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Holocene Sedimentation History of the Shallow Kangerlussuaq Lakes, West Greenland

Nico W. Willemse

GEOSCIENCE • 41 • 2002

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Cover: Kangerlussuaq area showing lake SFL4 (right central area) surrounded by dwarf shrub heaths. In the background the Inland Ice, with the Russell outlet glacier (photo M. Marbus, July 1991).

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Contents

Abstract

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The sedimentation history and Holocene development of seven small and shallow closed-basin lakes in the continental Kangerlussuaq region of West Greenland was studied by detailed analysis of their sedimentary infill, chronology, and litho- and chronostratigraphy of sediment cores. The lakes are situated between the present-day Inland Ice margin and Sukkertoppen Iskappe and have generally simple basin morphometries. Accumulation and distribution of organic rich sediments indicate a rather regular lake infill since deglaciation. According to AMS ^{14}C ages of basal organic sediments, start of organic infill in the lakes follows the general timing for deglaciation as defined by regional moraine chronostratigraphy, but include evidence for the diachronous development of individual lakes. Holocene accumulation rates of organic-rich sediments were rather constant over time, and the longest record extends back to ca. 10,500 cal yrs B.P. Floristic shifts in the first diatom record presented from this area record significant limnological changes, with relatively mesotrophic conditions prevailing throughout lake history. Results of high-resolution loss-on-ignition (LOI) analysis on diatomaceous and silt-rich organic sediments suggest a dominant lacustrine signal in these lakes. A characteristic sedimentary sequence occurs in all lakes investigated where highly variable LOI fluctuations show a close correlation between sites, well corroborated by radiocarbon ages. The results are discussed in the context of on-going investigations of the palaeoenvironmental and climatic record in the continental interior of West Greenland and present information on the environmental background, basin characteristics, sediment stratigraphy, and chronostratigraphy of the shallow Kangerlussuaq lakes.

Key words: Lake sediments, Greenland, stratigraphy, Holocene, AMS ^{14}C, climate change, diatoms.

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Introduction

Hindcasting climatic trends in the polar region has become increasingly important following the recognition of a probable polar amplification to global warming (Kattenberg *et al., 1*996). High latitude feedback mechanisms play an important role in the climate system (Dickinson *et al., 1*996), but in the absence of long term meteorological records, natural climate variability and feedback mechanisms remain poorly understood (Overpeck *et al., 1*997). Arctic areas are often rich in lakes of all sizes and their sediments contain one of the few continuous archives for deciphering past climate variability (Smol *et al.,* 1996, Hughen *et al., 1*996; Bradley *et al.*, 1996). Recent efforts focussed on the high-resolution climate signal in annually laminated lake sediments of the eastern Canadian Arctic (Hughen *et al.*, 1996; Bradley *et al., 1*996; Lamoureux & Bradley, 1996; Hardy *et al.,* 1996). Apart from ice core data from the Greenland Ice Sheet (Johnsen *et al.,* 1995; White *et al., 1*997), no such detailed information is yet available from West Greenland sites. However, the marked geographical variability between palaeoclimate records and instrumental data from the eastern Canadian Arctic and Greenland point to the importance of increasing spatial coverage (Overpeck *et al., 1*997).

The present study aims at a more comprehensive understanding of the type of environmental signal archived in sediment records from the numerous small (0.5-2 ha) and shallow (3-8 m) lakes in the West Greenland Kangerlussuaq area (Fig. 1), following previous research on the postglacial sedimentary record (Dijkmans, 1988; 1989; Dijkmans & Törnqvist, 1991; Eisner *et al., 1*995). Palaeolimnological and palaeoecological approaches for several of the larger lakes are currently under study (Anderson & Bennike, 1997; Anderson *et al.,* 1999; Bennike, 2000; Anderson *et al*., 2000; Brodersen & Anderson, 2000; Brodersen & Anderson *in press*), but our efforts focus on shallow lakes with comparable basin morphometries, small drainage areas, and no distinct inlets or outlets. Controls on physical, chemical and biological processes for such lakes are in general relatively simple, and the response of small lake-watershed systems to short term (interannual) climate fluctuations is likely to be maximal (e.g. Smol *et al., 1*991; Douglas *et al., 1*994).

The majority of these lakes are unstudied, but previous work in the area showed the presence of a strong limnological and climatic signal (Eisner *et al., 1*995; Willemse & Törnqvist, 1999). Little precise information is yet available about the chronology, sedimentary infill and sediment stratigraphy of these shallow lakes. The purpose of this paper is to provide details on sediment and basin characteristics, sediment stratigraphy, environmental background, and the paleoenvironmental signal recorded in seven shallow lakes situated along a roughly 120-km-long transect, extending from the present-day Inland Ice margin towards Sukkertoppen Iskappe (Fig. 1). A large number of new acceller-ator mass spectrometry (AMS) ^{14}C ages from Kangerlussuaq lake sediment records are presented, providing the chronostratigraphic framework for reconstructing the environmental and climatic history.

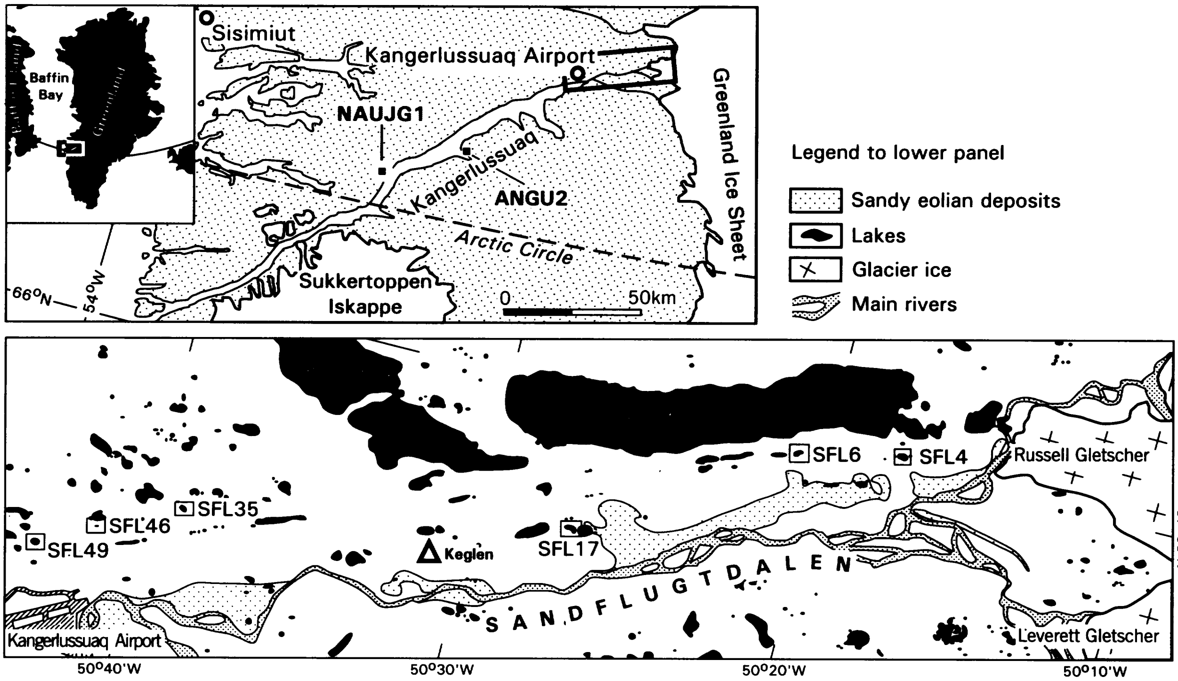


Fig. 1. Location of sampling sites in the Kangerlussuaq region of West Greenland.

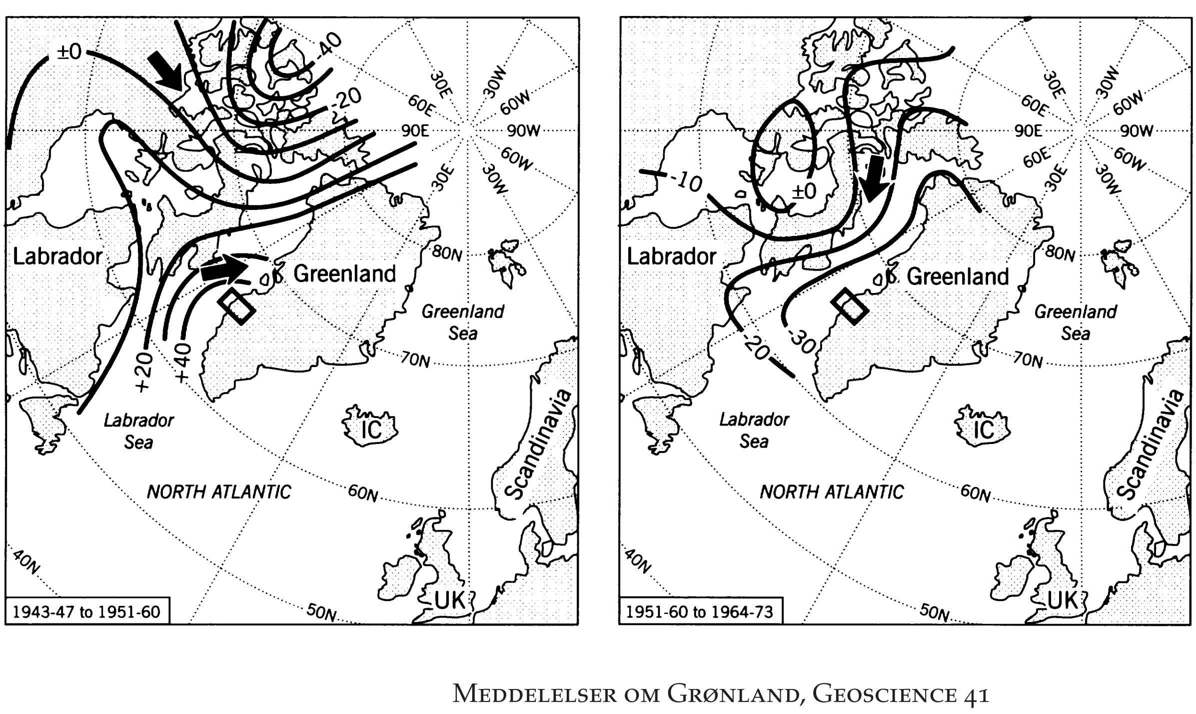
Climatic significance of West Greenland lakes

Climate variability and West Greenland

The Baffin Bay region (Baffin Bay, West Greenland, and Baffin Island) is excellently suited for studying Arctic climate variability. The present-day synoptic climatology is influenced by a major feature of northern Hemisphere circulation: a semi-permanent upper level trough in the circumpolar westerly flow over Baffin Bay (Jacobs *et al.*, 1985; Barlow *et al.*, 1997). The Baffin Bay trough largely reflects the northward diversion of depression tracks from the central Canadian Arctic due to blocking highs over Greenland and the orographic influence of the Greenland Ice Sheet (Putnins, 1970; Barry & Kiladis, 1982). Regional atmospheric circulation, and the associated pattern of temperature and precipitation in the Baffin Bay region are strongly affected by changes in the position of the trough axis (Fig. 2, also Keen, 1980; Williams & Bradley, 1985), which exhibits interannual variability related to the Icelandic Low and the North Atlantic Oscillation (Barlow *et al.*, 1997). Below or above normal air temperatures in Greenland are associated with an eastward/westward shift of the trough axis as a response to unusual stronger/weaker westerlies over the North Atlantic (Keen, 1980). In addition, oceanographic and sea-ice conditions play an important role in regional climatic differences (Jacobs *et al.*, 1985; Williams & Bradley, 1985). In contrast to most other regions of the Earth, the Baffin Bay region has experienced decreasing temperatures over recent decades.

Temperature trends (from Nuuk since 1784, Ilulissat 1840) document drastic interannual and decadal climate variability in the region over the last hundred years (Fig. 3). Briefly, these general

Fig. 2. Changes in the mean July 700 millibar heights between periods 1943-47 to 1951-60 and 1951-60 to 1964-73. From 1943 to 1960 the mean trough position moved westwards, resulting in increased southerly and westerly circulation with moist and warm maritime air over West Greenland and Baffin Island. From 1964-73 the mean trough position was more east, with enhanced northerly and northeasterly airflow (cold and dry arctic airmasses) over the region (Williams & Bradley, 1985).



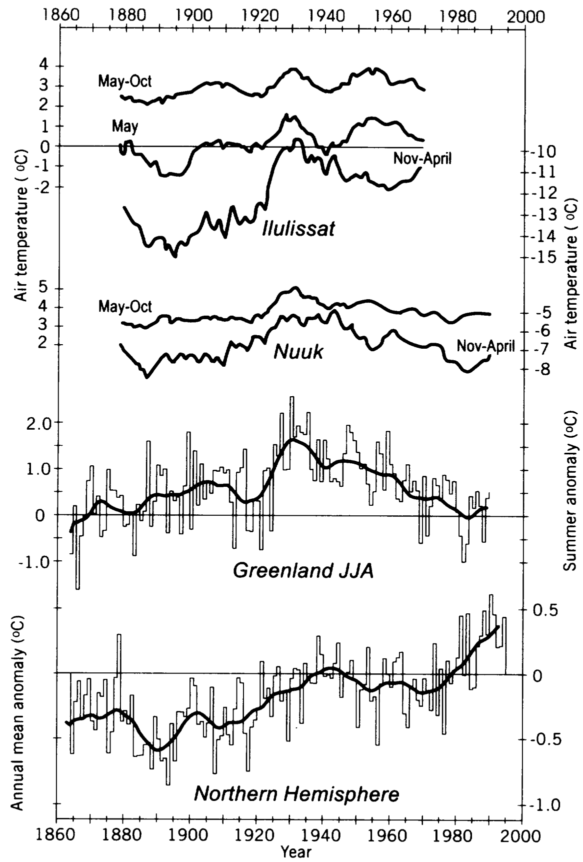


Fig. 3. Selected West Greenland coastal temperature (°C) records for the period 1878-1971 (Ilulissat) and 1878-1990 (Nuuk) and summer (JJA) temperature departures relative to 1866 to 1885 averaged over Greenland (Warrick & Oerlemans, 1990 and National Climatic Data Centre) compared to Northern Hemisphere combined annual land-surface air and sea surface temperature anomalies (°C) 1864-1995, relative to 1961 to 1990. (Nicholls et al., 1996). Solid smoothed curves are 20-yr running averages of monthly and annual means.

climatic changes are associated with variable cyclonic activity and depression tracks over Greenland bringing either warm/moist maritime or dry/cold arctic air masses over West Greenland. As a result, the inland regions experience alternating phases of continental or maritime conditions (Hasholt & Søgaard, 1978). The influence of atmospheric circulation shifts on cyclone tracks over the Greenland Ice Sheet during major Holocene warming and cooling phases is also evident from snow accumulation records (Kapsner *et al., 1995*) and long-distance pollen records in the region (Nichols *et al., 1978*; Fredskild, 1984).

Lake sediments as climate archives

The biological, chemical and physical cycles of lakes in extreme environments with steep climatic gradients are strongly influenced by meteorological variability (e.g. Welch *et al., 1987*; Schindler *et al., 1996*; Hardy, 1996; Doran *et al., 1996*). 7-9 months of lake ice cover and the presence of catchment permafrost prevents sediments and nutrients to enter the water column during the larger part of the year. As a result, sediment fluxes and deposition in Arctic lakes are usually strongly seasonal (e.g. Miller, 1992; Zolitschka, 1996). Moreover, the opacity of thick, partially snow covered ice and the low incident angle of sunlight during the 7-9 months of complete ice cover effectively reduces the active biological period (Hobbie, 1973; Welch *et al., 1987*; Fee *et al., 1987*; Smol *et al., 1996*; Vincent & Pienitz, 1996). Like for other Arctic regions, meteorological conditions are notoriously variable, with differences of 5°C in annual air amplitude not being exceptional. Variable temperature and precipitation regimes greatly affect important environmental conditions, such as length of the ablation season (Bradley & Miller, 1972; Tramoni *et al., 1985*; Magnuson *et al*., 2000), growing season length (Myeni *et al., 1997*), and the hydrological balance (e.g. Fritz *et al., 1999*). Small lake ecosystems are relatively simple and minor climatic shifts are expected to generate large changes in biota and depositional processes. In combination with such lake sensitivity, the majority of limnological and sedimentological changes recorded in lake sediments are, therefore, likely to represent climate-lake-catchment interactions.

Environmental background

Physiography

The Kangerlussuaq region is situated in the interior area of southern West Greenland (Fig. 1). Together with the coastal region, the inland area forms the most extensive ice-free continuous land in West Greenland. The Kangerlussuaq region can be characterized as a gentle WSW-ENE trending hilly landscape. Relief amplitude is of the order of 200-300 m with a few summits exceeding 600 m a.s.l. where glacially abraded bedrock knolls alternate with extensive plains and large, bedrock controlled lakes (Hansen, 1970).

Depositional features related to former ice expansion are widespread in the area (Weidick & Ten Brink, 1970) and include moraines, thin patches of loose gravelly and sandy diamicton, and scattered erratic boulders (Funder, 1989). Glaciofluvial and fluvial sediments, outwash plains, and fluvial terraces cover the floors of all major valleys. Deglaciation and subsequent relative land uplift resulted in terraced proglacial sediments that grade down-valley from ice-marginal deposits into emerged marine deltas in the valley mouths near the fjord (Ten Brink & Weidick, 1974; Van Tatenhove *et al.*, 1996). Usually the valleys contain several generations of river terraces (Funder, 1989).

