates and Chronology

Four sediment bulk samples and ten samples of aquatic moss remains were dated using AMS (Table 9.1, Fig. 9.5). ^{14}C dating of organic matter from lake sediments is considered to give reliable ages when dissolved carbon in the water column is in equilibrium with the atmosphere (Björck & Wohlfarth 2001, Wolfe *et al*., 2004). However, the ^{14}C exchange may be restricted due to permanent lake-ice coverage thus increasing the amount of older carbon in the lake water resulting in reservoir ages (Doran *et al.*, 1999, Wagner *et al.*, 2006, Wagner *et al.*, 2008, Klug *et al.*, 2009). Older carbon may also be contributed by glacial melt, which is a common phenomena in lakes from high latitudes (Björck and Wohlfarth, 2001). In addition rework of sediment that contains organic matter from previously living organisms may also account for older radiocarbon ages in bulk samples.

Of the four bulk sediment samples the humic acid fraction (HA) and the alkali residue (humic acid free fraction, HAF) were radiocarbon dated. Both fractions show large age differences and also reversals occur (Table 9.1). Age differences between the two fractions indicate the problem of bulk sample contamination (Grootes *et al.*, 2004) that might derive from rework of older sediments or in-situ migration of younger humic acids from overlying sediments. The reliability of HA and the HAF ages has been subject to numerous studies and controversial results have been discussed (Walker and Harkness, 1990, Olsson, 1991, Björck and Wohlfarth, 2001). Nevertheless, Olsson (1991) suggested the HA ages from fine-grained sediments to be more reliable.

Three of the four bulk sediment samples derive from unit A. The HA and HAF ages from 431-432 cm, 367-368 cm, and 331-332 cm depth (KIA34798, KIA34403 and KIA34797, respectively) scatter within the period from 42800 to 34200 cal. yr BP but large differences between individual ages and also reversals occur (Table 9.1). The latter may point to reworking of sediment and admixture of older organic material. However, the ^{14}C ages date close to the detection limit of the radiocarbon method and the counting limit of the activity of a sample. Consequently, the ages of unit A should be regarded as minimum ages and we argue that unit A was deposited sometime before 34 cal. kyr BP.

Sample KIA34402 originates from 201-202 cm depth in unit C1. The HA and HAF fraction from this sample date to 22830 - 23371 and 25888 - 27520 cal. yr BP, respectively (Table 9.1). The carbon content of the HAF, however, is only 0.4 mg and thus very sensitive to contaminations during sample preparation (Grootes *et al.*, 2004). The carbon content of the HA sample was sufficient for providing reliable ages and dates in a period when temperatures in central Greenland were about 20°C colder than today (Andersen *et al.*, 2004). Harsh conditions likely have also prevailed in coastal North-East Greenland.

However, it is known that microbial productivity occurs even under permanent lake-ice coverage where light availability is very limited (Hawes and Schwarz, 1999, Gordon *et al.*, 2000). Although biogenic production may occurred we cannot exclude that sample KIA34402 contains old carbon from redeposited organic matter or from melt water and that the HA age of c. 23 cal. kyr BP is inaccurate.

The ten samples of moss remains are all younger than 10 kyr BP and show decreasing ages with decreasing depth. The oldest moss sample (KIA-34796) is from 188-187 cm depth in unit C1 and reveals an ^{14}C age of 9706 ± 49 yr BP, corresponding with 11235 – 10809 cal. yr BP (Table 9.1). This age suggests that the biogenic production in Melles Lake started shortly after the beginning of the Holocene. The topmost moss sample from 8-9 cm depth (KIA-34397) dates to 332 - 509 cal. yr BP and seems unaffected by ^{14}C from atomic bomb testing. The age of the topmost sample indicates that the aquatic mosses are likely not affected by a reservoir effect, thus providing reliable ages. A more detailed description of the moss remain dates and the chronology of the uppermost 190 cm of the Melles Lake record is provided by Klug *et al.* (in press).

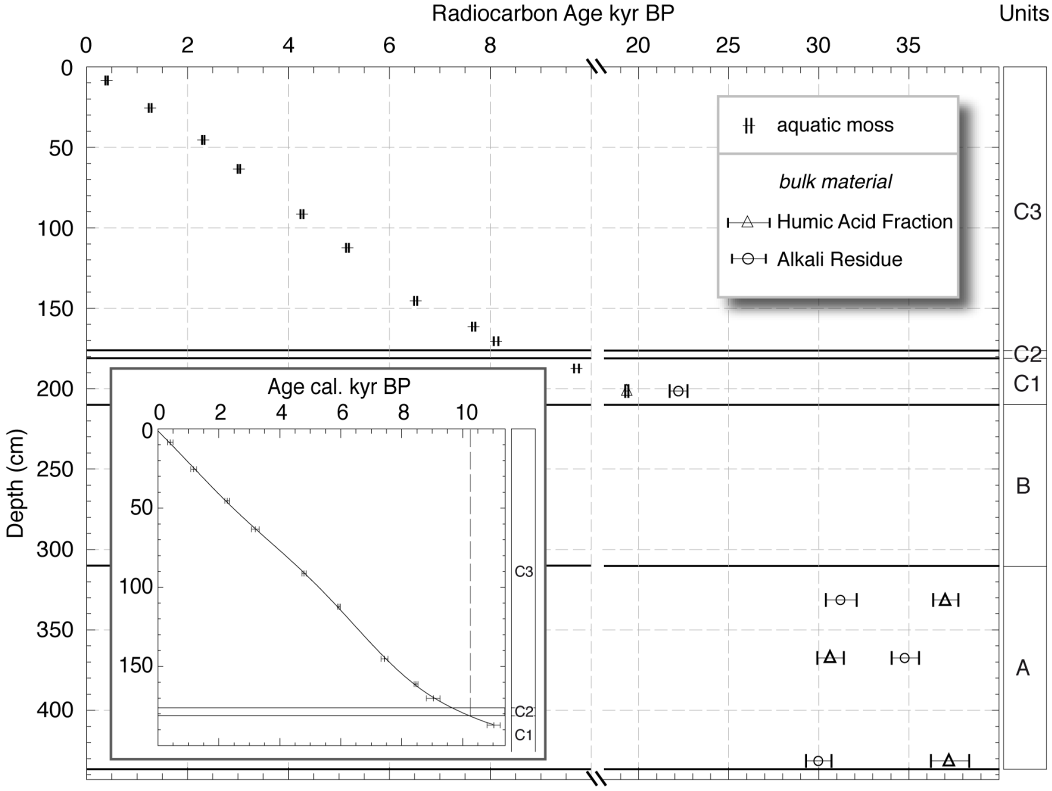

Fig. 9.5: Radiocarbon dates of Melles Lake plotted versus depth. Ages of bulk organi carbon were measured in humic acid fraction (HAF) and alkali residue (HAF). For sample weight, carbon content, and δ ^{13}C ‰ values see Table 9.1. The left insert shows an age-depth model of calibrated dates accorting to the calibrated ages in Table 9.1 and Klug et al. (in press).

9.4.3 Sediment Properties

Unit A. - The mainly fine-grained sediment in unit A (Fig. 9.6), combined with the weak lamination suggests a deposition under more or less calm conditions. Evidences for subaerial conditions as structural traces and for desiccation of the basin as evaporites (Raab et al., 2003) were not found. More likely, unit A was deposited in an aquatic setting. Since fine-grained and laminated sediments can occur as lacustrine bottomsets in a proglacial environment (Brodzikowski and van Loon, 1991), melt-water from a glacier in the catchment of Melles Lake may have provided suspended sediment.