Floodplains provide the source of widespread aeolian deposits in the area (Hansen, 1970; Dijkmans & Törnqvist, 1991). Coarser sediments are deposited as small dunes and sand sheet deposits in the valleys. Silt-sized sediments are carried over larger distances, and most of the area close to the present Inland Ice margin is covered by a thin (0.2–1.0-m-thick) layer of loess (Böcher, 1949; Dijkmans & Törnqvist, 1991).

Geology and glacial history

The Kangerlussuaq area is dominated by Precambrian crystalline metamorphic rocks with a general ENE strike, belonging to the Nagssugtoqidian mobile belt (Escher *et al.*, 1976). The bedrock consists mainly of granodioritic gneisses locally intruded by deformed amphibolitic and doleritic dykes.

Ice cover extended beyond the present day coastline during the larger parts of the Sisimiut glaciation of Late Wisconsinan-Weichselian age (Funder, 1989). During the course of deglaciation, the Inland Ice margin progressively recessed about 175 km in central West Greenland. Frequent halts or re-advances interrupted the retreat and formed extensive moraine systems, providing one of the most complete records of Greenland deglaciation history (Ten Brink & Weidick, 1974; Funder, 1989; Van Tatenhove *et al.*, 1996). ^{14}C dating of basal organic lake sediments indicated that the terrain became ice free as early as 7250 cal yrs B.P. (calendar years before 1950) close to the present Inland Ice margin (UtC-1987, Van Tatenhove *et al.*, 1996). Between 6000 and ca. 4000 cal yrs B.P. the ice margin was behind its present position (Weidick *et al.*, 1990, Anderson *et al.*, 1999) followed by a general readvance reported to begin at ca. 4500-4000 cal yrs B.P. (Anderson *et al.*, 1999; Bennike & Weidick, 1999).

Due to isostatic rebound, the Holocene marine limit rises from approximately 40-60 m a. s. l. at the head of Kangerlussuaq (oldest shells K-1664: 7140 ± 130 cal yrs B.P. and UtC-3522: 7100 ± 70 cal yrs B.P., reservoir corrected) to a position 130-140 m above relative sea level at the outer coast, reported to be older than 9500 ^{14}C yrs B.P. (Ten Brink &

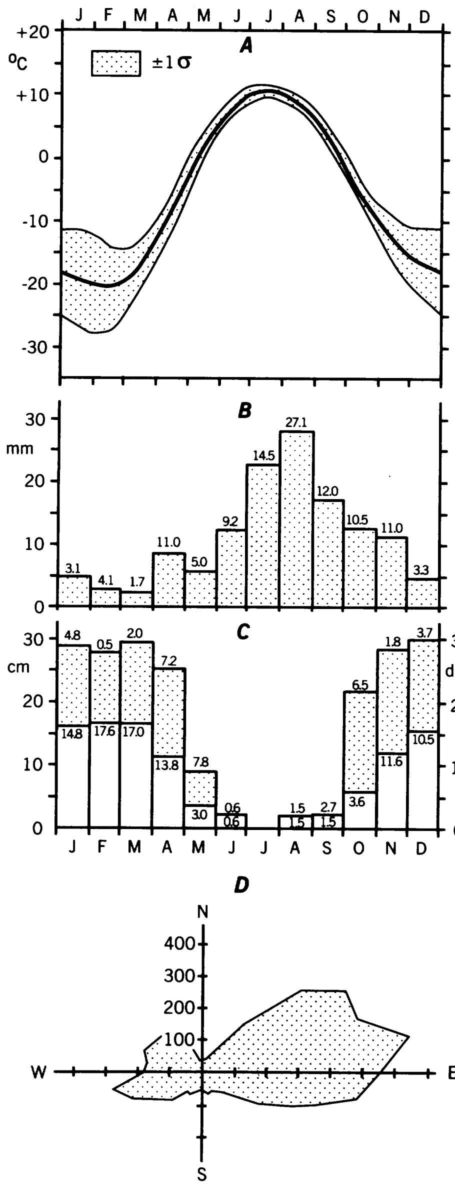


Fig. 4. Climatological data from the Kangerlussuaq Airport meteorological station (W.M.O 04231) for the period 1974-1987 (Danish Meteorological Institute, Copenhagen). A) Mean monthly temperature. B) Mean monthly precipitation, mm water equivalent. C) White bars: mean monthly height of snow cover for days with a complete snowcover. Shaded bars: mean number of days per month with a complete snow cover. Numbers in B) and C) indicate one standard deviation. D) Windrose for the Kangerlussuaq Airport meteorological station showing total number of 3-hour measurements with a wind speed > 5 m/s for the period 1974-30/9/1988 (Dijkmans & Törnqvist, 1991).

Weidick, 1974). According to Weidick (1993), isostatic adjustment ended after ca. 3000 cal yrs B.P. followed by a subsequent transition to submergence as a result of general ice expansion.

Climate and vegetation

Major climatic differences are observed within the ice free area where the interior position and the proximity to the Greenland Ice Sheet enhances the continentality of the Kangerlussuaq area. The present-day climate is arid continental subarctic (Fig. 4). There are about 4 months of continuous sunlight in summer and 2 months of complete darkness in winter. Mean annual temperature for Kangerlussuaq Airport weather station is –5.2 °C (1943-1992). Average temperature for February is –26 °C and +11 °C for July. Interannual temperature variability is large, especially during winter months. Continuous permafrost occurs throughout the region (Van Tatenhove and Olesen, 1994).

Extremely low precipitation in the inland areas compared to the coastal region (Fig. 5) is due to the rain shadow effect of the coastal ranges and the highlands around Sukkertoppen (Putnins, 1970; Reeh, 1989). Kangerlussuaq Airport weather station records a mean of 155 mm (period 1943-1992), with a summer maximum when evapotranspiration rates exceed net precipitation (Rott & Obleitner, 1992). Excluding meltwater discharge from the Inland Ice and a brief spring meltwater peak, runoff is very low.

Wind speed and direction close to the Inland Ice margin is largely controlled by the influence of the ice cap and the effect of topography. Dominantly easterly winds at ground level reflect the effects of both thermally induced katabatic (gravity) winds from the Inland Ice and airflow channeling in the valleys (Rott & Obleitner, 1992; Van den Broeke *et al.*, 1994). At greater altitudes, wind directions show a much wider distribution and no pronounced maximum of frequency (Hedegaard, 1982; Rott & Obleitner, 1992). In general, wind speeds are low (Hedegaard, 1982).

The dry climate, combined with the scarcity of meltwater, limits the presence of humid, luxuriant vegetation (Holt, 1980). Drier upland sites and south-facing slopes support a variety of dwarf-shrub heaths, dominated by the deciduous shrub *Salix glauca*, dwarf shrub vegetation (*Betula nana*, *Vaccinium uliginosum*), and steppe-like communities including xerophytic grasslands. Meso-hygrophytic vegetation occurs on north-facing slopes and in wet meadows developed in depressions, where bryophytes (*Aulacomnium* spp., Sphagnacaea), sedges (*Carex* spp.) and grasses (*Poa, Calamagrostis*) dominate. Most lakes support a 2-10 m wide fringe of aquatic macrophytes (*Hippurus vulgaris, Menyanthes trifoliata*) together with submerged species (*Myriophyllum spicatum, Potamogeton* spp.) and aquatic bryophytes.

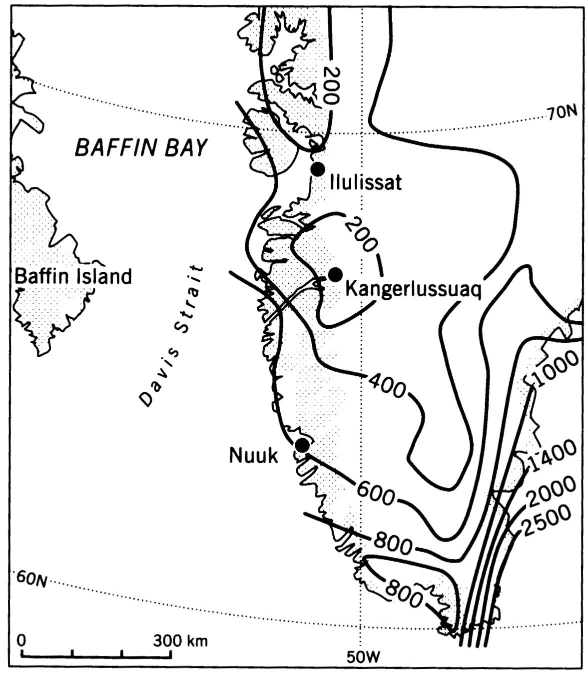


Fig. 5. General distribution of precipitation in western and southern Greenland, mm water equivalent per year, based on accumulation rates in ice-covered areas and recorded precipitation at coastal weather stations (after Reeh, 1989). Note lee-effect of mountain highlands in the West Greenland interior.

Kangerlussuaq lakes

Lakes of all sizes are a conspicuous feature of the inland ice-free region, which makes up about 5-10% of the surface area (Table 1). The lakes are either large, oblong shaped, bedrock controlled basins, ice-disintegration depressions formed in glaciofluvial sediments, or related to damming to varying degrees by glacial drift (Fig. 6). Basin subsidence due to degradation of ice-rich permafrost and/or buried glacial ice is thought to be an important process for the formation of many of the shallow lakes used in this study. Surface lowering due to melting of ground-ice in areas occupied by lakes amounted up to 4.0 m in the period 1943-1968 and 2.3 m for 1968-1985 in recently deglaciated terrain (Van Tatenhove, 1996). Small 1-2 m deep thaw-lakes are another conspicuous feature in peat-filled basins, and most likely develop from degraded ice-wedge polygons (Billings & Peterson, 1980).

The most extensive limnological studies of West Greenland lakes are those of Hansen (1967) and Anderson and coworkers (Anderson *et al*, 1999; 2000; 2002; Brodersen & Anderson, 2000; Bro-

Table1. Percentage lake distribution in the ice-marginal Kangerlussuaq area (map co-ordinates (UTM) x:488 y:7449 and x:545 y:7410). Means for 5 randomly selected 5x5 km square areas. *n* denotes number of lakes. Data based on Vandrekort Vestgrønland, sheet Kangerlussuaq 1:100.000 (Greenland Tourism A/S, 1996) and aerial photographs (series 282 /1968/ scale 1:47.400).

Upland areas (300-800 m a.s.l.)	Ponds (%)	0.5-2 ha (%)	2-25 ha (%)	> 25 ha (%)	*n*
Mean	60,9	30,8	6,1	1,7	89,8
Standard dev.	7,5	7,8	3,7	1,2	6,6
Lowland areas (0–300 m a.s.l.)					
Mean	47,3	31,3	14,3	7,0	41,3
Standard dev.	27,8	9,9	8,9	9,8	13,6

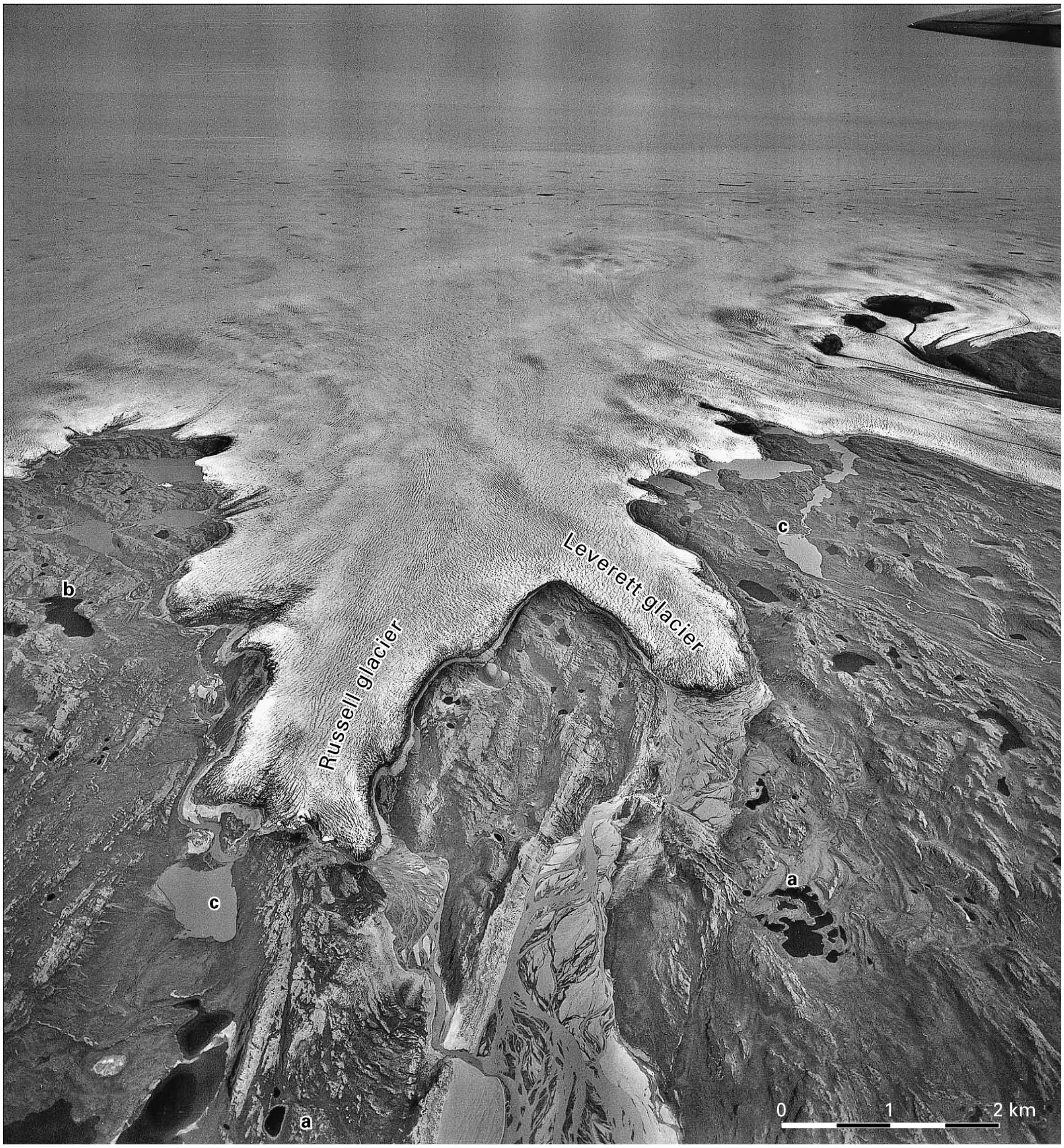


Fig. 6. Oblique aerial photograph of the ice-marginal landscape east of the Kangerlussuaq Airport, showing the Greenland Ice Sheet and the proglacial area in 1954. View is towards the east. Lower foreground: Russell and Leverett outlet glaciers, the outwash plain, lakes in peat filled depressions (a), bedrock controlled basins (b) and glacially fed lakes (c). Photo B23 A-L, no. 87 (1954) (reproduced with permission of National Survey and Cadastre – Denmark).

dersen & Anderson, *in press*). Details on water chemistry and flora of the lakes in the Kangerlussuaq region are presented in Böcher (1949), Brodersen & Anderson (*in press*), and Anderson *et al*. (2002). Notes on lake fauna are provided in Røen (1962), and Brodersen and Anderson (*in press*). The shallow lakes are generally moderately nutrient rich (total phosphorus 7-34 µg/L), soft water lakes, although the dissolved chemical constituents and transparency due to dis-

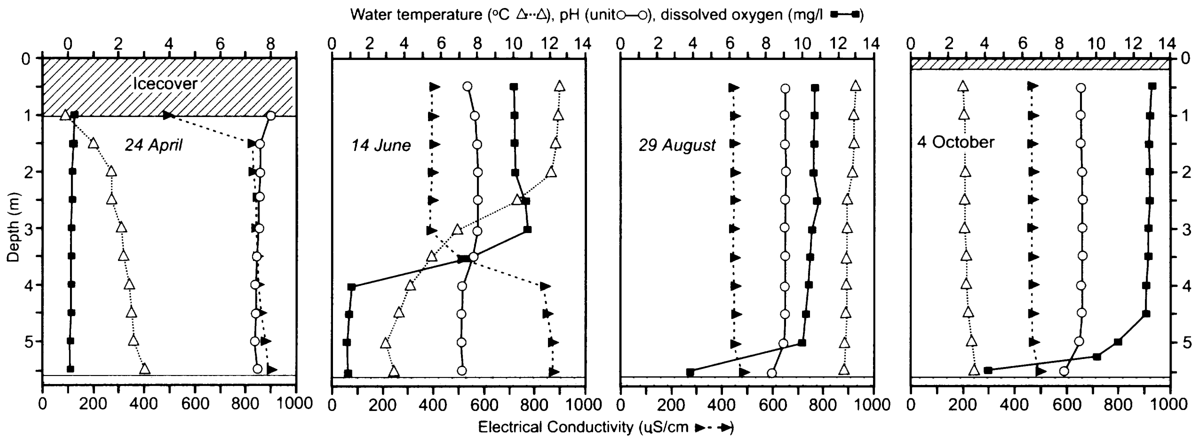


Fig. 7. Profiles of water temperature, dissolved oxygen, electrical conductivity (EC) and pH for lake SFL4 during four selected days (April 24th, June 14th (five days after ice-out), August 29th and October 4th, 2001) plotted against depth in the lake. Note the EC gradient due to dilute water input just after the lake became ice free and the resulting density stratification that developed until its breakdown end of August 2001.

solved coloured organic matter varies between lakes (Brodersen & Anderson, *in press*; see also Table 4).