The elemental profiles of unit A show high levels of K, Ti, Mn and Fe with some minor changes (Fig. 9.6). K and Fe are common elements in feldspars and micas, respectively, as they are common in the gneissic bedrock on Store Koldewey and in the Dove Bugt embayment. Ti can be used as an indicator of terrigenous sediment supply (Rothwell and Rack, 2006), whereas Mn can indicate manganese oxides that are more frequent in marine sediments than in terrestrial material (Revenko, 2002). High levels of K, Ti, and Fe therefore indicate a terrestrial provenance of the sediment, whereas the relatively high Mn concentrations point to a marine origin. High levels of Mn:Ti ratio may indicate post-depositional alteration (Reynolds *et al.*, 2004, Thomson *et al.*, 2006). The MS shows only slight variations throughout the unit and slightly decreases topwards. The low water content of c. 20% is at least partly a result of the sediment compaction and of the small amount of organic material (partly around 0.5 % TOC). The latter is specific to low productive environments, but can also derive from dilution due to high sedimentation rates.

Fine-grained sediments in the basal parts of lacustrine records are frequent in East (Björck *et al.*, 1994a) and North-East Greenland (Björck and Persson, 1981, Björck *et al.*, 1994b, Wagner *et al.*, 2000, Wagner and Melles, 2002). Comparable sediments at the base of lacustrine records from lakes on Store Koldewey are reported by Wagner *et al.* (2008) and Klug *et al.* (2009). Wagner *et al.* (2008) suggested glacial meltwater supply rather than aeolian of fluvial supply as origin for the occurrence of fine-grained clastic matter. Klug *et al.* (2009) concluded that the relevant sediment was most likely deposited in an aquatic milieu such as occurring in a proglacial lake. However, AMS dating of these basal sediments of the latter record revealed an infinite age. Furthermore the transition to the overlying unit is sharp and therefore erosion and loss of sediment is proposed (Klug *et al.*, 2009). The gradual shift in the grain size, the elemental profiles and the MS in Melles Lake to the overlying unit (Fig. 9.6) suggests that the transition to the overlying unit was non-erosional and that there is no hiatus between the units. Moreover, the radiocarbon dates from unit A suggest deposition sometime before 34 cal. kyr BP, which is considerably older than the basal sediments in

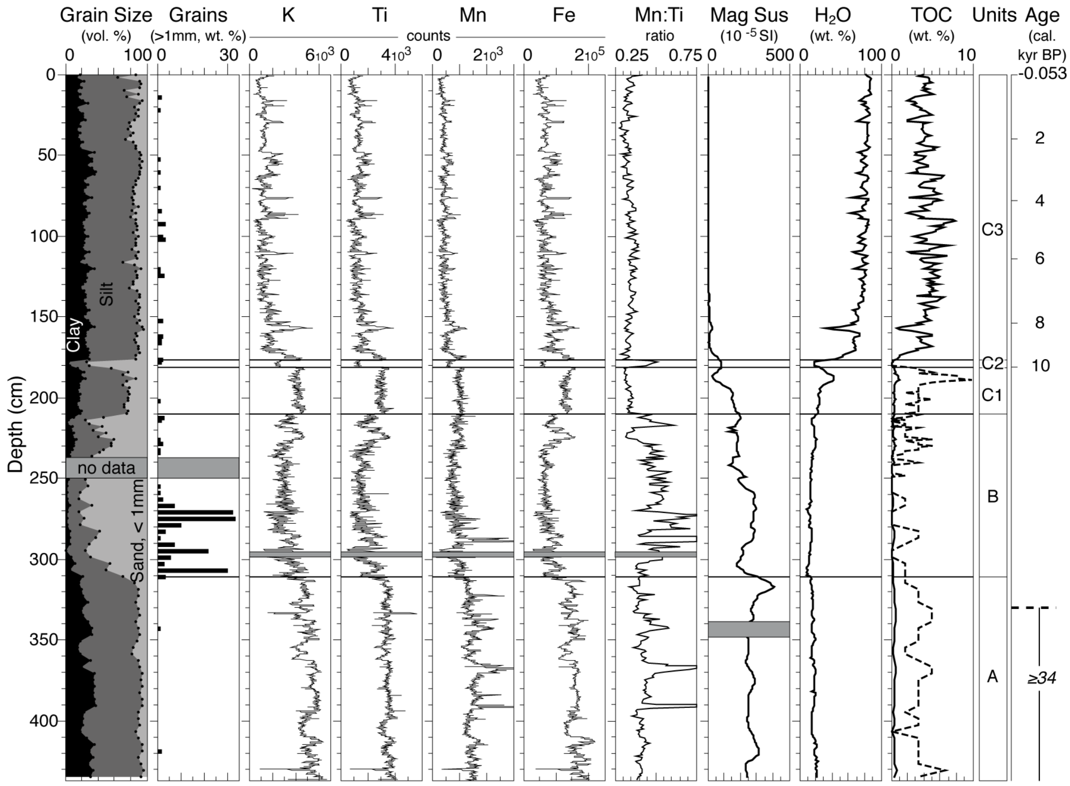

Fig. 9.6: Grain size data, XRF elemental profiles, geophysical, and biogeochemical data of Melles Lake and stratigraphic units versus depth (left) and and ages in cal. kyr BP (right) according to the age-depth model shown in Fig. 9.5. Mn:Ti ratio is averaged over 1 cm, for better resolution TOC values of units A, B, and C1 are exaggerated by factor 10 (dashed line). Grey bars indicate sections were no data are available.

other lakes in East and North-East Greenland (Björck and Persson, 1981, Björck *et al.*, 1994a, Björck *et al.*, 1994b, Wagner *et al.*, 2000, Wagner and Melles, 2002, Wagner *et al.*, 2008) and clearly pre-dates the late Weichselian glaciation.

Unit B. - The alternating sand and gravel combined with less fine-grained clastic matter and the occurrence of larger rocks imply a significant shift in depositional conditions in the Melles Lake basin. The upward coarsening that started already at the upper part of unit A and continues in the lower section of unit B, thus indicating a gradual transition. The elemental profiles of K, Ti, Mn, and Fe simultaneously decline in the lower part of unit B (Fig. 9.6) and vary on a relatively low level. Small-scale changes of the elemental counts could be due to variations in the sediment source, but could also be related to grain-size changes. MS alternates marginally and a general decreasing trend continues towards the upper boundary of unit B. The water content is lowest at 270 cm depth, but slightly increases topwards, which can likely be related to less compaction, decreasing grain sizes composition and slightly increasing TOC contents. Overall, TOC remains however, on a very low level throughout unit B, thus indicating a non-productive environment.

The deposition of coarser sediments in unit B requires increased transport energy. Fluvial transport is unlikely due to the poor sorting and the occurrence of subangular rocks embedded in a relatively fine matrix. These sediments are more characteristic for glacial transport, although glacial sedimentation depends on type of sediment transported, distance to the ice margin, or locality where the sediments are deposited (Ehlers, 1996). The subangular rocks indicate rather short glacial transport distance. The upward fining in the upper part of unit B could thus be due to glacier retreat and increasing transport distances. Therefore it is likely that unit B was deposited during a glacial advance with subsequent retreat. The lack of erosional features and the moderate compaction suggest that the coring position in the eastern part of the Melles Lake basin was not overridden by a glacier. This would also imply that the area east of the lake most likely remained ice-free.