The lakes begin to freeze-up in late September and ice grows to 1-2 m thickness during winter. Ice-free conditions extend from late May-early June to late September, although spring thaw for the larger and deep lakes can be some weeks later (Hansen, 1967). Hansen (1967) suggested that in the absence of strong winds, a weak and shallow summer stratification develops during summer, where dilute sheltered lakes might have two periods of full circulation throughout the watercolumn (see also Brodersen & Anderson, 2000). Spring warming (Late April-May) brings dilute water from the melting snowpack towards the lakes. These combine with dilute waters from the melting lake ice to generate a spring/summer salinity gradient in lakes with somewhat elevated salinities of 400-1000 μS/cm, a condition that can last throughout most of the summer (Fig. 7).

Due to the dry climate and the thin snowcover, meltwater flux and overland flow towards the lakes is limited, and, together with the lack of clear outflows, the water balance is for most cases controlled by spring overflow, direct precipitation and evaporative losses. Some groundwater exchange may occur, however, through percolating water on the permafrost table and from the open talik beneath lakes floored by mantels of unconsolidated sediment. A number of lakes have such high conductivities that they can be classified as oligosaline lakes (Böcher, 1949; Williams, 1991). Situated well above the local marine limit, their relative high salinity and conductivity is most likely a result of the high evaporation in relation to precipitation and lack of a distinct overflow (Anderson & Bennike, 1997; Anderson *et al.*, 1999), and most of these more saline lakes are meromictic (Brodersen & Anderson, 2000).

The biological structure of the small and shallow lakes is in general simple. Most lakes contain abundant aquatic macrophytes in terms of biomass, including emergent and submerged species and aquatic bryophytes. In general, the lakes support a large population of zooplankton and benthic invertebrates. The dominant groups of algae can vary distinctly between the shallow lakes (Hansen, 1967), but consist in general of desmids and diatoms (periphytic and benthic species). No fish species are found in the shallow lakes owing to low oxygen levels in the watercolumn during the long ice-cover period (Fig. 7).

Typically, surficial sediments consist of highly aqueous, olive green amorphous organic matter mixed to varying degrees with silt-sized mineral grains and limnic organic detritus (gyttja).

Fieldwork and methods

Site selection and sampling

Bathymetry, sediment infill and sediment characteristics of 41 relatively small (0.5-2 ha) and shallow (3-8 m) lakes were surveyed during two summer field seasons in 1991 and 1993. Most lakes are unnamed, and an un-official lake code was assigned to each of these sites. Six lakes were selected for detailed study in the Sandflugtdalen area (Fig. 1). Data were collected on bathymetry, basin morphometry, lake area, drainage area, postglacial sediment thickness, and the absence of distinct inlet or outlet channels. In addition, two more lakes were cored along Kangerlussuaq in 1996.

Prior to sampling, the bathymetry and postglacial sediment distribution were mapped along 10x10 or 20x20 m rectangular grids. Mapping included routine sediment description (sediment texture, colour, plant remains) with the aid of a 1.20 m long, 3.5 cm diameter Russian sampler. A rubber boat, fixed to the shoreline was used as a platform for coring. A number of sediment cores were taken in all lakes. Coring was carried out using a modified Livingstone sampler with a removable inner tube, a length of 1.20 m, and an inner chamber of Ø 50 mm. Overlapping core segments of at least 30 cm were taken throughout each section. Sediment cores (piston type Ø 75 mm) from the uppermost fragile ~60 cm were carefully collected by a scuba diver, to minimize the risk of non-recovery of topmost sediments.

Surface sediment samples (1-5 cm thick, 100 cm^3) from lake 17 in Sandflugtdalen (SFL17), were collected over a regular (10x20 m) grid to analyze the spatial variability and composition of (sub)recent surficial sediments. In addition, surface water samples were collected for chemical analysis during three consecutive days in July 1997, together with grab samples from the central surficial sediments. Subsamples for conductivity (EGV), alkalinity, pH, water temperature and transparancy (Secchi-depth) were measured in the field. Dissolved chemical constituency was measured on acidified subsamples using colometry (FIA) and ICP-AES spectrometry, together with bulk density estimates, organic matter content and Kjehldahl-N of surficial sediment samples.

Bathymetry, sediment volume and distribution

Bathymetry and sediment distribution was calculated by contouring the gridded sampling points using the software package SURFER and ordinary kriging for interpolation. For some cases, permafrost in shallow littoral areas (generally <1 m water depth) inhibited penetrating deeper than 50 cm into the sediments. For those cases the sediment thickness was estimated by extrapolation from surrounding points. Accordingly, subaqueous slopes were calculated throughout the modern lakes in order to identify lake-areas liable to sediment slumping (Larsen & McDonald, 1993). Bathymetry of the lakes prior to the start of organic infill was calculated by subtracting postglacial sediment thickness from the modern bathymetrical maps. Although the level of the lake surface at the time of onset of organic sedimentation is not known, it is assumed to be similar, or close to the modern level.

Sediment analysis

Wet surficial samples (100 cm^3) were

sieved over a 210-μm screen to split coarse particulate organic matter and macrofossil remains from the bulk sediment matrix, and subsequently analyzed using a dissecting microscope. Both fractions were thereafter oven-dried at 105°C for bulk weight and ashed in a muffle furnace at 4 hours/550°C for organic matter content (loss on ignition (LOI), Dean, 1974).

Cores were split lengthwise and photographed at the Laboratory of Physical Geography, Utrecht University. Precise subsampling of the sediments was done by allowing frozen sediment cores to warm to –2°C overnight, after which the sediments were sliced in contiguous 2-4 ± 0.3 mm samples in a refrigerated room. Samples from selected core segments were split into two equal parts: I) for LOI analysis, and II) to determine biogenic silica concentrations and further geochemical analysis. Sample I was oven-dried at 105°C for 12 hours to determine bulk dry weight, and subsequently at 550°C for 4 hours in a muffle furnace for LOI. The analytical error of LOI estimates (0.8% for standard 0.5 g dry weight samples) was determined by measurement of 39 replicate samples. Additionally, ashed samples were analyzed for remaining macrofossils. For carbonate content, the ashed sample I was placed in a muffle furnace at 950°C for 4 hours. For sample II, dry bulk samples were oxidized to remove organic matter. Authigenic mineral matter (formed within the lake proper) was separated from allogenic (detrital) mineral matter following extraction procedures of Engstrom & Wright (1984). The biogenic silica concentration in the authigenic mineral fraction (selective dissolution with 0.2 M NaOH), and authigenic and allogenic elemental composition was measured using ICP-AES spectrometry. The relative analytical error for biogenic silica content is ~15%.

Diatom and pollen analysis

Samples for diatom analysis were taken from one of the lake cores at intervals of 5 cm. Preparation and analysis of diatom samples followed standard techniques of the Paleobotanical Laboratory, Geological Survey of the Netherlands. Briefly, samples were treated with hydrogen peroxide (H_2O_2) and hydrogen chlorid acid (HCl) to digest organic material. $KMNO_4$ was added to remove coarse organic matter. Digested samples were repeatedly rinsed to neutrality in distilled water. Mineral grains were removed from the sample by bringing samples in suspension and resettle for a few seconds after which an 0.5 ml sample was taken just above the surface of the tube, and subsequently transferred and dried on 22x22 mm coverslips. Coverslips were mounted on slides with Naphrax, a medium with a high refractory index. For each sample 200-250 valves were counted and identified along random transects using oil immersion objectives at magnifications of 1000x or more. Identifications were made with references to the works of Foged (1958, 1972), Krammer & Lange-Bertalot (1986), Hendey (1964) and Patrick & Reimer (1966).

A small number of samples for detailed pollen analysis were selected from a new core, and the pollen diagram is discussed in Eisner *et al.* (1995). A set of subsamples were analysed from the upper 40 cm of the sediment profile from this lake, encompassing levels of lithostratigraphic change with a common regional significance (Willemse & Törnqvist, 1999). Subsamples were taken, and prepared using standard pollen procedures (Eisner *et al.*, 1995).

^{14}C analysis

Twenty eight accelerator mass spectrometry (AMS) ^{14}C ages were obtained for temporal control. All samples are from bioturbated sediments and were collected during subsampling. Macrofossil remains, when present, were picked from the central part of the sediment surface during slicing, carefully cleaned, and stored in distilled water acidified to pH<2 with HCl before further identification and preparation. In some cases bulk sediment slices were sieved over a 350-µm mesh to extract macrofossil remains, after removal of the outer, possibly contaminated, surface layer. As far as possible, samples were selected that: I) contained sufficient and well preserved identifiable terrestrial macrofossils for AMS ^{14}C dating, and II) were evenly spaced throughout the sequences. Because in a number of cases there was not enough material to fulfil these criteria, bulk sediment samples had to be used. To determine possible reservoir effects, comparitive dating between bulk sediment samples, terrestrial, and aquatic macrofossil material was performed on single samples.

Macrofossil material for ^{14}C analysis was submitted within five months after collection of the cores (Wohlfart *et al.*, 1998). AMS ^{14}C measurements were carried out at the Robert J. Van de Graaff Laboratory, Utrecht University, and converted to calendar years (cal yrs before 1950) using the CAL20 programme (Van der Plicht, 1993) and smoothed calibration curves (Törnqvist & Bierkens, 1994). Three samples were rejected from the dataset. Two bulk gyttja samples yielded ages that were too young (UtC-5602 and UtC-1988, see Table 6). Both consisted of fine detritus gyttja from cores (SFL17.1 and SFL4.2) collected from the marginal parts of the lakes. Contamination is suspected for these samples because this sediment type may contain minute plant rootlets. This problem is absent in cores taken from the central parts of the lakes where the sediments consist entirely of structureless gyttja. One macrofossil sample yielded an age inconsistent with its stratigraphic position (UtC-5613) and most likely consisted of reworked (older) material from outside the lake proper. Possible ageing of bulk sediment samples (e.g. reservoir effects) is unlikely due to the absence of carbonate rocks, other rocks containing old carbon in the catchments, and the rapid water mixing in shallow lakes. The absence of reservoir effects is also suggested by replicate samples of terrestrial (*Salix glauca* wood fragments) and (semi)-aquatic (*Hippuris vulgaris* fruits) macrofossils from the same stratigraphic level in core SFL17.1, yielding ^{14}C ages of 4016 ± 38 (UtC-5604) and 4037 ± 39 (UtC-5612), respectively. Bennike (2000) reported a possible small hard-water effect (50-100 yrs) in limnic fossil remains of Store Saltsø, but the residence time in this lake are likely longer than for the smaller lakes, and effects of sediment mixing on the sample time-width (Törnqvist & Bierkens, 1994) have not been accounted for. Radiocarbon ages from the second core in lake SFL17 (SFL17.1) are largely based on terrestrial macrofossils. According to extrapolated ages the timing of main lithostratigraphic changes are consistent with those reported for the first core SFL17.2 (Eisner *et al.*, 1995), which were entirely based on bulk gyttja material.

Results

Bathymetry and sediment distribution

Details on location, morphometry, drainage area, hydrochemistry, sediment infill, and sediment volume of the lakes are presented in Tables 2 to 4. The modern lakes have regular, slightly concave shapes, with maximum depth (Z_{max}) being highly correlated to mean depth ($\bar{Z}$): r = 0.968, and $\bar{Z}$:Z_{max} ratios averaging 0.33. Subaqueous slopes are low, averaging 1.5-2°, with 3.4° ± 3.1° for the steepest lake SFL46, and, together with the stabilising effect of benthic vegetation, gravity induced sediment redeposition due to slumping is unlikely (Larsen & McDonald, 1993). Drainage

Table 2. Location and morphometric parameters of shallow Kangerlussuaq lakes.

Lake	Location	Elevation (m a.s.l.)	L_a[a] (ha)	D_a[b] (ha)	D_a : L_a	Z_{mean}[c] (m)	Z_{max} (m)	Z_{mean} : Z_{max}	n[d]
SFL4	67⁰ 05'11"N / 50⁰ 17'22"W	247	2.17	17.95	8.27	1.58	5.30	0.30	45[(f)]
SFL6	67⁰ 04'56"N / 50⁰20'37"W	261	1.05	4.37	4.16	1.39	5.37	0.26	23[(f)]
SFL17	67⁰ 03'24"N / 50⁰27'03"W	128	0.96	22.33	23.22	1.16	2.80	0.41	60[(g)]
SFL35	67⁰ 02'43"N / 50⁰37'58"W	490	1.75	9.90	5.65	2.18	8.20	0.27	56[(g)]
SFL46	67⁰ 02'02"N / 50⁰ 41'46"W	385	0.27	5.24	19.40	2.30	5.50	0.42	35[(e)]
SFL49	67⁰ 01'37"N / 50⁰ 44'15"W	465	1.15	8.16	7.09	1.31	4.20	0.31	108[(e)]
ANGU2	66⁰ 43'45"N / 51⁰ 25'51"W	225	[h]	[h]	–	1.80	2.10	–	8
NAUJG1	66⁰ 40'01"N / 51⁰58'11"W	300	0.56	23.31	41.44	1.15	3.15	0.37	25[(g)]

[a]lake area; [b]drainage area excluding lake area; [c]waterdepth; [d]number of observations; [e]10 x 10 m coring grid; [f]20 x 20 m coring grid; [g]10 x 20 m coring grid; [h]restricted dataset available only for the western basin.

Table 3. Lake sediment distribution and basin volume.

Lake	S_{mean}[a] (m)	S_{max} (m)	S_{mean}: S_{max}	Volume: Basin[b] m^3 x 10^3	Volume: Sediment[c] m^3 x 10^3	S_v : B_v[d]	S_v : L_a[e]	n[f]
SFL4	1.22	3.25	0.37	62	30	0.48	13.8	45
SFL6	0.93	3.85	0.24	34	16	0.47	15.2	23
SFL17	1.35	2.30	0.59	41[g]	21[g]	–	–	60
SFL35	1.68	3.20	0.53	59	30	0.51	17.1	56
SFL46	1.08	2.50	0.43	11	6	0.54	22.2	35
SFL49	1.14	2.40	0.48	28	12	0.42	10.4	108
ANGU2	1.80	2.80	0.64[h]	[h]	[h]	–	–	8
NAUJG1	1.10	2.45	0.45	19	9	0.52	16.1	25

[a]postglacial sediment thickness; [b]waterdepth plus postglacial sediments; [c]excluding glaciolacustrine sediments; [d]ratio sediment volume to basin volume; [e]ratio sediment volume to lake area; [f]number of observations; [g]limited area only (see main text); [h]restricted dataset available only for the western basin.

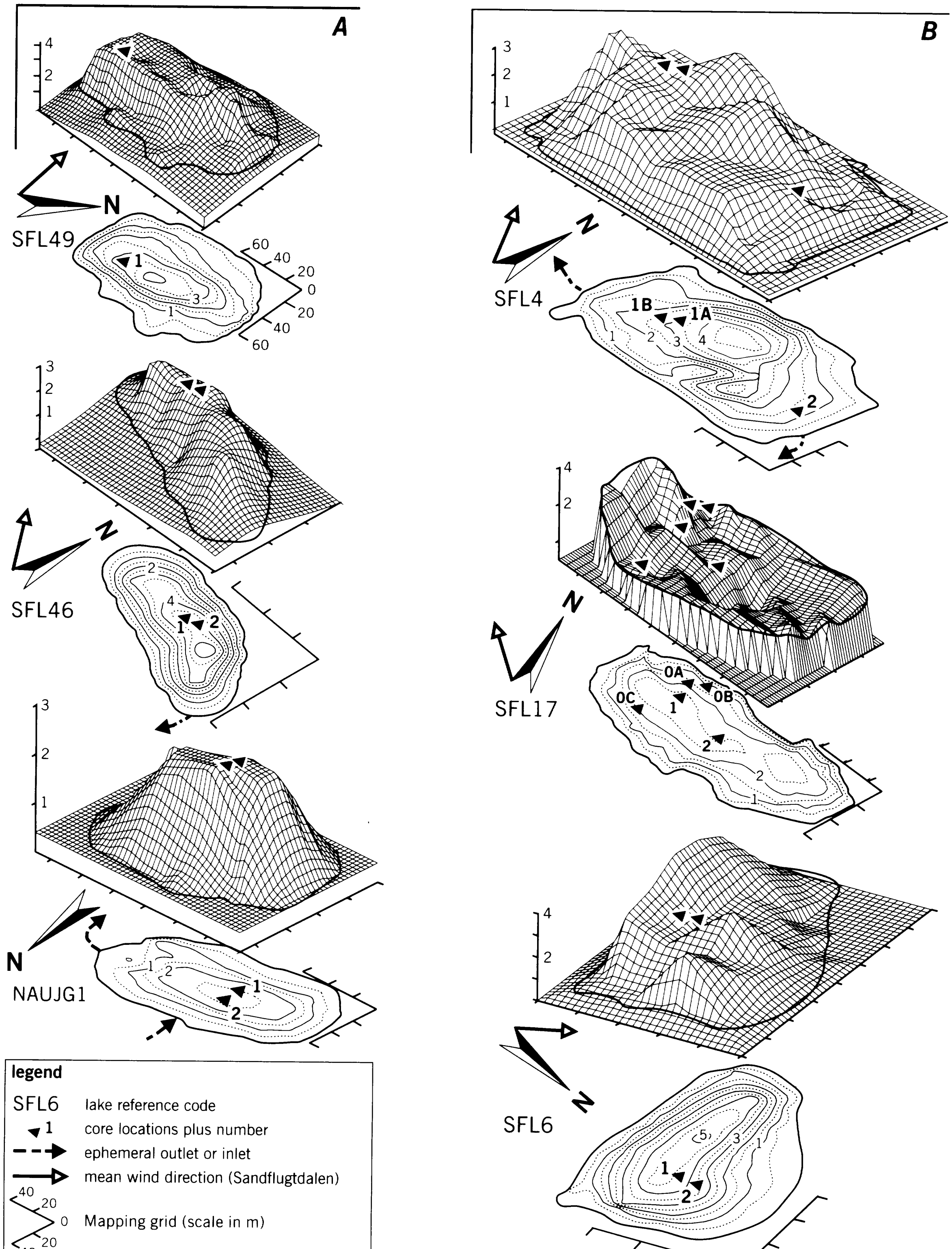


Fig. 8. Bathymetry and postglacial sediment infill of selected Kangerlussuaq lakes, showing the coring locations for lakes with symmetrical sediment distributions (A) and a-symmetrical or complex sediment distribution (B). Postglacial sediment bodies and bathymetrical maps shown as aligned 3D and 2D maps respectively (grid and depth scales in m).