Unit C1. - The transition from unit B to C1 is comparatively sharp. Grain-size characteristics of unit C1 differ markedly from the underlying unit (Fig. 9.6). The accumulation of fine-grained clastic matter with a clear lamination points to deposition under calm conditions, as they can occur in lacustrine environments. Assuming that the topmost part of the underlying unit was deposited by a retreating glacier, this glacier also might have supplied the fine-grained material in the base of unit C1. A slight increase in the elemental profiles of K, Ti, Mn, and Fe to values similar of those in unit A (Fig. 9.6) can partly be explained by decreasing grain sizes, but also may hint to a similar provenance (Fig. 9.6). The Mn:Ti ratio decreases to a comparatively low level. The decrease of MS throughout unit C1 along with slightly increasing values of Fe and Ti could also derive from enhanced formation of hematite (Hunt *et al.*, 1995). The occurrence of moss remains at 190 cm depth marks the onset of biogenic productivity owing to amelioration of living conditions (Klug *et al.*, in press).

Unit C2. - The sandy layer of unit C2 interrupts the sedimentation of fine-grained clastic matter. The elemental profile characteristics correspond with those from unit B and suggest a similar sediment source. The water content is again low and organic matter is not present in the almost homogenous layer (Fig. 9.6).

Sandy layers within fine-grained lacustrine sediments are known from many other lake sediment records from East and North-East Greenland. These layers are formed due to mass movement processes (e.g. Wagner *et al.*, 2000), when the low elevated basin became isolated from the marine environment or when an receding ice margin supplied large amounts of clastic material (e.g. Björck *et al.*, 1994a, Björck *et al.*, 1994b). Since the transitions of unit C2 to the under- and overlying units are very sharp, a lower lake level with the change to littoral facies or sediment slumping from the gently inclined slopes around the coring position is unlikely. More likely is that the coarser sediments were deposited during a period of rapid melting in the catchment.

Unit C3. - The overall fine-grained, partly layered to laminated sediments suggest accumulation under calm conditions such as in a lacustrine environment comparable to the present. The elemental concentrations decrease to lowest levels (Fig. 9.6), similar to those of units B and C2. This could indicate a similar sediment source, but low levels in unit C3 could be biased by increased water and TOC content. The Mn:Ti ratio remains on a low level throughout the entire unit C. Loss of manganese due to solution under anoxic conditions is in the shallow aquatic environment with a mixed water body and a likely fair oxygen supply probably not a matter (Cornwell and Kipphut, 1992, Burdige, 1993). The relatively constant values indicate an absence of post-depositional alteration (Reynolds *et al.*, 2004). The occurrence of moss layers in varying thicknesses indicates that lacustrine conditions similar to those of today prevailed. Variations in the frequency and the thickness of the moss layers can likely be attributed to climate changes and variations in the sea-ice coverage (Klug et al. in press). The most pronounced shift in elemental profiles, water content and TOC content at a depth of c. 158 cm depth can likely be associated with the 8.2 cold event (Klug *et al.*, in press).

9.5 The palaeoenvironment of Melles Lake basin during the late Quaternary

The sedimentary record of Melles Lake comprises five units, which represent a succession of palaeoenvironmental phases. Figure 9.7 provides potential scenarios for the development of the basin and its surrounding during these phases.

9.5.1 Phase 1 - pre-Last Glacial Maximum (early Glacial?)

The almost homogenous and fine-grained sediments of unit A with a faint lamination can be best explained by deposition in a subaquatic environment. The elemental profiles indicate sediment provenance from the gneissic bedrock in the vicinity of Dove Bugt, probably with some admixture of marine sediments as revealed by relatively high Mn contents. Melles Lake basin is located about 110 m above the local marine limit (Hjort, 1981) and pre-Quaternary marine sediments were not found in its catchment. Therefore, marine sediments could have been supplied by the transgression of the Greenland Ice Sheet (GIS) over Dove Bugt in the west of Store Koldewey (Fig. 9.7), thus incorporating shallow marine and coastal sediments being likely enriched in manganese (Calvert and Pedersen, 1996, Force *et al.*, 2008). The radiocarbon dates from unit A have to be regarded critically, since they are close

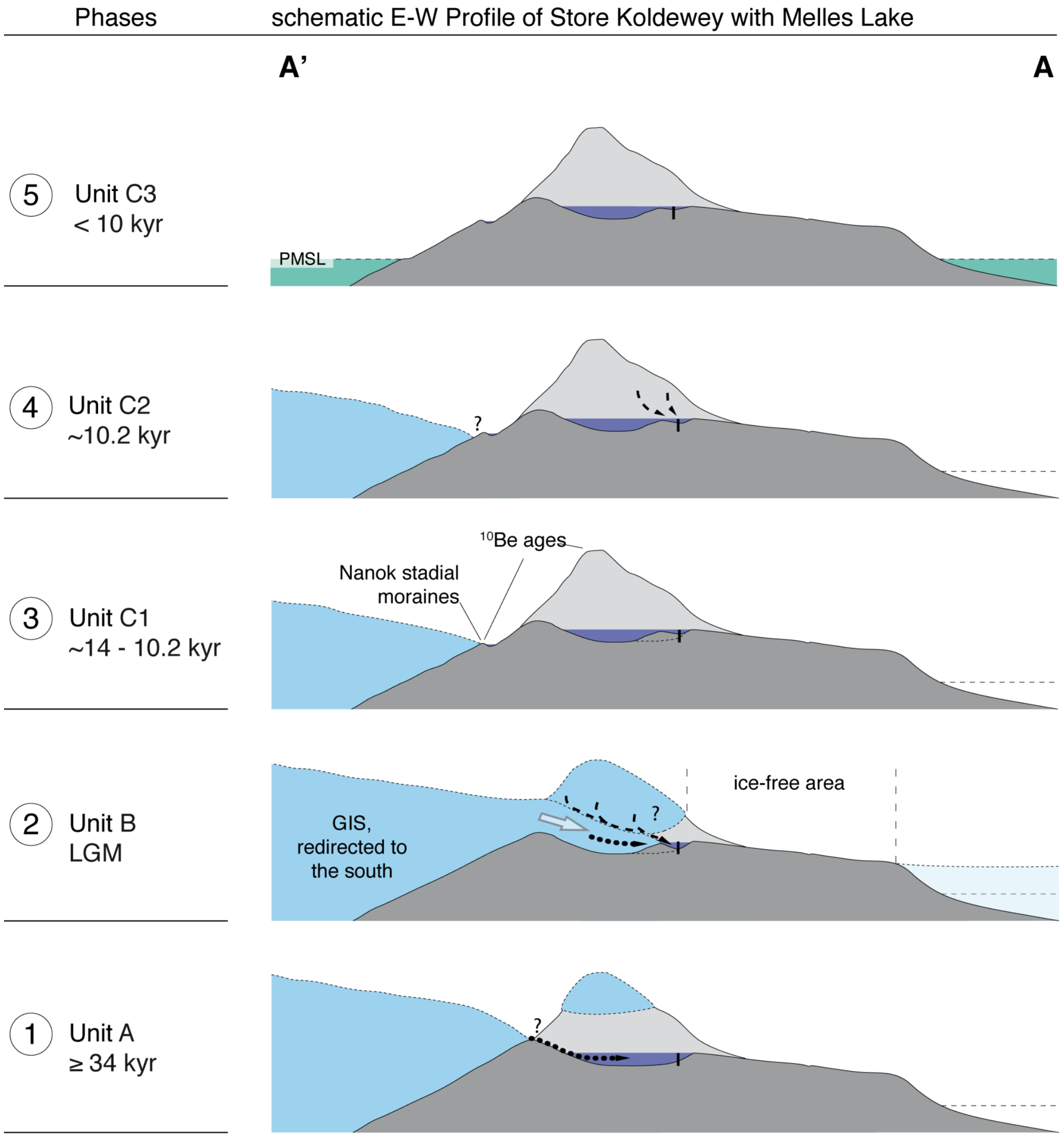

Fig. 9.7: Sketch of potential scenario of Melles Lake development during the late Pleistocene and Holocene. Profile (not to scale) through Melles Lake basin and view direction as indicated in Fig. 9.1. Arrows indicate distinct sediment delivery and different provenance with: dotted line = fine-grained, mainly marine material, dashed line = coarse sediments of local provenance. Ages of units are estimations according to calibrated ages in Table 9.1. The black bar indicates the coring position. Local glaciations on top of the mountain range are according to Håkansson et al. 2007.

to the dating limits and apparently comprise at least one reversal. Therefore, the ages of between 43 and 34 cal. kyr BP have to be regarded as minimum ages and indicate that unit A was most likely deposited prior to the LGM.

To provide the supposed marine sediment over the depression west of Melles Lake basin a minimum altitude of about 200 m above present sea level of a glacier transgressing Dove Bugt is needed but moraines of a pre-LGM glacial advance were not found on western Store

Koldewey. However, on Germania Land marine shell fragments occur in proglacial lacustrine sediments at 160 m a.s.l. and are associated with a glacial advance prior to the LGM (Landvik, 1994). Furthermore, the altitude of moraines in maximum position on Germania Land is above 200 m a.s.l. (Landvik, 1994). At Kap Herschell, central North-East Greenland glaciolacustrine sediments containing marine material with an age younger than 43 kyr BP have been found at 120 m a.s.l. (Houmark-Nielsen *et al.*, 1994).

9.5.2 Phase 2 - Last Glacial Maximum

The predominant fine-grained sediments of unit A gradually fade to the coarser sediments of unit B. The concomitant decline in the XRF elemental profiles is likely due to the grain-size changes, but can at least partly also be attributed to a change in sediment provenance. Less manganese in unit B may consequently indicate less input of marine sediments from Dove Bugt. The change in sediment source along with coarser sedimentation can be best explained by a glacial advance that pushed into Melles Lake basin. This advance likely picked up rocks from the western part of Store Koldewey and from the blockfields in the surrounding of the lake basin (Fig. 9.7). According to Hjort (1981), Björck and Hjort (1984), Funder (1989) and Landvik (1994) the western margin of Store Koldewey may have served as a barrier for the GIS and redirected the advancing ice towards the south. A small lateral lobe-like glacial tongue could have protruded into the western part of Melles Lake basin, thereby forming the subaquatic threshold that separates the western from the eastern basin (Figs. 9.2 and 9.7). Since there is no indication for glacial overriding at the coring location in the eastern basin, the lateral glacier tongue likely had only minor extent.

The cosmogenic exposure ages from the top of the mountain range on Store Koldewey imply that also the higher elevated areas were glaciated during the late Pleistocene (Håkansson *et al.*, 2007). A growth of local ice caps on Store Koldewey contemporaneous to the GIS advance during the LGM is likely and could also have caused the deposition of the subangular rocks and the changes in elemental profiles. However, since either a local cold-based plateau ice cap or an extensive polythermal ice sheet advance may have covered the Store Koldewey plateau (Håkansson *et al.*, 2007) the impact on the depositional environment in Melles Lake basin is inconclusive.

The sedimentary characteristics of unit B, the lack of evidences for erosion at the transition to the underlying unit, and the moderate compaction of underlying sediments suggest that the Melles Lake basin was not overridden by a glacier. More likely is that unit B was deposited in a very proximate proglacial, glaciolacustrine environment. Although no radiocarbon dates exist from unit B, it is likely that this unit was deposited in conjunction with the glacial advance during the LGM.

9.5.3 Phase 3 - Late Glacial

The fine-grained sediments of unit C1 suggest prevailing calm conditions as found in a proglacial lacustrine environment that is fed by glacial melt. Additionally, the distinct and fine lamination sedimentation likely indicates repeated periods of stagnant water due to permanent or periodic lake-ice coverage. The changes of the XRF elemental profile balance between values of unit A and unit B, which probably indicate a mixture of sediment from the glacier and from the catchment. The AMS dating of bulk sediment from the depth of 201-202 cm revealed an HA age of ca. 23 cal. kyr BP. Since this unit most likely is composed of glacially transported material and the lamination suggest calm condition as occur in permanent ice-covered lakes this age may also be biased by redeposited organic matter or by substantial reservoir age and could therefore be to old. Due to the lack of other datable material for the period shortly after deglaciation extrapolated sedimentation rates may give chronological constraints for the timing of glacier decay. The uniform lamination of unit C1 continues, except for the coarser sediments of unit C2, into unit C3. After subtracting 4 cm for unit C2 the calculated sedimentation rate between samples KIA-34796 at 188-187 cm depth and KIA-31912 at 171-170 cm depth amounts to about 6.5 cm/kyr. Extrapolating this sedimentation rate to the onset of the uniform sedimentation at 212 cm depth an age of about 14.7 cal. kyr BP is obtained. This estimated age coincides with the ^{10}Be age of about 14 kyr BP for ice-free conditions from the mountain plateau of Store Koldewey and another ^{10}Be age of 13 kyr BP from a perched boulder found on the Nanok stadial moraine at the western coast of Store Koldewey (Håkansson et al., 2007).

Melles Lake basin is located east of the depression (Figs. 9.2 and 9.7) that separates the lake basin from the lower elevated Nanok stadial moraines. It can therefore be assumed that even if an ice remnant remained in the western basin there was no substantial delay in the basin deglaciation compared to the ice-free conditions found on the surrounding mountains and the formation the moraine at the west coast of Store Koldewey. We therefore propose that the western lake basin was also deglaciated at about 14 cal. kyr BP or shortly thereafter.

9.5.4 Phase 4 - Late Glacial - Holocene transition

With the increasing temperatures after the Younger Dryas living conditions in Melles Lake ameliorated and aquatic mosses started to grow. The occurrence of mosses at 11.2 cal. kyr BP give the most unequivocal indication for ice-free conditions at the beginning of the Holocene. Ages for the deglaciation from other lacustrine records of North-East Greenland’s coastal area widely scatter (Cremer *et al.*, 2008) but no reliable ages for ice-free conditions prior to the Pleistocene-Holocene transition are reported. The oldest ^{14}C ages indicating ice-free conditions on Hochstetter Foreland from Peters Bugt Sø (ca. 13 kyr ^{14}C BP) and Ailsa Sø (ca. 14 ^{14}C kyr BP) (Björck and Persson, 1981) have been regarded as erroneous and corrected for Peter Bugt Sø to 9 kyr ^{14}C BP (Björck *et al.*, 1994b). Dating of organic matter of two other lacustrine records from central Store Koldewey (Wagner *et al.*, 2008, Klug *et al.*, 2009) indicate ice-free conditions at least at about 9.5 cal. kyr BP. After the colonisation of Melles Lake by aquatic mosses the laminated sedimentation was abruptly replaced by a layer of sand. According to the age model proposed by Klug et al. (in press) this layer was deposited at about 10.2 kyr cal. BP (q.v. Fig. 9.5). Although the temperatures at the beginning of the Holocene generally increased short periods with stagnant temperatures and also short inversions occurred (e.g. Andersen *et al.*, 2004, Vinther *et al.*, 2008). If a glacial advance, as it occurred on Hochstetter Foreland during the Nanok II stadial (Hjort and Björck, 1984) and in the Scoresby Sund area during the Milne Land stade (Funder, 1989, Funder *et al.*, 1994, Funder and Hansen, 1996), provided the coarser sediments of unit C2 cannot be determined.