Table 4. Physical and chemical variables of Kangerlussuaq lakes.

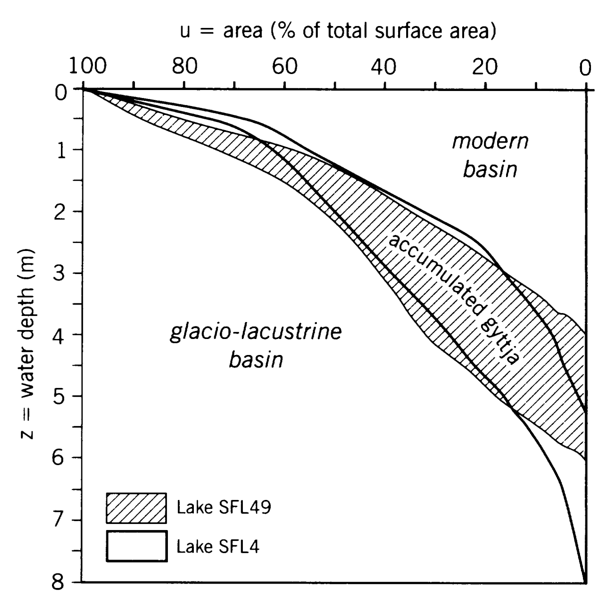
Lake	Cond[a]	Alk[b]	pH	Sec[c]	T[d]	Major hydrochemicals (mg/l)								
	(µS/cm)	(mmol/l)		(m)	(°C)	SO_4	Cl	Ca	K	Na	Mg	NH_4	TP (µg/l)	NO_3
SFL4[e]	497	0.9	8.95	>5	16.3	3.19	39.48	15.87	31.16	34.24	39.56	0.081	9.78	0.39
SFL6[e]	1131	1.6	8.45	4.2	16.2	3.87	154.4	13.44	63.25	92.84	78.60	0.083	10.43	0.55
SFL17[e]	229	2.0	7.04	2.1	15.3	4.03	13.01	10.80	11.58	13.82	19.37	0.108	14.99	2.34
SFL35[e]	90	0.7	6.81	3.8	17.6	2.05	1.17	7.47	2.08	3.87	4.51	0.103	17.28	2.21
SFL46[e]	57	0.6	6.9	2.2	17.1	52.6	1.17	4.97	2.05	2.73	3.83	0.066	9.78	1.23
SFL49[e]	66	0.55	6.92	2.0	17.0	59.5	1.78	5.06	2.15	3.59	3.78	0.145	19.56	1.85
ANGU2[e,f]	850	–	9.22	1.8	15.2	–	–	–	–	–	–	–	–	–
NAUJG1[e,f]	120	–	7.06	1.3	12.1	–	–	–	–	–	–	–	–	–
Large lakes and salt lakes														
Tasersuatsiaq	74	0.55	5.68	–	18,5	15.18	5.31	8.30	4.82	9.14	6.06	0.080	0.65	8.76
Mt. Evans[e]	375	0.1	6.65	–	17.5	15.177	33.48	29.54	23.84	56.36	29.08	0.065	10.76	0.47
Store Saltsø	2925	2.95	8.42	–	17.5	28.36	436.00	16.75	150.72	356.86	171.02	0.138	6.85	0.85
Hundesø	3950	3.65	9.01	–	18.3	165.49	663.00	12.17	180.06	491.19	210.32	0.010	10.76	0.47

[a]conductivity; [b]alkalinity; [c]secchi depth; [d]water temperature (at ~1 m);[e]unofficial name; [f]no watersamples available.

areas are relatively small, with drainage area to lake area ratios ranging from 5-22. Fig. 8A-B show the bathymetry and sediment distribution for six of the eight lakes that were mapped in detail. Accumulated organic rich sediments consist of soft, unstructured, olive green pelleted organic detritus mixed with silt sized mineral clasts. These are in general overlying stiff, faintly laminated sandy silts and fine sands, presumably of glaciolacustrine origin. The transition between these deposits is in most cases rather sharp, indicating a rapid change from an open glacially fed system to a closed system outside direct glacial influence.

The distribution of organic-rich sediments is relatively regular (average S_{mean}:S_{max}=0.44). Sediment distribution for topographically sheltered lakes with a small effective fetch is highly symmetrical (Fig. 8A). For less sheltered lakes (Fig. 8B), sediments have accumulated rather asymmetrically, preferably towards the downwind side of the lake (Odgaard, 1993). Maximum sediment thickness is generally found in close relation with the initial shape of the basin at the start of organic infill (e.g. lake SFL6). For lake SFL17, the situation is less clear (Fig. 8B). The modern lake is situated within a large elongated peat filled depression, probably a major former meltwater spillway. The presence of permafrost outside the lake proper inhibited mapping of the subsurface topography, but the distribution of accumulated sediments within the lake sug-

Fig. 9. Plot of area (u) vs. depth (z) for lakes SFL 49 and SFL4, where u is the percentage of the lake area deeper than z. The difference between the plot for the original basin derived from lowermost organic sediments and the modern basin represents accumulated gyttja.



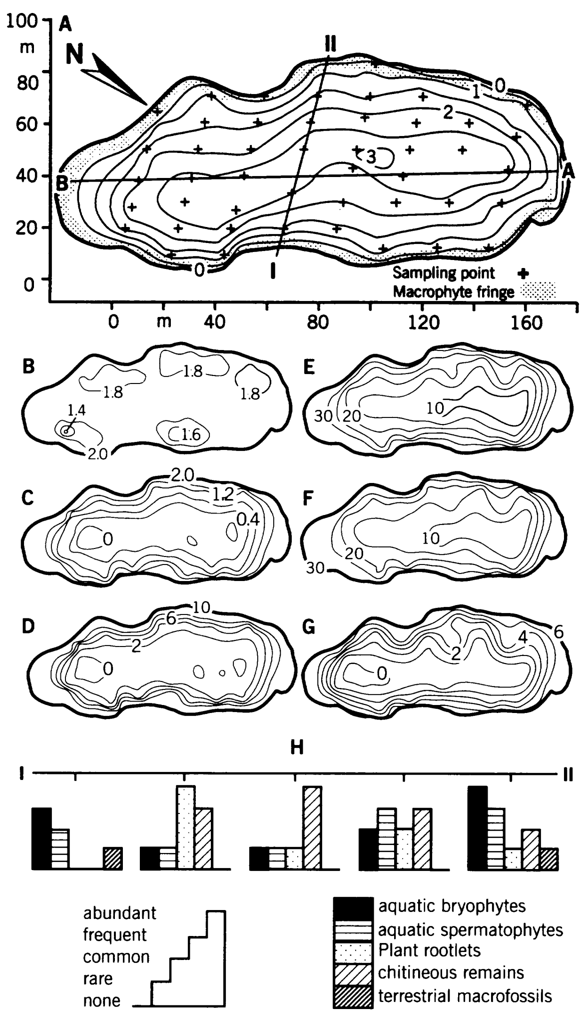


Fig. 10 A) Bathymetry and sampling points of lake SFL17. Contour lines in m water depth; +: Sampling point ; I-II and A-B: transects for Figs. 10H and 14, repectively. B) Dry density for bulk sediment samples (g/cm^3). C) Dry mass of coarse particulate organic matter (fraction > 210µm in g/dm^3 bulk sediment). D) Weight percentage coarse particulate organic matter. E) Organic matter content (%LOI) for 100 cm^3 bulk samples. F) Organic matter content (%LOI) gyttja fraction < 210µm. G) percentage difference LOI bulk gyttja and gyttja fraction < 210µm. H) Relative abundance of macrofossils along transect I-II.

gests that the larger part of the original basin has already been filled in.

The distribution and cumulative thickness of organic sediments as a function of water depth are exemplified in Fig. 9 for lakes SFL4 and SFL49. For lake SFL49 sediment accumulated conformably to the shape of the basin, suggesting sediment focusing towards the deeper parts of the lake to be the most important control on sediment accumulation. For the larger, more exposed lake SFL4, sediments accumulated more uniformly across the lake floor, notably for areas with overlying water deeper than the 1.5 m isobath, suggesting a more complex control on sediment accumulation.

Sediment volume to basin volume ratios (S_v : B_v) average 0.49, indicating that almost half of the lakes have filled in since glacial influence halted. Total postglacial sediment volume ranges from 6 x 10^3 m^3 for the smallest lake to over 60 x 10^3 m^3 for lake SFL4. Total sediment volume to lake surface area ratios are highly correlated (r=0.921), averaging 16.7 ± 3.9. Accumulated sediments are only weakly related to the size of the drainage area (r=0.292). This is in accordance with the closed character of the lakes, suggesting that accumulated sediments are primarily derived from within-lake sources, or from regional aeolian input.

Surficial sediments, composition and spatial patterns

Surficial sediments in lake SFL17 are representative for the small and shallow freshwater lakes around the head of Kangerlussuaq, and 47 evenly distributed surficial samples were collected to obtain information on the composition and spatial pattern of subrecent sediments (Fig. 10A). A sediment, pollen and macrofossil record from this lake is reported in Eisner *et al.*, (1995), and is based on a single core collected in 1985 by researchers from Stockholm University. Unfortunately, the precise location of the lake was misidentified as lake

SFL31 by the Swedish group, but a field check aided by photographic documentation pointed to lake SFL17 as the correct coring site.

Surficial sediments of lake SFL17 are highly aqueous (95-98% water content), and consist primarily of silt-sized detrital mineral grains (65-95% of bulk dry weight). These are homogeneously mixed with pelleted amorphous organic matter, and, to varying degrees, with organic detritus, mainly minute plant rootlets and plant fragments. Siliceous diatom cells, mainly epiphytic species, are present but not abundant. Fragmented cells are common and found as pelleted nests, indicating modification by bottom fauna ('coprogenesis' sensu Cole, 1994), and the accumulated sediments can be classified as gyttja. Comparison between grain-size data from grab samples of the central surficial sediments from all studied lakes with those for aeolian silts in the area (Dijkmans & Törnqvist, 1991) shows that the clastic lake sediments are somewhat coarser (Fig. 11). Although siliceous diatom frustules (10-200 μm range) were not removed from the lake samples, sorting of the material suggests an aeolian origin (Eisner *et al.*, 1995).

The vegetation at water depths < 2 m consist of both submerged and emergent species, is relatively rich in terms of biomass, and zoned according to water depth. This is reflected in the relative content of coarse plant fragments and plant microrhizomes in the sediment matrix. Fig. 10B-G presents the spatial characteristics of surficial sediments over the lake floor. Mean dry bulk densities average 1.8 g cm^{-3} with the lowest values found within the lake area <2 m depth (Fig. 10B). The most striking pattern is the close relationship between sediment composition and the depth of the overlying water column. Coarse organic detritus (fraction > 210 μm) makes up as much as 10% of bulk dry sediment weight in areas <2 m, with gradually decreasing values to less than 1% below the 2 m depth contour (Fig. 10C-D).

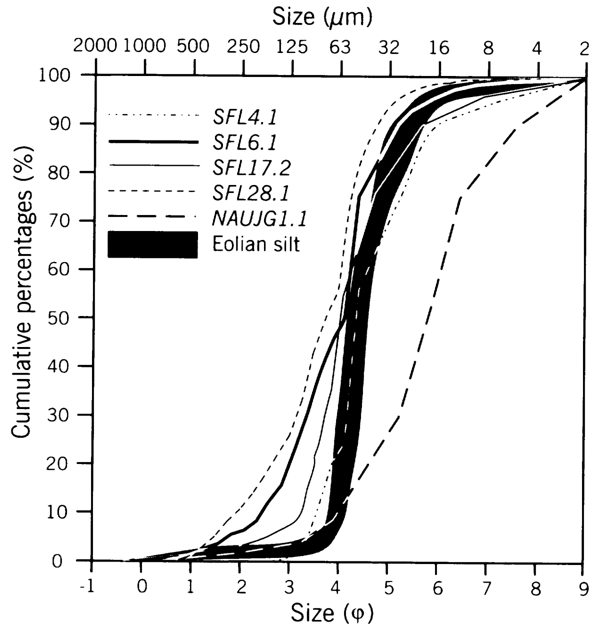


Fig. 11. Comparison of grain-size distributions of lake sediment surface samples and eolian silt in the Sandflugtdalen area, partly after Dijkmans & Törnqvist (1991) and Eisner et al.,(1995).

Figure 10H summarizes the relative abundance of different macrofossil types along a transect through the central parts of the lake. The coarse organic detritus in the shallower areas is dominantly derived from limnic plants (*Hippuris vulgaris*, *Menyanthes trifoliata*, *Myriophyllum spicatum*, *Potamogeton* spp.) and aquatic bryophytes (*Drepanocladus* spp.). Coarse organic detritus is largely absent in the central parts of the lake, and macrofossil material consists almost entirely of chitineous filaments of crustaceans (Fig. 10C-D), suggesting little zonal exchange of coarse organic material

Lake-wide patterns of organic matter content closely follow the zoned distribution of coarse organic detritus (Fig. 10E-G). Sediments in the central part of the lake are fine-grained with LOI values of 5-15%. Towards the shallower areas LOI increases to values of 30%, even if the coarse organic fraction is excluded from the analysis (Fig. 10F). Both distributions suggest that benthic and planctonic communities contribute organic matter directly from the overlying

Table 5. Bulk surface sediment composition of nonglacial lakes in West Greenland.

Lake[a]	L_A[b] (ha)	Z_{max}[c] (m)	I/O[d]	Lithology	LOI %	TOC[e,f] %	N %	C/N	$CaCO_3$ %	Si[g] %
1. Lake at Niaqornat	8.0	3.0	c	fine detritus gyttja	31.4	13.8	1.1	12	3.3	3.2
2. Lake at Ikerasak	0.8	6.0	o	amorphous gyttja	47.0	18.6	1.8	10	2.3	10.0
3. Taserssuaq, Ataa	1060	135	i	organic poor silt	4.7	2.4	0.6	4	0.7	10.8
4. Waterlake Qasigiannguit	2.0	4.0	i	amorphous gyttja	35.8	18.2	1.7	11	0.8	10.1
5. Waterlake Aasiaat	40.0	8.0	i	silty gyttja	21.2	10.0	1.3	8	0.4	14.3
6. Lake at Blæsedalen	7.5	5.0	o	diatomaceous gyttja	15.3	7.3	1.0	8	2.4	22.4
7. Thygesens Sø, Qeqertarsuaq	3.0	3.0	c	diatomaceous gyttja	31.7	15.9	1.0	16	1.1	17.3
8. Lake on Kangaarssuk	24.0	7.0	i	diatomaceous gyttja	40.0	16.7	1.3	13	1.5	19.7
9. Lake Evqitsoq, Disko	12.0	5.5	i	diatomaceous gyttja	13.9	9.1	1.3	7	1.8	35.6
10. Lillesø, Disko	7.0	12.5	o	organic poor silt	7.2	1.9	0.4	5	3.2	3.2
11. Mellemsø, Disko	36.0	26.0	i	organic poor silt	6.7	3.3	0.6	6	4.1	4.2
12. Lake at Qivitut, Disko	8.0	4.5	o	organic poor silt	5.6	4.0	0.3	14	0.2	2.5
13. Sandflugtdalen SFL4	2.17	5.3	c	silty amorphous gyttja	18.3	8.6	0.9	10	0.3	–
14. Sandflugtdalen SFL6	1.05	5.4	c	silty amorphous gyttja	15.2	7.1	1.0	7	0.8	–
15. Sandflugtdalen SFL17	0.96	2.8	c	silty amorphous gyttja	23.6	11.1	1.3	9	0.5	–
16. Sandflugtdalen SFL35	1.75	8.2	c	silty amorphous gyttja	26.1	12.3	1.1	11	1.2	–
17. Sandflugtdalen SFL46	0.27	5.5	c	silty amorphous gyttja	24.3	11.4	1.1	10	2.1	–
18. Sandflugtdalen SFL49	1.15	4.2	c	silty amorphous gyttja	28.1	13.2	1.3	10	0.9	–
19. Angujaartorfik ANGU2	3.04	2.1	c	calcareous gyttja	33.6	15.8	1.8	9	20.1	–
20. Naujaallup NAUJG1	0.56	3.2	c	diatomaceous gyttja	36.1	16.9	1.9	9	1.3	20.0

[a]1-12 data, site names and location in Hansen (1967), 13-20 this study; [b]lake area; [c]maximum water depth; [d]inlet/outlet or closed lake; [e]total organic carbon; [f]using conversion factor of 0.47 for % LOI (Dean, 1974) for 13-20; [g]percentage biogenic silica of bulk dry weight.

watercolumn to the sediments, reflecting the zoned differences in productivity and the biological structure of the overlying water column.