9.5.5 Phase 5 – early Holocene to present (<10 cal. kyr BP)

Subsequent to the short period of perturbation at about 10.2 cal kyr BP the fine-grained sedimentation continued and, as indicated by increasing TOC values, biogenic productivity slowly increased. Aquatic mosses reoccurred in the sediment at about 9.5 cal. kyr BP (Klug *et al.*, in press). The lack of coarse sediments and the change to a biogenic dominated sedimentation indicate a change to a lacustrine setting comparable to present conditions (Fig. 9.7). Short periods when biogenic productivity almost ceased and the aquatic mosses are less frequent or disappeared are interpreted as biological adjustment to climatic deteriorations and lake-ice coverage in relation to severe sea-ice conditions of the East Greenland current off Store Koldewey (Klug *et al.*, in press).

Evidences for a glacial re-advance in conjunction with the neoglaciation from 5 kyr BP and onwards are not documented in the lacustrine sequence. At present, there are no glaciers or glacial remnants in the catchment of Melles Lake basin and only snowfields persisted during summer 2003.

9.6 Conclusions

Based on grain size changes, µXRF elemental profiles and geophysical and geochemical characteristics of Melles Lake record in concert with chronological constraints the following conclusions can be drown.

The five distinct sedimentary units of the lacustrine record cover at least the last 30 kyr and document changes of the local ice front. Fine-grained laminated sediments with a comparable high content of manganese suggest that a transgression of the Greenland Ice Sheet filled Dove Bugt and marine sediments were eroded and transported into the lake basin before the advance of the Greenland Ice sheet at the end of the Weichselian. With proceeding growth of the Greenland Ice Sheet in conjunction with the Last Glacial Maximum the western mountain range of Store Koldewey acted as a barrier and redirected the ice advance to the south. Only a narrow valley cutting through the mountain range allowed a restricted glacial lobe from the major ice in Dove Bugt to laterally protrude and inundate the western part of the lake basin. Following the LGM ice advance sedimentation in the lake was characterised by deposition of fine-grained sediments that indicate a proximal glacier margin. However, the continuous, gradual transition between and the weak compaction of the sedimentary units suggest that the coring position was not overridden by the glacier and the area east of the lake remained consequently ice-free, which can also be assumed for the adjacent low elevated areas to the north and south. With continuing ice decay after the LGM the lake became ice-free at about 14 kyr BP and deposition of fine-grained, laminated clastic sediments prevailed. After the onset of biogenic production at the beginning of the Holocene rapid melting in the catchment area at about 10.2 cal. kyr BP caused deposition of a sandy layer. It remains open, whether this short event is in relation to the period of re-advancing ice during the Nanok II stadial or due to some other climatic perturbation at the beginning of the Holocene. Following this episode, the Holocene history of the lake was characterised by varying biogenic productivity that primarily mirrors the regional and local climatic history. Evidences of advancing glaciers as occurred in conjunction with the neoglaciation or the Little Ice Age are not documented in the sediment sequence.

10 Synthesis

10.1 The variety of information from East and North-East Greenland's coastal lacustrine records - from local to supra-regional inferences

The studies presented here were aimed at the investigation of recent freshwater ecology of lakes and ponds in the high Arctic and of the detailed reconstruction of the climatic and environmental history during the late Quaternary. Study was carried out on two coastal islands off North-East Greenland. The studies address different time intervals at various temporal resolutions; particular attention was paid to the latitudinal differences of climatic and environmental changes during the late Quaternary. However, information obtained from these studies ranges from locally restricted to supra-regional inferences.

Local inferences. - Investigations of the recent hydrology and plankton of nine lakes and ponds on Store Koldewey (NE Greenland) revealed information about the freshwater and diatomic phytoplankton community characteristics of the investigated water bodies. The limnological survey included the recording of temperature, conductivity, oxygen concentration and saturation, pH, ionic composition, transparency, and the diatom phytoplankton community. In summer 2003 the lakes were cold, monomictic, thermally unstratified, alkaline and likely oligotrophic water bodies. The concentration of diatom phytoplankton, present in six lakes and dominated by four species (*Aulacoseira tethera*, *Cyclotella pseudostelligera*, *C. rossii*, and *Fragilaria tenera*), varied distinctly between the lakes. The freshwater ecosystems are controlled by the present climatic conditions on the coastal island. Furthermore, the lakes and ponds are also influenced by variations in environmental conditions like the duration of lake-ice coverage, the status of soil evolution, melting processes in the catchment and the sediment and nutrient input into the lake.

The sedimentary record from Duck Lake, central Store Koldewey, contains a succession from clastic to biogenic-clastic sediments that cover almost the entire Holocene. High compaction of the fine-grained basal sediments in combination with the erosional transition to the overlying unit suggests that the lake basin was covered by glacial ice. Subsequent coarse to fine-grained sedimentation can be attributed to the decay of the glacier and deposition in a proglacial environment. Information as to whether the glacier was of local origin or connected to the Greenland Ice Sheet, could however, not be obtained. Organic matter began to accumulate in the lake at 9.1 cal. kyr BP, which provides a minimum age for the deglaciation of the basin. Although the early to mid-Holocene is known as a thermal maximum in E Greenland (e.g. Wagner *et al.*, 2000, Christiansen *et al.*, 2002, Wagner and

Melles, 2002, Kaufman *et al.*, 2004), organic matter accumulation in the lake remained low during the early Holocene and increased during the middle and late Holocene (q.v. Fig. 7.7), when temperatures in East Greenland gradually decreased. Late plant immigration and lack of nutrient availability due to retarded soil evolution rather than temperature seem to have affected the biogenic productivity in the lake. Chironomids are abundant throughout the record after 9.1 cal. kyr BP (q.v. Fig. 7.7). Besides their sensitivity for temperature changes the chironomid assemblages from central Store Koldewey are largely controlled by changes in nutrient availability (pers. comm. Steffi Schmidt). Nevertheless, chironomids may be a suitable tool for climatic reconstructions even in those high arctic environments. However, a better understanding of the ecology of chironomids under these extreme conditions and detailed investigation of palaeotrophic conditions of the lakes are needed. It is also concluded that relative temperature reconstructions based on biogeochemical data have to be regarded critically, particularly in the period shortly after deglaciation when nutrient availability was low.

The results of the limnological survey give information about recent conditions of the lakes on Store Koldewey (Fig. 10.1A). The results fill the gap of limnological studies in NE Greenland and add to the growing body of information about recent ecological conditions of freshwater lakes across the arctic (e.g.