Under the microscope, amorphous grey-brown aggregates dominate the organic compounds, and the general lack of terrestrial macrofossils in littoral sediments suggests that terrestrial sources for organic matter are limited. This is consistent with the closed character of the lakes and the lack of significant amounts of overland runoff. A dominant aquatic source for sedimentary organic matter in the lakes is also evident from the low C/N ratios (Meyers & Lallier-Vergès, 1999) obtained for surficial samples in the lakes (Table 5), which are comparable with those for other West Greenland lakes (Hansen, 1967).

Sediment stratigraphy

Sediment stratigraphy for the different lakes is illustrated by focusing on cores taken from the central areas of four widely separated lakes that exhibit three distinctly different types of lithology (Fig. 12).

Lakes with a large aeolian input, dominated by highly organic silty gyttja

Lithologically, sediments from lakes around the head of Kangerlussuaq are characterized by a high proportion of silt-sized aeolian grains (Eisner *et al.*, 1995). Core SFL4.1B was collected from the western central part of lake SFL4 (Fig. 8). The lake is situated at an elevation of 247 m a.s.l. and only 4 km to the

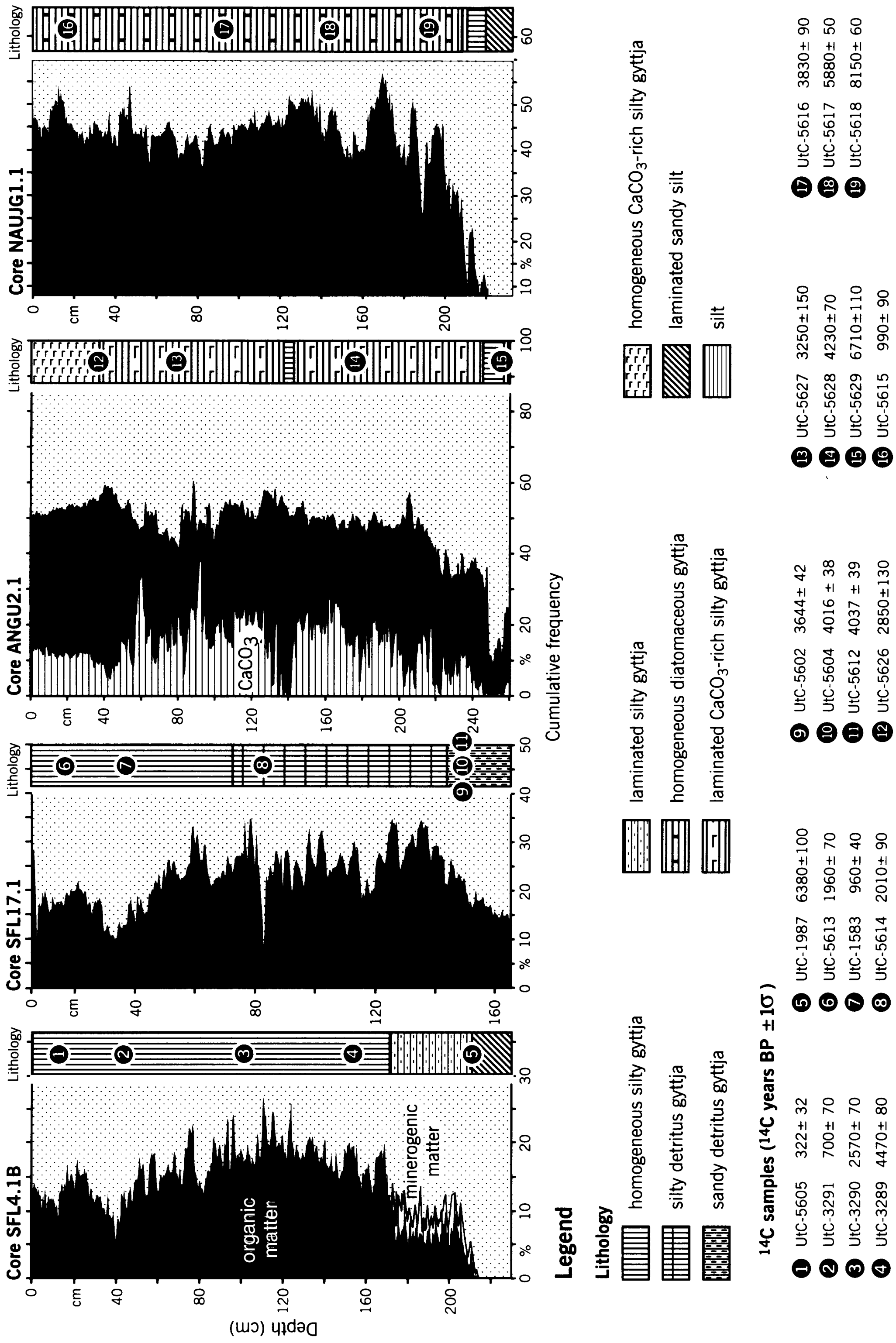


Fig. 12. Sedimentological properties and major lithofacies for cores from lakes SFL4, SFL17, ANGU2 and NAUJG1, including location of ^{14}C samples (Table 6).

Inland Ice margin (Figs 1 and 13A). The locality of the lake is a small peat filled basin. Northern and southern shorelines are dammed by outcropping bedrock and to the east by a low ridge of glacial drift. Lake SFL4 is located just outside the outer limits of the Ørkendalen moraine system (Ten Brink, 1975). A ^{14}C age

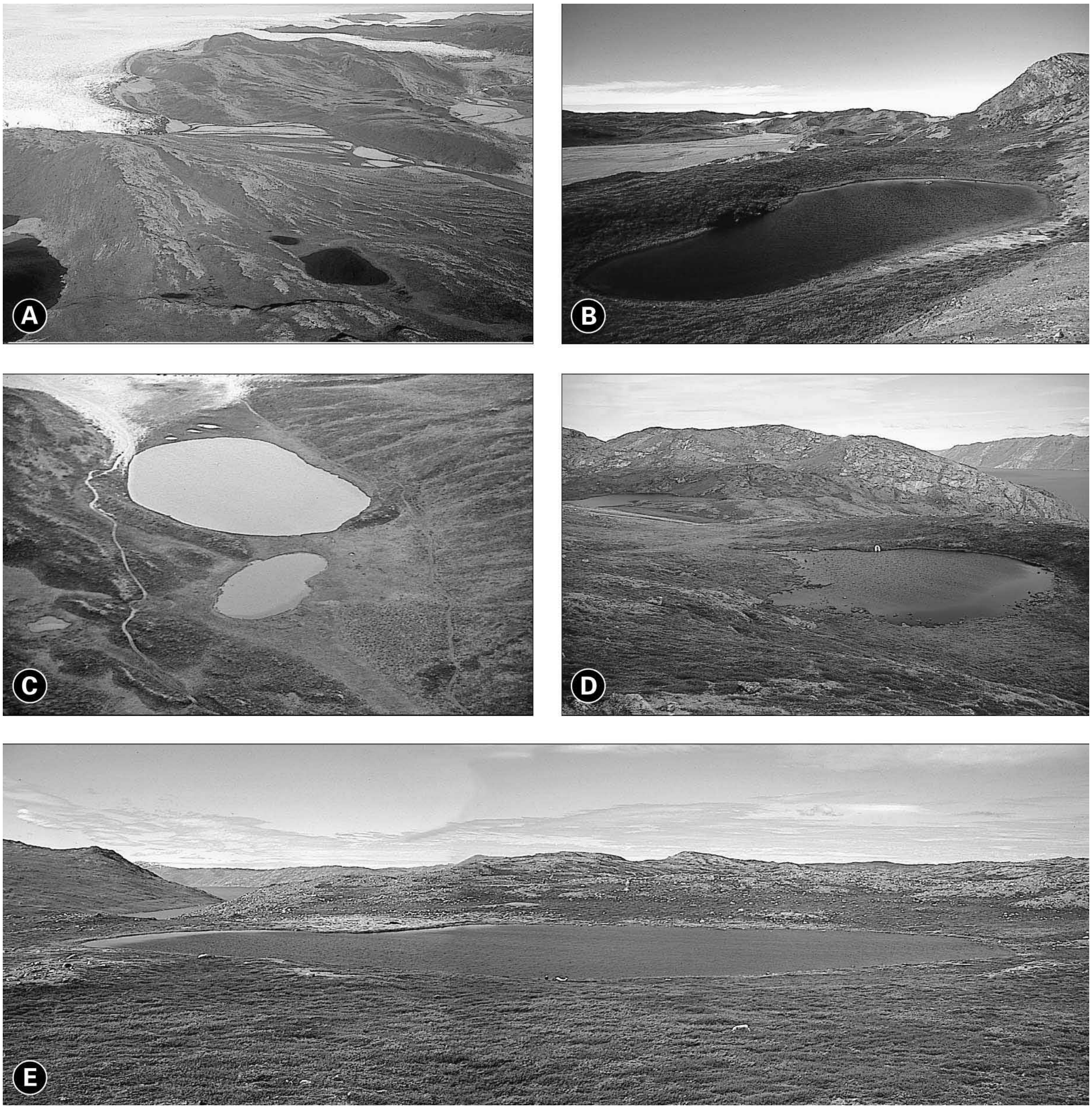


Fig. 13. Shallow lakes in the Kangerlussuaq region. A) Lake SFL4 (lower central lake) in Sandflugtdalen as seen from the west (July 1995). In the background the Inland Ice, with the Russell (lower) and Leverett (upper) outlet glaciers. B) Lake SFL6 in Sandflugtdalen as seen from the northwest (July 1991). In the background the Sandflugtdalen floodplain can been seen. Lake SFL6 is a fresh-brackish soft water lake (conductivity in 1997: 1131 µS/cm) with abundant benthic Characeae. Note the rim of exposed littoral sediments and littoral vegetation, indicating a recent drop in water level of ca. 30-50 cm. which is observed not to be a-typical for interannual variations in water level. C) Lakes SFL16 (upper lake) and SFL17 (lower lake) in Sandflugtdalen as seen from the west at an altitude of 500 m (August 1995). Both lakes are situated in a peat-filled former major meltwater channel, in which peat forming wet-meadow vegetation predominates. Note the eroded frontal morainic ridges and the eolian deposits (upper left corner). D) Lake NAUJG1 in the Naujaallup Nunaa area (June 1996). View is towards the northeast. White dingy for scale. In the upper right background Kangerlussuaq and the 800-1000 m high Ivnajuagtoq Plateau. E) Lake ANGU2 just north-east of the Angujaartorfik embayment (June 1996). View is towards the northwest and a white dingy for scale In the left central background Kangerlussuaq.

of the lowermost organic infill records a Holocene deglaciation before 6380 ± 100 ^{14}C yrs B.P. (UtC-1987, Van Tatenhove *et al.*, 1996).

The bathymetry of the lake is somewhat irregular, with $\bar{Z} : Z_{max} = 0.30$ and maximal 5.30 m water depth. The shallow eastern part reflects the presence of a small sandy delta lobe, now buried by organic rich sediments. Maximum sediment thickness (3.25 m) is found close to the western shore of the modern lake, which probably reflects preferential sediment focusing towards the down-wind site of the lake.

The basal part of core SFL4.1B consists of laminated sands and silts of variable thickness, lacking organic matter (< 1%). Lowermost organic-rich sediments consist of faintly laminated sandy/silty calcareous gyttja, exhibiting a gradually upward decrease in detrital mineral matter content. The uppermost 1.70 m of accumulated sediments is completely uniform and the lithological properties are similar to those of the central surficial sediments in lake SFL17. It can be described as a homogeneous silty gyttja with a detrital mineral content of up to 90% and a mean organic matter content of 18% of dry bulk weight and no carbonates. Throughout the core LOI values are highly variable, between 10-27%, superimposed on a gradually varying trend. This succession of main lithofacies is in general typical for non-glacial Greenlandic lakes (Fredskild, 1977), where glacially derived sediments are usually succeeded by finely laminated silty/clayey gyttja layers which after a couple of centuries change often rather suddenly into a 1.5-2.5 m thick completely uniform fine detritus gyttja. According to Fredskild (1977), the gradual change towards decreased minerogenic content reflects the stabilization of catchment soils by vegetation.

Sediment infill for lake SFL17 (Fig. 13C) is somewhat more complex. The lake is situated 8 km west of lake SFL4 (Fig. 1) and a basal age of 5030 ± 60 ^{14}C yrs B.P. (UtC-1586; Table 6) based on core SFL17.2 (Eisner *et al.*, 1995) suggests that glacial influence ceased 1500 yrs later than lake SFL4. Basal sediments are highly heterogeneous and range from boulder sized material to coarse and fine fluvioglacial sands. Cores penetrating into the sandy/silty substratum contained slightly peaty sands, including wood fragments and coarse terrestrial plant material, interpreted to reflect a former soil. Detailed lithological cross-sections of the sediment infill (Fig. 14) shows a succession from basal peaty sands to fine detritus gyttja which contain numerous plant macrofossil remains and are typical for modern littoral sediments. The fine detritus gyttja changes rather slowly into the extremely fine gyttja that make up the central surficial sediments of the present lake, suggesting a retraction of the littoral zone. Based on the modern composition of surficial sediments, it seems therefore reasonable to argue that the mean lake-wide sediment succession is consistent

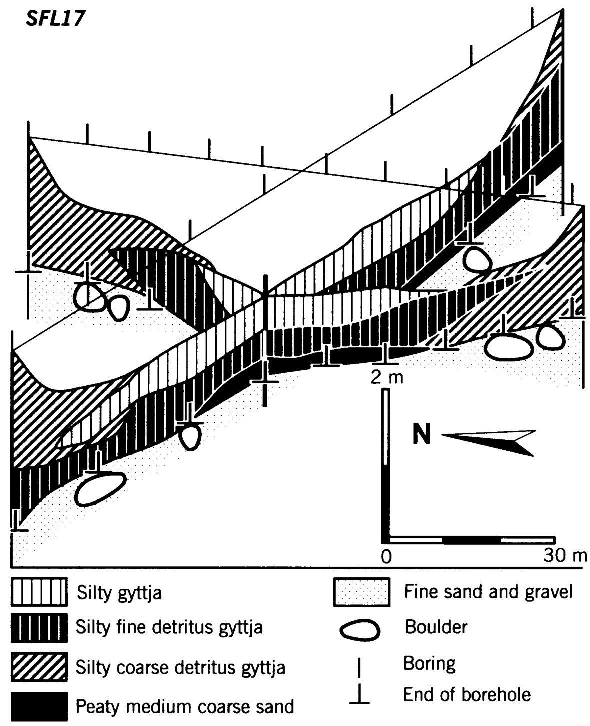


Fig. 14. Cross-sections through the central part of lake SFL17, showing the distribution of main lithofacies (for location see Fig. 10A). Most lithofacies changes are gradual. Depths are relative to water level.

Table 6. AMS ^{14}C analytical data from West Greenland lake cores.