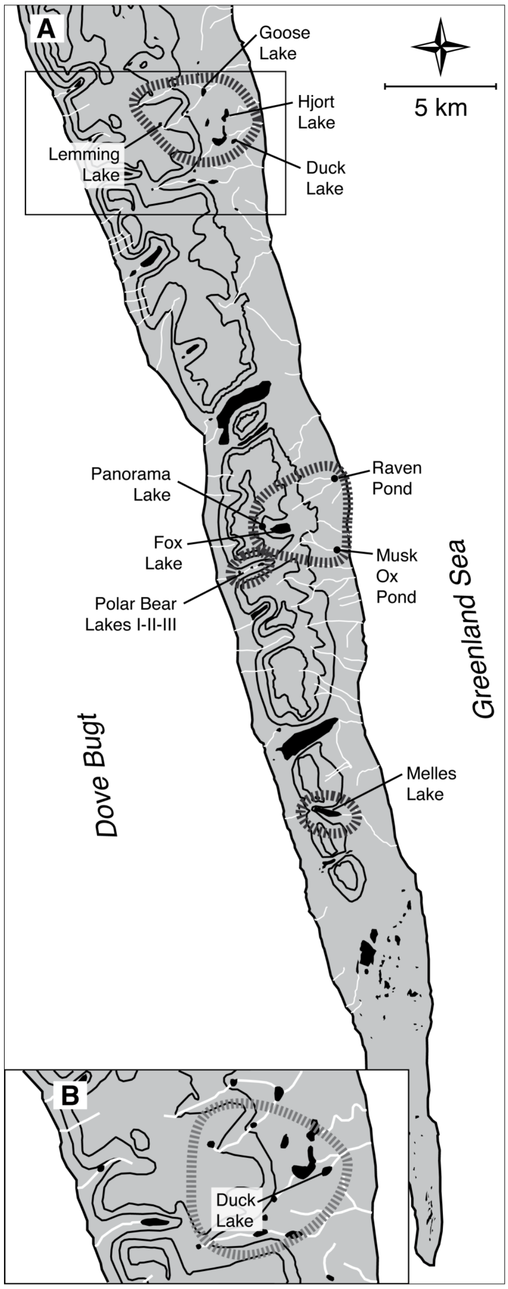

Fig. 10.1: A) Catchment areas (dark dashed encircling) that, besides climatic conditions, mainly influence the recent hydrology and diatom phytoplankton of lakes on central and southern Store Koldewey and B) Inferences drawn from Duck Lake's sedimentary record are locally restricted (gray dashed encircling) and mainly give information about the local environmental history.

Antoniades *et al.*, 2003, and references therein; Hobbie, 1984, Cremer and Wagner, 2003, Duff *et al.*, 1999, Willemse, 2002, and references therein). The interpretation of changes in sedimentary, geophysical, biogeochemical, and fossil data from Duck Lake records (q.v. Fig. 7.7) shows that the Holocene evolution of the lake was mainly influenced by local environmental changes on Store Koldewey and inferences drawn from this record are locally restricted (Fig. 10.1B). The analysis of data from Duck Lake corroborates the interpretation by Wagner *et al.* (2008) of restricted nutrient availability that hampered biogenic productivity and thus restraining a detailed climatic reconstruction. Inferences drawn from Duck Lake record that could explain latitudinal differences between the climatic history of East and North Greenland are therefore not possible with data from this record (Fig. 10.3).

Regional inferences. - The sedimentary record of Melles Lake, southern Store Koldewey, is composed of five distinct units that comprise a succession from glaciolacustrine to limnic setting. Results of XRF studies in combination with ^{14}C dating suggest that the basal sediments of the record are of marine origin with a minimum age of 34 cal. kyr BP. The origin and the age of the basal sediments imply that a pre-LGM transgression of the Greenland Ice Sheet covered Dove Bugt west of Store Koldewey and transported eroded marine sediments, which were most likely deposited in the lake basin during a subsequent melting. In the course of the advance of the Greenland Ice Sheet in association with the LGM the glacial ice was redirected to the south at the western shore of Store Koldewey and only a small lateral glacier lobe intruded into the western basin of Melles Lake. Sediment characteristics as moderate compaction and the lack of erosional features in combination with the presence of a subaquatic moraine suggest an unglaciated eastern lake basin (q.v. Figs. 9.6 and 9.7). These features give the first evidence of ice-free conditions on eastern Store Koldewey during the Late Weichselian as suggested by Hjort (1979, 1981). With the subsequent glacier decay in the western part of the basin, the lake experienced deposition of fine-grained and laminated sediments. The estimated time of ice-decay of about 14 cal. kyr BP has to be regarded critically but coincides with ice-free conditions on higher plateaus of Store Koldewey (Håkansson *et al.*, 2007). If the estimated age mirrors ice-free conditions in Melles Lake the record contains sediments from the Late Weichselian-Holocene transition, thus predating the time of deglaciation in other lacustrine records from Store Koldewey by about 3 kyr (Wagner *et al.*, 2008, this study). However, clastic matter dominated the sedimentation and first macrofossils that indicate ameliorated living conditions in the lake are not found until about 11 cal. kyr BP. Following a short period of intensive melting in the catchment area at about 10.2 cal. kyr BP and related coarser sediments of local provenance the Holocene sediment sequence is dominated by fine-grained sediments and variable biogenic deposition. Glacial advances as occurred during the Nanok II stadial at about 11.2

cal. kyr BP (Hjort and Björck, 1984), corresponding to the Milne Land stade in E Greenland (Funder, 1989, Funder and Hansen, 1996) or later in association with the neoglaciation from about 5 kyr onwards (Funder, 1989, Andrews *et al.*, 1997, Jennings *et al.*, 2002, Reeh, 2004) are not documented in the record.

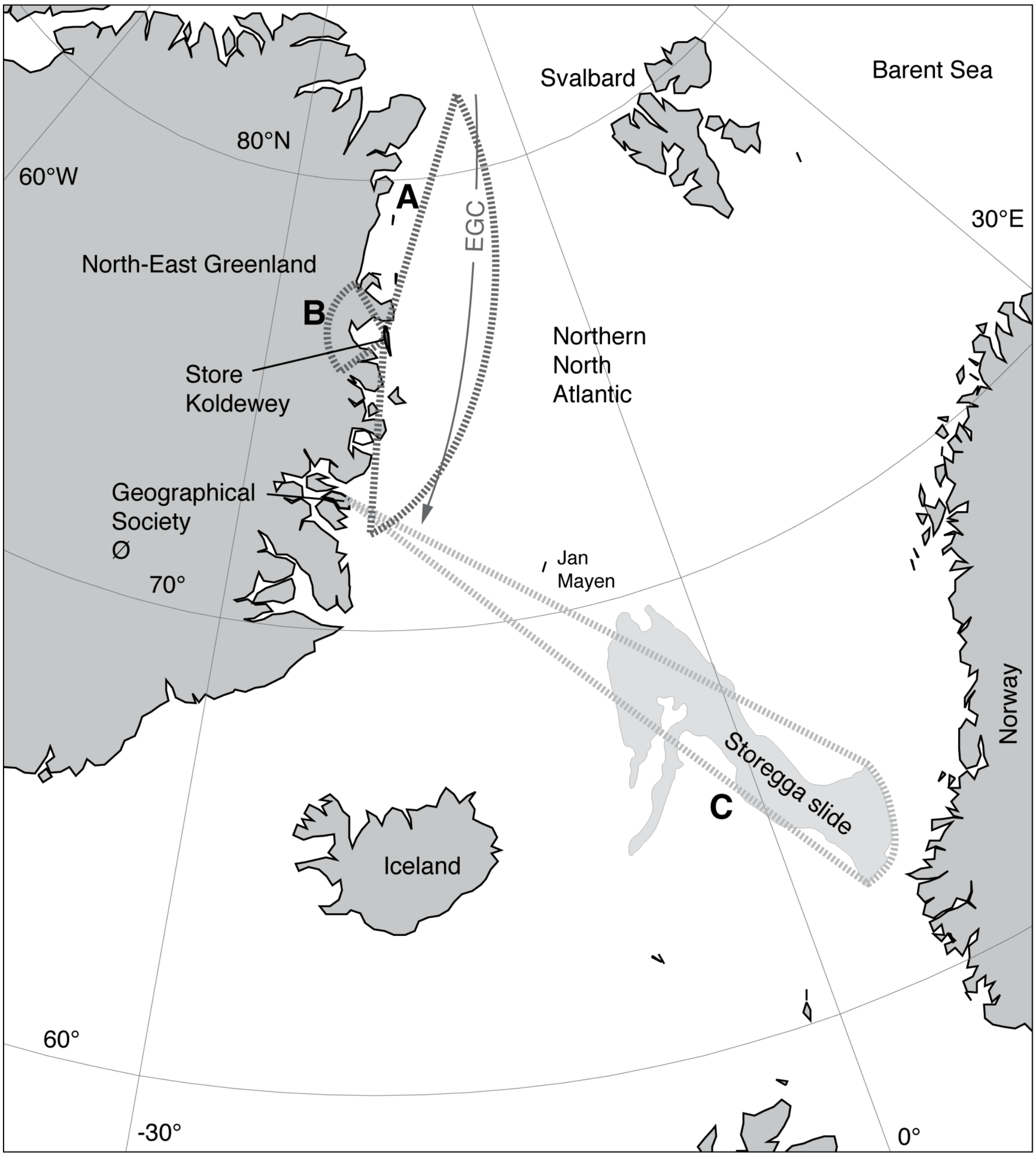

Fig. 10.2: From regional to supra-regional inferences: areas or events with impact on records presented in this thesis: A) regional inferences drawn from the clastic record of Melles Lake with information about the spatio-temporal changes of the Greenland Ice Sheet during the Late Weichselian; B) regional inferences drawn from the biogenic productivity in Melles Lake that is affected by the variability of sea-ice coverage of the East Greenland Current throughout the Holocene; C) supra-regional inferences drawn from the clastic sediment sequence of Loon Lake record, Geographical Society Ø, with the first deposits of the Storegga tsunami in East Greenland.

After the period of about 3 kyr with ice-free conditions and a dominant fine-grained clastic sedimentation, living conditions ameliorated and aquatic mosses colonised Melles Lake. Dating of the oldest mosses shows that lacustrine biogenic productivity began as early as around 11 cal. kyr BP. This age predates the onset of bioproductivity in Hjort Lake (Wagner *et al.*, 2008) and Duck Lake (this study) on central Store Koldewey by about 2 kyr and in other lakes on coastal E Greenland by at least 0.6 kyr (Cremer *et al.*, 2008). In spite of the early onset of biogenic production organic matter accumulation remained low and minerogenic sedimentation continued. At about 9.5 kyr cal. BP moss, sulphur, organic carbon, and biogenic silica content started to increase, indicating that the environment stabilized and the biogenic production in the lake adjusted to more preferable conditions. Subsequently, the bioproductivity experienced repeated changes and varied on long-term and short-term scales. The long-term trend shows a maximum during the Early Holocene (q.v. Fig. 8.5) thus responding to increased temperatures during the Holocene Thermal Maximum as recorded in various archives in E Greenland (Johnsen *et al.*, 1992b, Fredskild, 1995, Wagner *et al.*, 2000, Christiansen *et al.*, 2002, Wagner and Melles, 2002, Kaufman *et al.*, 2004). Superimposed on the long-term trend, bioproductivity also experienced repeated short-term fluctuations on multi-decadal to centennial scales that match partly the NGRIP temperatures (q.v. Fig. 8.5). The most pronounced decrease of bioproductivity occurred at around 8.2 cal. kyr BP. Perennial lake ice coverage due to low temperatures is supposed to have caused decreased lacustrine bioproductivity. From the middle Holocene to the present repeated decreases of bioproductivity occurred that could be related to periods with severe sea-ice conditions of the East Greenland Current (q.v. Fig. 8.5). Besides the dependence on air temperature it therefore demonstrates the sensitivity of lacustrine bioproductivity in coastal high arctic areas to short-term cold spells that are mediated by the currents emanating from the Arctic Ocean.

Inferences drawn from the clastic record of Melles Lake are regionally restricted to the spatio-temporal fluctuation of the Greenland Ice Sheet and its impact on the local glacier front environment (Fig. 10.2). The detailed climatic reconstruction following deglaciation during the Late Weichselian to the beginning of the Holocene was, however, hampered by the low biogenic productivity, most likely as a result of the still prevailing inhospitable low temperatures at the end of the Weichselian, the restricted availability of nutrients, and/or an assumed retarded plant migration.

The Holocene climatic evolution as shown by the long-term correspondence of biogenic productivity of Melles Lake with the Holocene Thermal Maximum in East Greenland indicates that the climate on Store Koldewey was in general controlled by climatic conditions that prevailed further to the south (Fig. 10.3). On a shorter time-scale, the atmospheric and/or