Core no.	Depth (m)	Analysed fraction[a]	Weight (mg C)	Lab-no.[b]	$\delta^{13}C$ ‰	^{14}C age (yr B.P.)	Median (cal yr B.P.)	2σ-age range (cal yr B.P.)
SFL4.1B [d]	0.110-0.120	silty gyttja	8,15	UtC-5605	–24.2	322 ± 32	335	262- 398
– [d]	0.370-0.380	silty gyttja	1.98	UtC-3291	–21.5	700 ± 70	670	550- 790
– [d]	0.945-0.955	silty gyttja	1.28	UtC-3290	–24.4	2570 ± 70	2664	2470-2830
– [d]	1.545-1.555	silty gyttja	2.10	UtC-3289	–15.2	4470 ± 80	5103	4870-5330
– [d]	2.070-2.110	silty gyttja	0.73	UtC-1987	–21.5	6380 ± 100	7250	7030-7430
SFL4.2A[d]	0.700-0.710	silty gyttja	1.70	UtC-1988	–21.5	4460 ± 50	5088#	4930-5230
– [c]	0.700-0.710	Wf	0.55	UtC-1989	–28.4	5370 ± 60	6159	6010-6290
SFL6.1[d]	0.160-0.170	silty gyttja	1.06	UtC-3293	–23.3	900 ± 60	836	710- 930
– [d]	0.700-0.710	silty gyttja	2.00	UtC-3292	–20.0	3940 ± 100	4379	4090-4650
– [d]	1.200-1.220	silty gyttja	1.59	UtC-1990	–21.1	6090 ± 50	6954	6830-7070
SFL17.2[d]	0.060-0.070	silty gyttja	1.26	UtC-1583	–27.4	960 ± 40	880	806- 950
– [d]	0.390-0.400	silty gyttja	2.30	UtC-1584	–28.5	2590 ± 40	2692	2590-2790
– [d]	0.650-0.660	silty gyttja	0.83	UtC-1585	–28.6	3950 ± 50	4394	4250-4530
– [d]	0.910-0.920	sandy silt	1.39	UtC-1586	–24.1	5030 ± 60	5786	5650-5910
SFL17.1[c]	0.120-0.130	Bn-Cyl-Bl	0.36	UtC-5613	–27[est]	1960 ± 70	1907#	1730-2070
– [c]	0.735-0.745	Bn-Can-Hipn	0.15	UtC-5614	–27[est]	2010 ± 90	1964	1750-2190
– [c]	1.420-1.430	silty gyttja	2.34	UtC-5602	–27.1	3644 ± 42	3964#	3830-4070
– [c]	1.420-1.430	Salw	2.26	UtC-5604	–28.4	4016 ± 38	4484	4370-4570
– [c]	1.420-1.430	21xHipn	2.27	UtC-5612	–27.1	4037 ± 39	4511	4390-4610
NAUJG1.1[c]	0.112-0.122	Emn-Erf-Colf	0.19	UtC-5615	–27[est]	990 ± 90	908	730-1070
– [c]	0.924-0.930	Emn-Bn	0.17	UtC-5616	–27[est]	3830 ± 90	4225	3970-4470
– [c]	1.414-1.420	Ern-Ll	0.49	UtC-5617	–27[est]	5880 ± 50	6715	6590-6830
– [c]	1.914-1.944	Emn-Can	0.47	UtC-5618	–27[est]	8150 ± 60	9082	8910-9270
ANGU2.1[d]	0.360-0.380	Bl-Cyn-Grn	0.08	UtC-5626	–27[est]	2850 ± 130	2996	2690-3310
– [d]	0.745-0.755	Bn-Bl	0.05	UtC-5627	–27[est]	3250 ± 150	3475	3090-3830
– [c]	1.745-1.755	Bn-Bl-Salf	0.29	UtC-5628	–27[est]	4230 ± 70	4778	4550-4930
– [d]	2.607-2.615	Can-Cyn	0.11	UtC-5629	–27[est]	6710 ± 110	7545	7330-7730
SFL28.1[d]	0.700-0.720	silty gyttja	1.64	UtC-1991	–21.1	4830 ± 60	5569	5410-5690
Lille Saltsø[c]	0.320-0.360	Bl-Eml	–	AAR-3503[f]	–26.6	1070 ± 55	964	909-1077
– [c]	0.640-0.698	Bl	–	AAR-3504[f]	–25.6	2235 ± 60	2291	2082-2354
– [c]	1.120-1.160	Bl	–	AAR-3505[f]	–24.4	4280 ± 70	4849	4607-4996
– [c]	1.620-1.660	DrL	–	AAR-3506[f]	–24.8	6260 ± 70	7180	6979-7263
– [c]	1.750-1.760	Clay gyttja	–	AAR-3507[f]	–18.2	7210 ± 60	7976	7894-8099

[a]Bn: *Betula nana* nuts, Bl: *Betula nana* leaf fragments, Can: *Carex* spp. ahenes, Colf: fragments of Coleoptera, Cyn: *Cyperaceae* ahenes, Cyl: *Cyperaceae* leaf fragments, DrL: *Dryas integrifolia*, Emn: *Empetrum* endocarps, Eml: *Empetrum* leaf fragments, Erw: *Ericaceae* wood fragment, Erf: *Ericaceae* flower bud, Ern: *Ericaceae* endocarps, Grn: *Gramineae* seeds, Hipn: *Hippuris vulgaris* fruits, Ll: *Ledum* leaf fragments, Salw: *Salix* wood fragments, Wf: wood fragment undiff.; [b]UtC: ^{14}C measurements carried out at the AMS-facility of the Robert J. Van de Graaff Laboratory, Utrecht University; AAR: AMS ^{14}C Dating Laboratory, Aarhus University; [c]calibration curve smoothing 100yr; [d]calibration curve smoothing 200yr; [f]data according to Bennike (in press); [est]assumed; # rejected.

with a slowly subsiding basin. This most probably reflects the gradual desintegration of buried ice (e.g. Britt-Florin & Wright, 1969), or subsidence due to degradation of ice rich permafrost, rather than the slow infill of a rapidly isolated basin, which is often assumed when interpreting basal dates from lakes in previously glaciated areas.

Lakes dominated by carbonate-rich highly organic silty gyttja

Core ANGU2.1 was collected from the western central part of lake ANGU2 (Fig. 13E). The lake is situated at an elevation of 225 m above the eastern shore of the Angujaartorfik embayment, some 42 km west of the head of Kangerlussuaq (Fig. 1). The locality of the lake is a relatively low-relief complex of moraine ridges and large erratic boulders belonging to outer margins of the Fjord stade moraine complex (7900 to 9100 cal yrs B.P., Ten Brink & Weidick, 1974). The presence of a <20 cm-thick loess cover found on fluvioglacial/glaciomarine terraces indicates the rapid thinning of aeolian sediments with distance from the ice-marginal area. Lake ANGU2 consists of two subbasins, a small and shallow western basin and a larger eastern basin. Because of limited equipment only the shallow western part of the lake was mapped. The organic-rich sediments are up to 2.70 m thick. The maximum water depth was 1.90 m, which make the sediments liable to disturbance by lake ice (Nichols, 1967). The surficial sediments are highly aqueous and sampling of the uppermost 60 cm was extremely difficult.

The sediment matrix in general consists of carbonate-rich (up to 30%) brownish green silty gyttja with abundant gastropod shells. Total organic matter content is high (averaging 30%) and detrital mineral grains make up 50% of the sediments. The basal part of the core is composed of homogeneous, slightly humic, poorly sorted dark grey sandy silt, sharply overlain by a thin (up to 20 cm) layer of brownish fine grained strong silty gyttja with abundant gastropod remains. Carbonate content is low (> 2%) and likely derived entirely from the calcareous shells. The uppermost 2.50 m consists of indistinctly, probably disturbed, laminated carbonate-rich silty gyttja. The top 45 cm is strongly disrupted, probably as a result of coring. Downcore proportions of especially carbonates, and to a lesser extent, organic matter, are highly variable (5-40% and 25-55%, respectively), whereas the relative amount of detrital minerogenic clasts is somewhat less variable (20-45%) (Fig. 12). Relative contributions of detrital mineral matter, organic carbon and $CaCO_3$ are inversely related throughout the larger part of the sequence (Fig. 15). The components are well balanced throughout the largest part of the core, and the inverse relation is a likely result of relative dilution.

Small discontinuous sections of the sequence are composed of distinct darkbrown/yellowish white laminae. They vary in thickness and most commonly are 0.1-1.4 mm thick. Under the microscope the darkbrown layers appear as amorphous organic-rich material, containing small amounts of fine organic detritus, while the light layers are composed of coarse amorphous calcite crystals, and show some resemblance to the laminated biogenic sediments described by Kelts & Hsü (1978). Silt-sized mineral clasts are most common in the calcite laminae but also occur within the organic laminae. Individual laminae are mostly irregularly formed, wrinkled and broken and because of the many structural disturbances they are difficult to trace laterally, limiting the clear identification of separate laminae.

Laminated calcareous lake sediments are relatively common for the more saline lakes in West Greenland where meromixis and anoxic bottom waters prevent benthic activity (Anderson et al.,

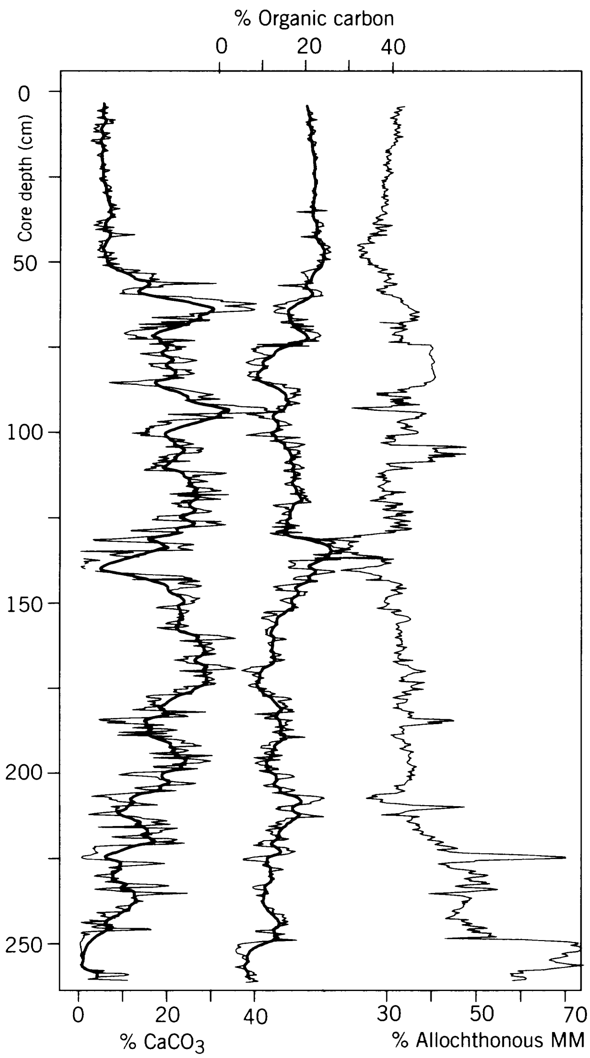


Fig. 15. Profiles of $CaCO^3$, organic carbon and allochthonous mineral matter in sediments from lake ANGU2. Organic carbon calculated using conversion factor of 0.47 for % LOI (Dean, 1974; 1999).

1999; 2000). Sedimentation of calcareous gyttja has been reported in North Greenland (Fredskild, 1977), where carbonate rocks are widespread. For the West Greenland lakes the causal process remains to be clarified, but most likely relates to calcite deposition in strongly alkaline lakes due to photosynthetically induced inorganic precipitation associated with CO_2 consumption and increased pH during phytoplankton blooms (Kelts & Hsü, 1978).

Lakes dominated by highly organic diatomaceous gyttja, lacking aeolian input

Lake NAUJG1 is a very small (0.56 ha) roughly bowl shaped lake (Fig. 13D), formed in a small peat-filled depression. It is situated 70 km southwest of Kangerlussuaq Airport (Fig. 1), close to the eastern spurs of the coastal mountains and in the vicinity of Sukkertoppen Iskappe (Naujaallup Nunaa area). Relief extents up to some 1000 m a.s.l. The lake is situated at an elevation of 300 m a.s.l., directly above the northern shore of Kangerlussuaq, and in the westernmost part of a hanging valley. The locality of the lake is just on the inner limits of the Sarfartoq/Avadtleq moraine system (8600 to 9800 cal yrs B.P., Ten Brink & Weidick, 1974). Evidence for mass wasting, mainly gelifluction, is common on the western slopes of the lower valley floor, but is absent in the immediate surroundings of the lake. Western shorelines are bounded by a small mire in front of a 600 m high moderately steep convex bedrock slope, while southern and southeastern shorelines are dammed by low relief bedrock. A low ridge of presumably glacial sediments acts as a threshold for the northeastern shore of the lake. No active tributaries, except a very small inactive ephemeral stream from the western slope, enter the lake.

No aeolian silt cover was found in the wider surroundings of the lake, and it seems that atmospheric input of detrital mineral grains is absent. Core NAUJG1.1 was collected from the central part of the lake (Fig. 8). Bathymetry is quite simple, with $\bar{Z} : Z_{max} = 0.37$ and maximal 3.15 m water depth. Accumulated sediments are regularly distributed over the lake floor, with maximum sediment thickness (2.20 m) found in the central parts of the lake. The basal part of the core consists of compacted dark grey silts with only a few minor sandy layers and no organic matter. These sediments are sharply overlain by a thin (up to 20 cm) layer of brownish silty gyttja with overall rapidly increasing LOI values (from < 1 to 30%). The uppermost 2.10 m of accumu-

Table 7. Mean sediment accumulation rates (mean values ± 1σ).

	Sediment accumulation rate[e]					
	SAR[a] *(cm/100 yr)*	LOI *(%)*	DBS[b] *(g m^{-2} yr^{-1})*	TOM[c] *(g m^{-2} yr^{-1})*	$CaCO_3$ *(g m^{-2} yr^{-1})*	TMM[d] *(g m^{-2} yr^{-1})*
SFL4.1A	2.88 ± *24 %*	18.3 ± *4.4*	72.4 ± *20.3*	12.8 ± *3.1*		59.6 ± *19.0*
SFL17.1	3.17 ± *21 %*	23.7 ± *5.8*	64.4 ± *20.2*	14.9 ± *4.9*		53.5 ± *8.2*
SFL17.2	1.58 ± *11 %*	23.6 ± *7.5*	46.6 ± *7.8*	9.1 ± *1.9*		37.8 ± *8.1*
SFL6.1	1.74 ± *18 %*	11.6 ± *5.6*	49.9 ± *13.8*	4.6 ± *1.6*		50.4 ± *15.3*
ANGU2.1[f]	11.1 ± *15 %*	33.6 ± *1.6*	104.2 ± *46.0*	32.6 ± *13.1*	17.3 ± *15.5*	58.7 ± *20.6*
NAUJG1.1	2.21 ± *11 %*	36.1 ± *3.6*	36.7 ± *9.6*	13.1 ± *2.8*	9.1 ± *2.3*[g]	23.6 ± *6.8*

[a]Sedimentation accumulation rates. Values for maximum or minimum sedimentation rates relative (%) to overall mean. Maximum and minimum sedimentation rates were calculated for sections between two successive ^{14}C-ages using +2σ and –2σ for maximum slope, and –2σ and +2σ for minimum slope; [b]SAR dry bulk sediment; [c]SAR total organic matter; [d]SAR total minerogenic matter; [e]mean sediment accumulation rates (±1σ) calculated from concentration data (g/cm^3) and mean sedimentation rates; [f]calculations for upper three ^{14}C-ages only; [g]biogenic silica (uppermost 45 cm only).

lated sediments is completely uniform, apart from a thin (3 cm) intercalated silty layer at 2.0 m. Under the microscope, grain sizes seem highly uniform throughout the core, and make up 50-55% of the sediment matrix. Mean grain size ranges from 8-32 µm (Fig. 11), which is finer than those typically found for aeolian silts around the head of Kangerlussuaq (Dijkmans & Törnqvist, 1991). Amorphous extremely fine flocculent organic material and abundant diatom cells (up to 20% of bulk dry weight, see also Fig. 18) together make up 35 to 50% of the sediment matrix and the complete upper 2.10 m of the sediment profile must be characterized as a diatomaceous gyttja. Coarse particulate organic matter is extremely sparse and macrofossil remains are in general absent. Similar to the other lakes, LOI properties are highly variable throughout the core superimposed on a more gradually varying trend.

Radiocarbon chronology

Table 6 presents ^{14}C age determinations for six of the nine lake sediment profiles. Calibration results are reported as 2σ (95%) calendar-age ranges, with median values derived from the cumulative calibration probability distributions. Including previously published ^{14}C ages (Eisner *et al.*, 1995; Bennike, 2000), there are 33 AMS ^{14}C ages constraining the stratigraphic studies. The distribution of ^{14}C ages over depth suggests fairly uniform accumulation rates, and linear interpolation between successive median values was used to extrapolate ^{14}C ages towards intermediate depths (Fig. 16). Uppermost ^{14}C samples are not located at the top of sediment cores because of the risk for sediment disturbance. The presence of in-situ aquatic mosses and oxidized sediments (typically 2-3 cm thick) was taken as an indication for the recovery of the complete water-sediment interface. Whenever absent, the age of the uppermost deposits was obtained by extrapolation. According to the age-depth curves, uppermost sediments of cores, NAUJG1.1, SFL6.1 and ANGU2.1 were lost during coring. For core ANGU2.1, about 2800 yrs of sediment is missing, which is likely due to a poor recovery of the uppermost sediments, and/or lake ice disturbance.

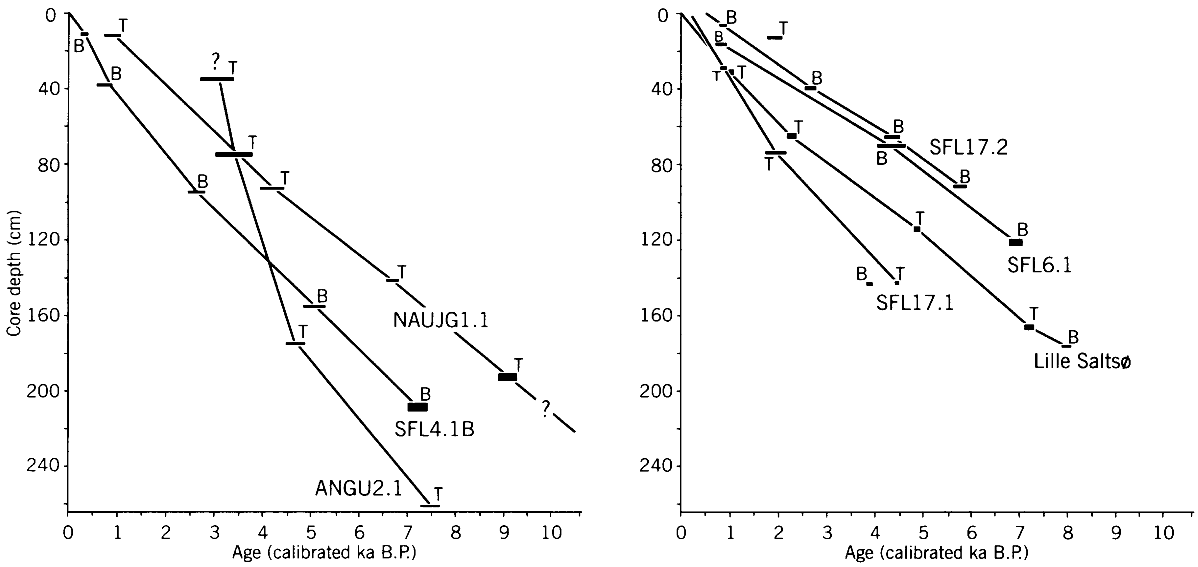


Fig. 16. Age-depth plot of Kangerlussuaq lake cores. Data for Lille Saltsø (Bennike, 2000). Solid boxes indicate calibrated 2σ ^{14}C age ranges and sample thickness. T: terrestrial plant macrofossils, B: bulk sediment. Age-depth models based on linear interpolation between the median values of the calibration probability distribution.

Long-term sedimentation rates are typically on the order of 1.6 to 3.2 cm/100 yrs (Table 7), comparable to other nonglacial lakes in Greenland (Fredskild, 1977; Bennike, 2000) and the Canadian Arctic (Doran, 1993). Given the large confidence limits for the ^{14}C-age determinations of core ANGU2.1, and the probable loss of uppermost sediments, sedimentation rates for this sequence remain somewhat ambiguous. The lowermost date for core ANGU2.1 suggests a significantly lower sedimentation rate for the first 0.90 m of the sequence. A hiatus is possible, however, given the sharp lithological break between the lowermost two ^{14}C samples.

Basal ^{14}C-ages of the lakes only yield minimum ages for the timing of local deglaciation since the samples were taken from the organic rich sediments. ^{14}C ages for the start of organic infill in the lakes in general follow the timing of deglaciation proposed by Ten Brink & Weidick (1974) and Van Tatenhove *et al.* (1996). However, the dataset also shows inconsistencies. For core NAUJG1.1 ^{14}C ages for terrestrial macrofossils were obtained at a level 20 cm above the lowermost organic deposits. The calibrated age-range of 8910-9270 yrs B.P. is well in range of the dates obtained for the period of deposition of the Sarfartooq/Avadtleq moraine system (8600-9800 cal yrs B.P.) although extrapolation suggests an age of 10,500 cal yrs B.P. for the basal organic sediments, which is significantly older. A ^{14}C age of 7545 cal yrs B.P. constrains the cessation of glacial influence in lake ANGU2.1 during the Fjord-stade (7900-9100 cal yrs B.P., Ten Brink & Weidick, 1974). If correct and given the uncertainties in the chronology proposed by Ten Brink & Weidick (1974), the onset of the Fjord-stade would shift to a younger date.

Sediment concentration and sediment accumulation rates

Relative increases in organic matter content can be attributed to several factors that are either related to within-lake processes (autochthonous), or derived from outside the lake proper (allochthonous). Accumulation of specific components can vary independently over time and their effects are difficult to address properly without quantified data and specific sediment accumulation rates. To determine the variable contributions, volume based concentrations (total dry density

(TDD) in mg/cm^3) of dry bulk sediments (DBD), total organic matter (TOM), total minerogenic matter (TMM), calcite and biogenic silica (BioSi) were calculated. The results are plotted for cores NAUJG1.1, SFL4.1A and SFL49.1 in Fig. 17.

All components are highly variable throughout the cores, whereas long-term variations at the base of the LOI profiles largely reflects increased minerogenic input. Specific sediment accumulation rates (SAR, g m^{-2} yr^{-1} ± 1σ) are presented in Table 7, and were calculated by multiplying TDD values with the mean sedimentation rates derived from the age-depth curves. Mean concentration values for organic matter in the sediment cores vary between 25 and 100 mg/cm^3 with a weighted mean SAR between 4.6 ± 1.6 g m^{-2} yr^{-1} (SFL6.1) and 14.9 ± 4.9 (SFL17.1). SAR values of organic matter for core ANGU2.1 are higher, measuring 32.6 ± 13.1 g m^{-2} yr^{-1}. This result is due to the higher apparent sedimentation rates in this lake. Bulk dry densities are highly correlated to concentrations of minerogenic matter, which for lakes in Sandflugtdalen vary between 100 and 300 mg/cm^3, with increased values towards more than 500 mg/cm^3 at the base of the cores. TDD values for minerogenic matter in core NAUJG1.1 are somewhat less and average 55 mg/cm^3. Weighted mean SAR values for minerogenic matter are typically in the order of 50 to 60 g m^{-2} yr^{-1} for lakes within the area covered by aeolian silts (except for core SFL17.2), which points to a regionally homogeneous influx rate of aeolian silts. Allochthonous clastic input in lake NAUJG1 is distinctly less and averages 23.6 ± 6.8 g m^{-2} yr^{-1}.

From Table 7 it is quite clear that accumulation rates for minerogenic matter are as variable as the accumulation rates for organic matter. Both sediment components are intrinsically related through relative dilution and bioturbation effects present in each sample. Comparison of accumulation rate (flux) data of homogenized sediments will not circumvent the interpretation problem of the percentage data. This is mainly a result of the inherent uncertainty when calculating SAR-values using ^{14}C-based chronologies. This can easily be seen from a comparison of means and standard deviations of SAR and SAR values

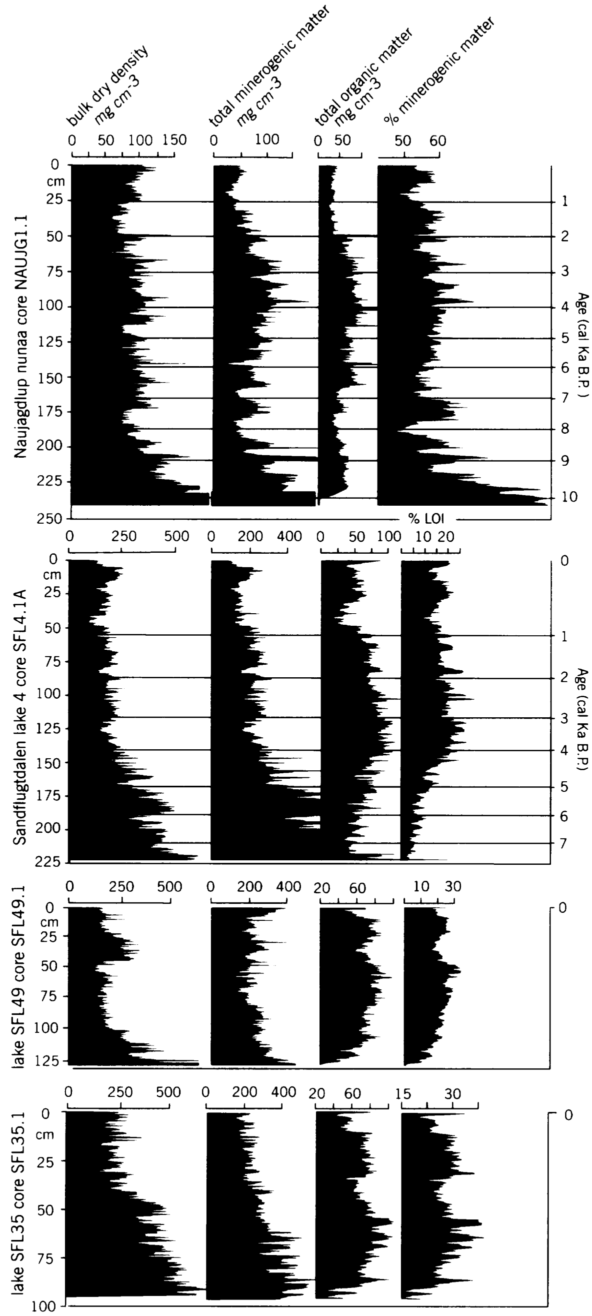


Fig. 17. Comparison between concentration diagrams for lithological properties and %LOI of lake cores SFL35.1, SFL49.1, SFL4.1A and NAUJG1.1 plotted vs. core depth (left axis) and time (right axis).

Table 8. Regression results between LOI and concentration data.

Core no.	OM[a] (g/cm^3)	MM[b] (g/cm^3)	DBD[c] (g/cm^3)	OM= *f*MM[d]	n[e]
SFL4.1A	0.68	–0.07	0.53	0.41	424
SFL49.1	0.67	–0.34	0.02	0.12	297
SFL17.1	0.74	0.26	0.51	0.30	447
SFL17.2	0.76	–0.81	0.41	0.34	57
SFL6.1	0.58	–0.43	–0.15	–0.20	362
NAUJG1.1	–0.13	–0.66	–0.78	0.51	587

[a]r-values calculated using %LOI and total organic matter; [b]LOI and total minerogenic matter; [c]LOI and dry bulk density; [d]r-values between organic matter and total minerogenic matter; [e]number of paired samples.

for TOM and TMM in table 7. To circumvent these problems, specific influences of either variable TOM or TMM concentrations, potential influences of variable SAR, and compaction on concentration and percentage data, were analyzed by calculating linear regression coefficients between bulk dry density weights, percentage data and specific TDD values (Table 8). Because the purpose was to analyse the small stratigraphic changes, the lowermost core sections were omitted from the datasets to avoid biased results. According to the regression results, concentration changes in TOM for the SFL-sites explain 45% to 57% of the overall variance in LOI values, whereas variations in TMM concentrations explain only 1% to 18%, suggesting that changes in concentration TOM drive the LOI fluctuations at the SFL sites. Positive relationships between DBD and TDD values for TOM indicate the effect of either variable sedimentation rates, compaction, or sample volume errors. These are in general weak except for a strong inverse relationship between TOM and DBD in core NAUJG1.1.

For core NAUJG1.1 from the westernmost lake NAUJG1, changes in TDD values of TMM account for 44% of the variance in %TMM, whereas there seems to be no direct relationship between LOI and organic matter. According to the data there is a weak positive relation between TOM and TMM for this lake, which explains up to 25% mutual variance. The sediments of NAUJG1 contain abundant diatom cells throughout the whole core, contributing up to 20% of dry bulk sediment weight. To analyze the effect of changing concentrations of diatom cells on the LOI results, the uppermost 45 cm of core NAUJG1.1 was selected as a dataset for the calculation of a linear regression between BioSi, total minerogenic matter, autochthonous mineral matter and organic matter. The correlation coefficient between biogenic silica and total bulk minerogenic matter is r=0.66 (α=0.05, 54 paired samples), explaining 44% of the variance in total minerogenic content (Fig. 18A). No significant correlation was found between variations in non-BioSi minerogenic matter and LOI, suggesting that variations in the minerogenic fraction throughout the core (residuals-on-ignition; ROI) are driven by variations in biogenic silica (Figs 18B-C). The weak positive relation between TOM and TMM concentrations suggests covarying fluctuations of organic matter and BioSi, where the dominance of biogenic silica changes over organic matter in the LOI signal can be explained by the higher specific weight of diatoms compared to the nonsiliceous biogenic fraction.

It is important to note that, considering the available data, there is only weak evidence for a dominating aeolian signal in the Sandflugtdalen lakes, irrespective of the recognition that the bulk of minerogenic sediments have a primary aeolian origin. Relative dilution of organic matter due to increased atmospheric inputs should in theory give a high negative relationship between TDD values for TOM and TMM, but no such evidence was found except for a weak signal in core SFL6.1. Nevertheless, small parts of the core sections might be influenced by increased aeolian influx, as it seems unlikely that both source and supply of aeolian silts have remained more or less invariable since deglaciation (Dijkmans & Törnqvist, 1991). Lack of a dominant aeolian signal is also evident for core ANGU2.1 (Fig. 15) which shows a more invariable allochthonous minerogenic component in the lake with depth when compared to the organic and carbonate content. This is somewhat surprising given the large aeolian component in lake sediments around the head of Kangerlussuaq and contrasts earlier postulations on the type of sediment variability present in the lakes (Eisner *et al.*, 1995).

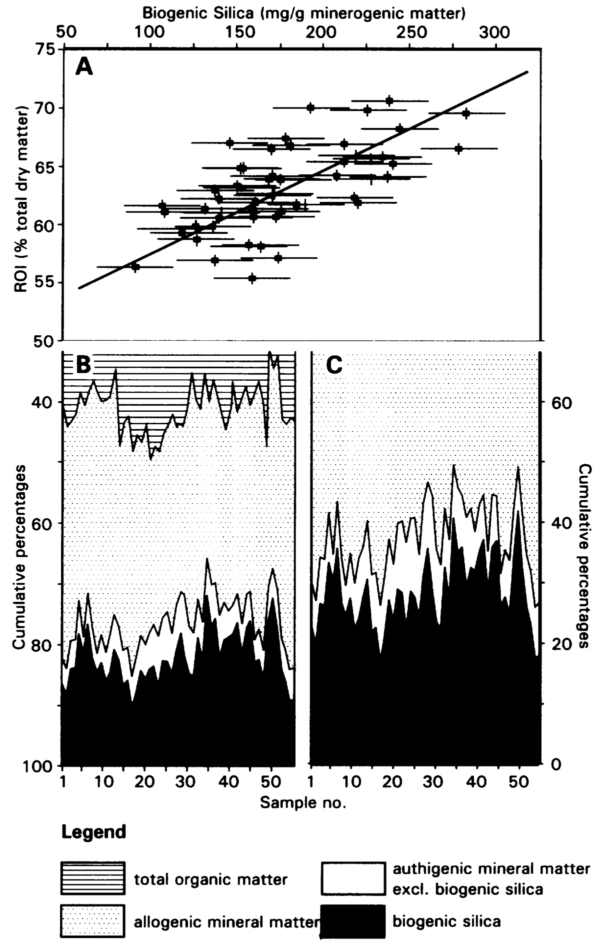


Fig. 18. A) Scatterplot including analytical errors for biogenic silica content (mg/g total minerogenic matter) vs. residuals of ignition (%ROI of bulk dry matter). B) Cumulative percentages of dry bulk matter of sediment components in core NAUJG1.1. C) Cumulative percentages of total minerogenic matter of sediment components in core NAUJG1.1. Selected section represents 0-45 cm core depth.

Diatom stratigraphy

Results of diatom analysis for the central core (SFL4.1B) of lake SFL4 have been synthesized in the form of a percentage diatom diagram, where relative diatom abundances are based on the sum of all counted valves. Fig. 19 shows the relative abundances of major taxa (>1% presence of total sum) in this core, and presents the first diatom stratigraphy from this part of West Greenland. The chronostratigraphy for this core is based on the age-depth curve presented in Fig. 16. When analysis started, the uppermost 40 cm of the sequence was incomplete and the stratigraphy ends at ca. 900 cal yrs B.P. Specification of ecological preferences is based on autoecological information summarised by De Wolf (1982), Denys (1992), with additional information from Cumming & Smol (1993), Dixit *et al.* (1991), and Pienitz *et al.* (1995).

Fossil diatom assemblages record significant floristic shifts and have been grouped into six biostratigraphic zones to discuss the inferred limnological changes:

- Barren sediments (-218 cm, older than 7550 cal yrs B.P.)

No diatom valves were found below 218 cm. This could be due to dissolution, but more probably conditions were to harsh to develop a diatom flora.

- Diatom zone I (218-213 cm, 7550-7400 cal yrs B.P.).

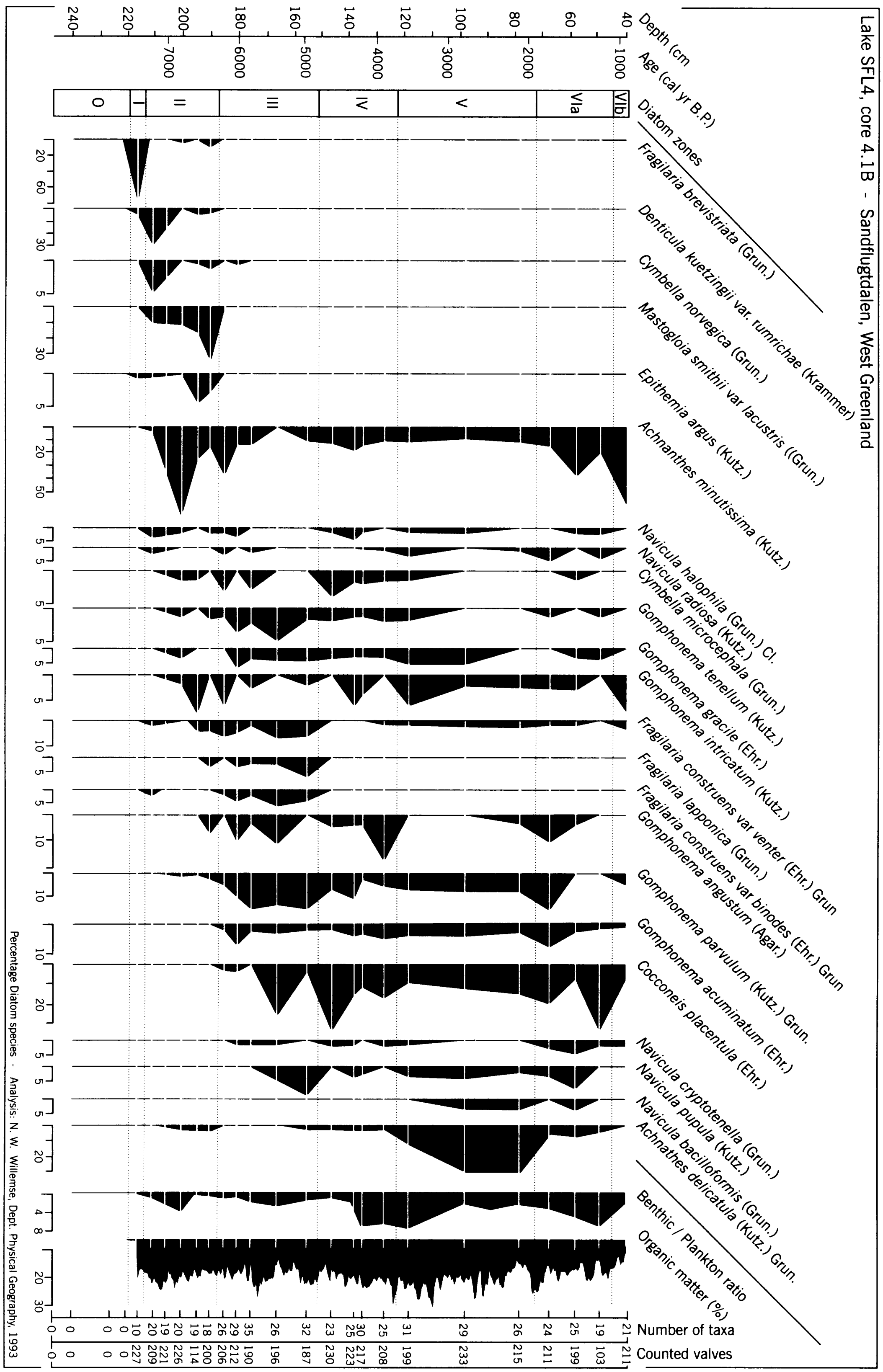


Fig. 19. Diatom percentage diagram for core SFL4.1B.