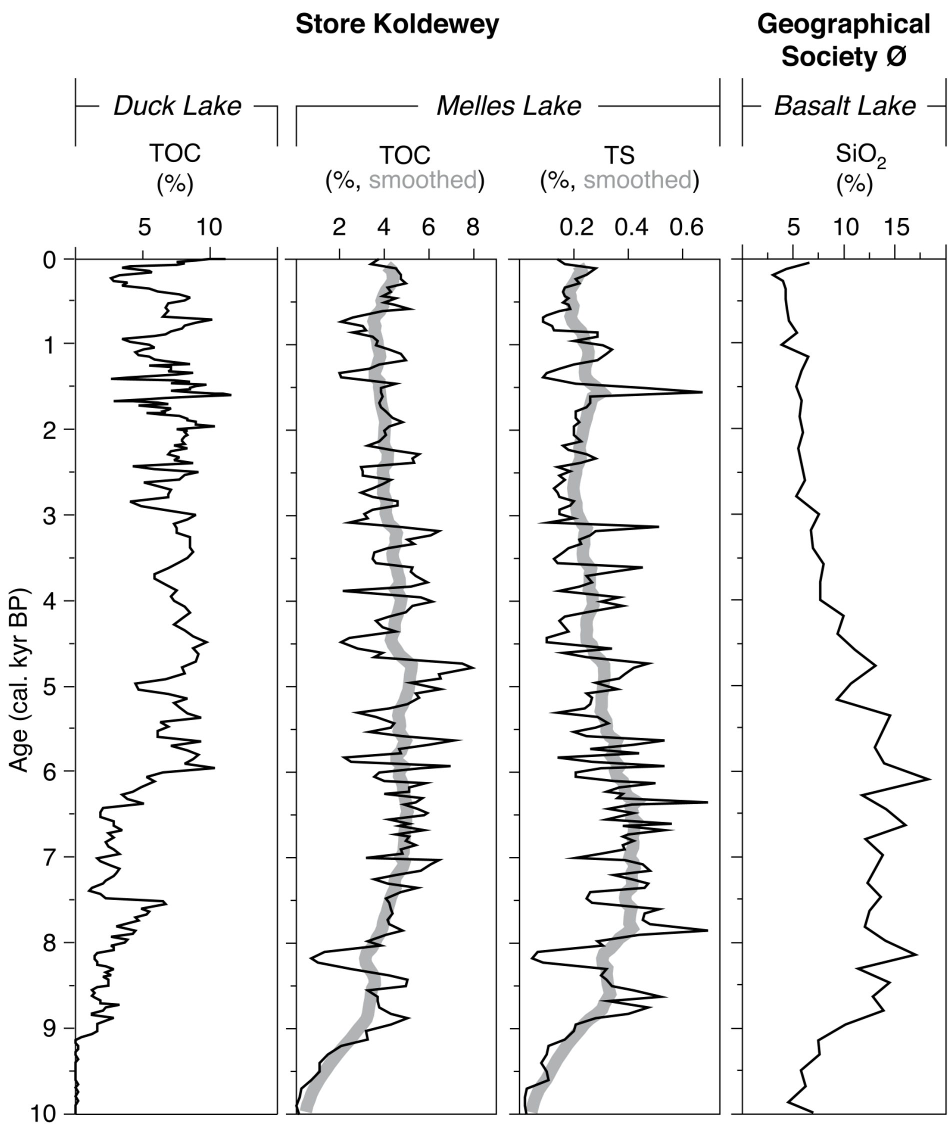

Fig. 10.3: Selected proxies from Duck Lake, Melles Lake (both Store Koldewey) and Basalt Lake (Geographical Society Ø) that show biogenic productivity related to environmental changes (Duck Lake, left column) or climatic changes (Melles Lake and Basalt Lake). The correspondence of bioproductivity in Melles and Basalt Lake to the Holocene Thermal Maximum indicate a similar climatic evolution in NE Greenland with those further to the South. Basalt Lake data are modified after Wagner and Melles (1999), published in Wagner et al. (2000).

oceanic circulation patterns, those related to the temperature over the Greenland ice sheet, or to the oceanic conditions off the coast of NE Greenland, influenced the biogenic productivity of Melles Lake. This correspondence indicates an impact of regional climatic conditions such as prevailing northerly winds, and low temperature due to reduced albedo, in relationship with increased sea-ice coverage of the EGC (Fig. 10.2). However, these short-term changes cannot be linked to the thermal maximum in N Greenland suggesting that either the impact of the Arctic Ocean was stronger in the north or that the lacustrine records of N Greenland are influenced by regional peculiarities like the lack of nutrients, as a result of retarded soil evolution or late plant emigration.

Supra-regional inferences. - The 2.73m long sediment sequence from Loon Lake, Geographical Society Ø, East Greenland, mainly consists of fine-grained homogeneous sediments interspersed with a 0.72m thick sandy horizon that shows distinct fluctuations in the grain-size distribution and in the physical and biogeochemical properties. The predominance of marine fossils throughout the entire sequence - dated to between 8630 and 7535 cal. yr BP - shows that the Loon Lake at that time was a marine basin. The sandy horizon was, according to the radiocarbon dates, deposited after 8500–8300 cal. yr BP (q.v. Fig. xx) and is interpreted as originating from the Storegga tsunami.

Besides the opportunity to reconstruct sea level changes and the local postglacial uplift history, this record demonstrates that lakes from low elevated areas in E and NE Greenland may contain deposits from the Storegga tsunami and show the long distance impact of tsunami waves (Fig. 10.2).

10.2 Future perspectives

Towards an understanding of the late Quaternary evolution of East Greenland and its surrounding area this thesis provides new and substantial findings about the late Weichselian and Holocene climatic and environmental history in North-East Greenland. Additionally, some questions need further investigations and some new issues arose, the clarification of which will contribute to our understanding of spatial and temporal climatic and environmental changes in the past.

Since proxies used for climatic and environmental reconstruction are controlled by local environmental factors the obtained proxy-related information may vary between lakes. In addition to the need for a multidisciplinary approach (Wagner, 2000) it is therefore suggested that, for a better understanding of information from lacustrine records, a denser sampling grid is preferable. Furthermore, investigation of more than one lake from one locality can help to distinguish between local and regional or even supra-regional signals. Once local environmental factors controlling biogenic productivity are understood and lacustrine bioproductivity is adapted to the prevailing environmental conditions, a detailed climatic reconstruction is achievable. To link the climatic and environmental history of East and North-East Greenland with those of North Greenland, supplementary information for latitudinal climatic and environmental peculiarities during the Holocene may be obtained from

Blåsø (ca. 79°38'N, 22°50'W), near Nioghalvfjerdsfjorden and by a multidisciplinary re-investigation of the lakes of N Greenland.

Lacustrine sediment sequences from low elevated areas in E and NE Greenland may contain sediments originating from the Storegga tsunami. However, more than one sediment sequence, preferably from different elevations, is needed to reconstruct the affected area in E and NE Greenland and to reconstruct runup heights of the tsunami waves and thus the impact of the tsunami, as done in coastal areas of Norway, Scotland and other North Atlantic regions.

To verify the sensitivity of lacustrine biogenic productivity to sea-ice coverage off coastal areas, lacustrine records from eastern or southern Svalbard could be promising and may present suitable targets for the reconstruction of sea-ice coverage changes of the western and south-western Barents Sea.

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12 Appendix

Sediment core data, including AMS dating results and fossil data, of Duck and Melles Lake are available at the geoscientific database PANGAEA: http://www.pangaea.de.

Erklärung

Ich versichere, dass ich die von mir vorgelegte Dissertation selbständig angefertigt, die benutzten Quellen und Hilfsmittel vollständig angegeben und die Stellen der Arbeit − einschließlich Tabellen, Karten und Abbildungen −, die anderen Werken im Wortlaut oder dem Sinn nach entnommen sind, in jedem Einzelfall als Entlehnung kenntlich gemacht habe; dass diese Dissertation noch keiner anderen Fakultät oder Universität zur Prüfung vorgelegen hat; dass sie − abgesehen von unten angegebenen Teilpublikationen − noch nicht veröffentlicht worden ist sowie, dass ich eine solche Veröffentlichung vor Abschluss des Promotionsverfahrens nicht vornehmen werde. Die Bestimmungen der Promotionsordnung sind mir bekannt. Die von mir vorgelegte Dissertation ist von Prof. Dr. Martin Melles betreut worden.

Teilpublikationen:

1) Cremer H, Bennike O, Håkansson L, Hultzsch N, Klug M, Kobabe S and Wagner B. 2005. Hydrology and diatom phytoplankton of high arctic lakes and ponds on Store Koldewey, Northeast Greenland. *International Review of Hydrobiology* **90**, 84-99.

2) Wagner B, Bennike O, Klug M and Cremer H. 2007. First indication of Storegga tsunami deposits from East Greenland. *Journal of Quaternary Science* **22**, 321-325.

3) Klug M, Schmidt S, Bennike O, Heiri O, Melles M and Wagner B. 2009. Lake sediments from Store Koldewey, Northeast Greenland, as archive of Late Pleistocene and Holocene climatic and environmental changes. *Boreas* **38**, 59-71.

4) Klug M, Bennike O and Wagner B. im Druck. Repeated short-term bioproductivity changes in a coastal lake on Store Koldewey, North-East Greenland, an indicator of varying sea-ice coverage? *Holocene*. im Druck

5) Klug M and Wagner B. Late Weichselian ice-front history on Store Koldewey, North-East Greenland, as illustrated by a lacustrine record. in Vorbereitung zur Veröffentlichung in Journal of Quaternary Science.

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