Both diatom zones I and II represent the transition from glacio-lacustrine to lacustrine conditions. First species to appear in the sediment is *Fragilaria brevistriata* (78%) that dominates the single sample from zone I, together with *Denticula kuetzingii* (11%), small *Fragilaria* spp. (5%), *Epithemia argus* (3%) and *Navicula pupula* (5%). Most species are alkaliphilous and conditions must have been alkaline, possibly caused by the leaching of the barren soils surrounding the lake (Fredskild, 1977). Whereas *F. brevistriata* has a wide nutrient tolerance, other species in this sample suggest moderately high nutrient availability during this phase.

- Diatom zone II (213-187 cm, 7400-6000 cal yrs B.P.).

Diatom zone II shows an abrupt change towards a more ion-rich environment. The fresh/brackish *Fragilaria brevistriata* disappears and *Denticula kuetzingii* var. *rumrichae* becomes dominant (32%) together with *Cymbella norvegica* (22%), and *Mastogloia smithii* var. *lacustris* (13%). Most remarkable is the appearance of mesohalobous species such as *Navicula halophila* (3%), *Anomonoeis exilis* (1%), *Amphora veneta* (5%), and *Mastogloia smithii* var. *lacustris* (13%) that tolerate salinities from 0.3 to about 10‰. The presence of planktonic species such as *Cyclotella striata* (2%) and *Aulacoseira italica* (1%) suggest more favourable conditions for open water species to flourish, although the frequencies are very low. Increased erosion of surrounding soils, possibly bringing dissolved mineral salts towards the lake, is indicated by the presence of small bands of sand and pebbles and species composition changes from a dominant epiphytic one, towards a more benthic flora.

- Diatom zone III (187-150 cm, 6000-4900 cal yrs B.P.).

This zone is considered to be the pioneering phase of a closed-basin lake as several pioneering species like small benthic and alkaliphilous *Fragilaria* spp. appear in this section. Mesohalobous species disappear, and a more diverse flora develops containing several *Achnanthes* spp., *Fragilaria* spp., *Cocconeis* placentula and several *Navicula* spp. Dominant species in this zone are *Achnanthes* spp. especially A. *minutissima* (40-18%) and *Cocconeis placentula*, together making up about 30-35% of total valve counts. Several alkaliphilic *Amphora* spp. appear together with *Cymbella* spp. (C. *cymbiformis*, C. *helvetica*, C. *microcephala*) and *Gomphonema* spp., indicative of nutrient-rich conditions. The appearance of small benthic *Fragilaria* spp. (F. *construens* var. *venter* and var. *binodes*) indicate clear water conditions. The sudden increase in the eutrophic and alkaliphilous *Cocconeis placentula* and the decrease or almost disappearance of the meso-eutrophic, pH-indifferent *Achnanthes minutissima* (at 167 cm) might be influenced by higher trophic conditions and higher alkalinity as is indicated by the presence of eutrophic alkaliphils as *Fragilaria* spp. and *Gomphonema* spp. Planktonic species as *Melosira* spp. and *Cyclotella* spp. disappear, suggesting less favourable conditions for open water species.

- Diatom zone IV (150-123 cm, 4900-3800 cal yrs B.P.).

Most *Fragilaria* spp. have disappeared in this zone. Dominant species are the epiphytic *Cocconeis placentula* (35%) and *Achnanthes minutissima* (14-20%) among with several other *Achnanthes* spp. (A. *hauckiana*, A. *exilis*), which are indicative of alkaline conditions. A change of clear water *Fragilaria* spp. towards *Synedra* spp. as S. *acus*, S. *ulna* and S. *rumpens*, and the increase in benthic *Navicula* spp. suggest a larger input of suspended clastics. Epiphytic taxa in this stage again make up some 70% of the total count. Benthic species make up some 12% and very few planktonic species occur.

- Diatom zone V (123-73 cm, 3800-1900 cal yrs B.P.).

This zone is dominated by the euryhaline *Achnanthes delicatula* (30%), A. *minutissima* (12%) and *Cocconeis placentula* (10-15%). Species composition remains remarkably constant throughout this zone. *Achnanthes delicatula* is most commonly found in slightly to moderately brackish water which is consistent with the presence of the halophobous *Navicula halophila*. Species composition also shows a marked shift from mesotrophic to eutrophic conditions as A. *delicatula* is thought to be an eutrophic species. Epiphytic species make up some 75% of total species in this zone. The transition to zone VI is characterized by the virtual disappearance of *Achnanthes delicatula* and a renewed increase in *Achnanthes minutissima* and *Cocconeis placentula*.

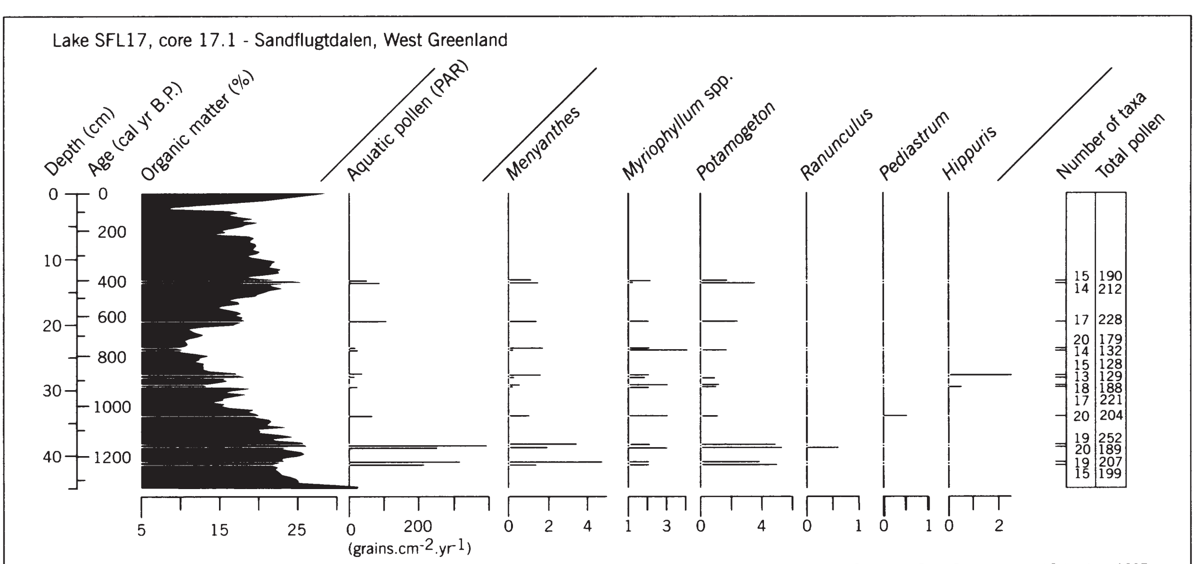
- Diatom zone VIa (73-45 cm, 1900-1100 cal yrs B.P.).

This zone shows some resemblance with zone III with *Achnanthes minutissima* and *Cocconeis placentula* being the dominant species (50%). Benthic and aerophilous species as *Navicula* spp. and *Gomphonema* spp. become more frequent coinciding with the disappearance of *Ranunculus* and *Hippurus* pollen from the sediments (unpublished data). At 60 cm depth, large benthic *Navicula* spp. appear. This coincides with lower counts of *Gomphonema* spp. and *Cocconeis* spp. and increasing numbers of alkaliphilic *Achnanthes* spp. and *Epithemia* spp. The transition to the last zone starts with a rapid increase of non-organic components as can be seen from the LOI curve.

- Diatom zone VIb (45-50 cm, 1100-900 cal yrs B.P.).

Although based on one sample only, this zone can be distinguished because of the marked change in species composition. The pH-indifferent *Achnanthes minutissima* becomes strongly dominant (61%) whereas alkaliphilic species as *Cocconeis* placentula and *Epithemia* spp occur with a very low frequency. Hydrophyte pollen are sparse. Low frequencies of *Hippurus* pollen occur, whereas most other hydrophyte pollen have disappeared (unpublished data). A markedly low occurrence of hydrophytes at this stage in time (after ca. 1000 cal yrs B.P.) is in accordance with pollen and macrofossil data from lake SFL17 (Eisner *et al.*, 1995, see also Fig. 20). The disappearance of major hydrophyte species in both lakes

Fig. 20. Percentage diagram for aquatic pollen and spores, and organic matter content in the uppermost 40 cm of core SFL17.1. The pollen sum includes both aquatic and terrestrial pollen, but excludes long-distance pollen. Pollen accumulation rate (PAR) for aquatic pollen and Pediastrum expressed as grains cm^{-2} yr^{-1}.



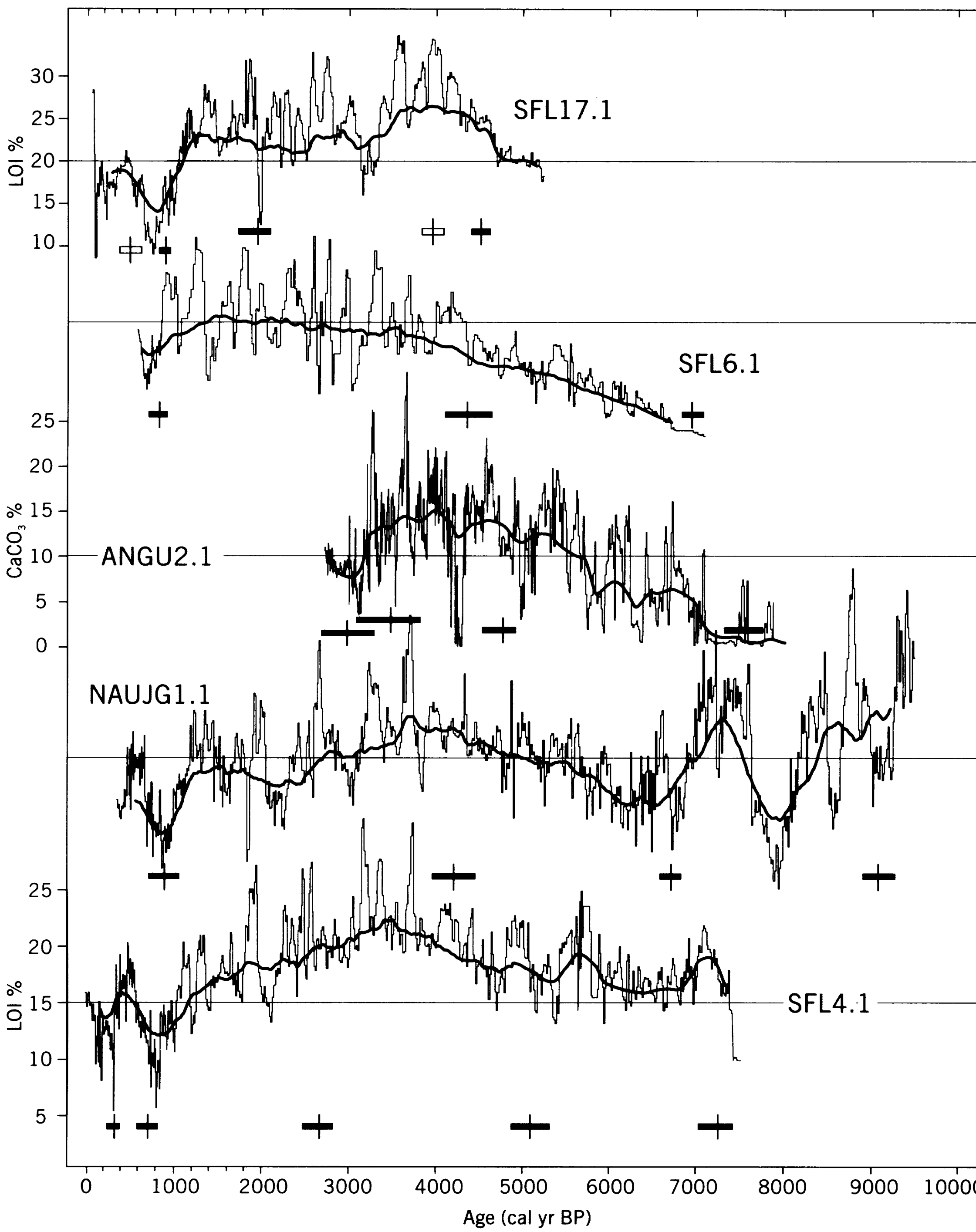


Fig. 21. Chronostratigraphic comparison of Kangerlussuaq lake sedimentary sequences. Solid boxes indicate 2σ-age range and median value of ^{14}C ages. Heavy lines are ~500 yrs equal-weight averages.

at the same time suggests unfavourable conditions for water plants to grow in the lake, which could relate to shortening of the summer open-water period and/or lowered temperatures. A potential deterioration of climate after 1000 cal yrs B.P. is also suggested from the LOI data and a pollen record from a nearby site (Eisner *et al.*, 1995).

Regional correlations

According to regression results for sediment concentration data, lithological

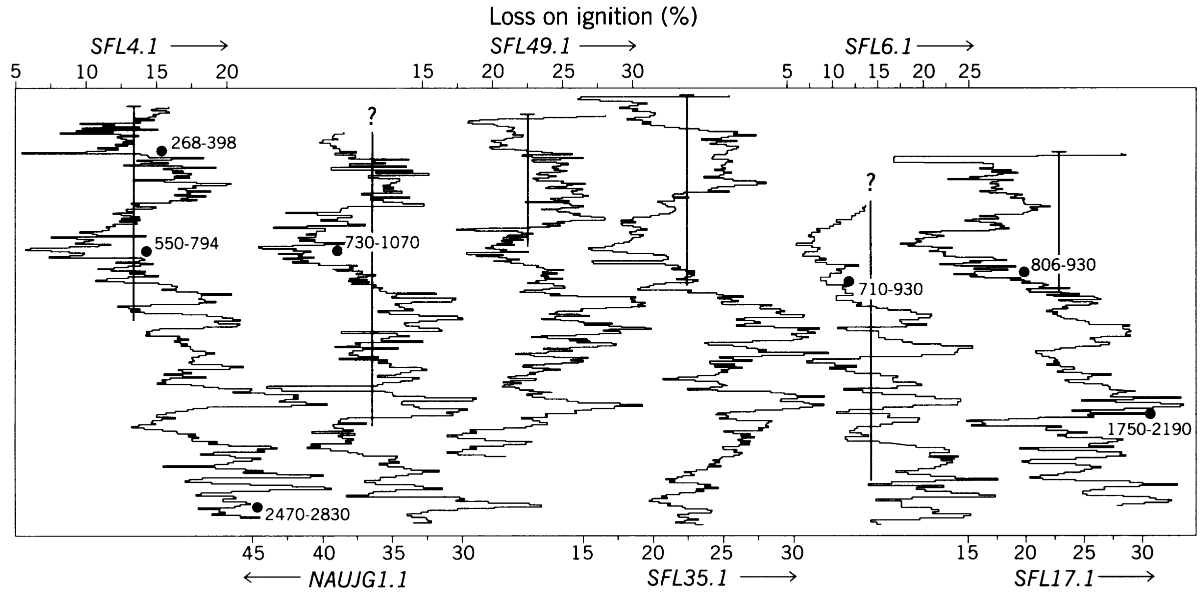


Fig. 22. LOI data from uppermost (50–100 cm) sediments in six West Greenland lakes plotted vs. depth. Vertical bars represent 50 cm core length, for scale. Question-marks indicate the absence of water sediment interfaces. Note the inverted scale for NAUJG1.1. Calibrated ^{14}C ages indicated by 2σ confidence intervals. Open circles represent bulk sediment samples; solid circles represent terrestrial macrofossils.

variations in the shallow lakes seem to be primarily driven either by changes in organic matter or biogenic silica content. Together with concurrent changes in major hydrophyte pollen, and evidence from limnic plant remains (see Willemse & Törnqvist, 1999), these observations favour the interpretation of a predominant lacustrine signal in the LOI records from small and shallow Kangerlussuaq lakes.

Fig. 21 presents the chronostratigraphic comparison of 5 different LOI profiles. For each individual curve, parameters were plotted to explain sediment variability in terms of lacustrine productivity changes. Lithological fluctuations exhibit the same order of high frequency variability, superimposed on a gradually varying trend. In terms of timing, there seems to be good agreement between short term fluctuations, especially for SFL4.1 and NAUJG1.1. Fig. 22 shows the lithostratigraphic correlation of LOI curves for the uppermost 50-100 cm (time window approximately 0-2200 cal yrs B.P.). Taking into account the before mentioned restrictions, similarity of individual LOI profiles between the lakes is surprisingly good. All curves show similar and robust fluctuations (in terms of timing, shape, and amplitude) and stratigraphically pronounced excursions (notably around 800 cal yrs B.P. and 2100 cal yrs B.P.) are within the 2σ-confidence limits of the calibration probability distributions. Two records (SFL49 and SFL35) have not been dated. As a chronostratigraphic tie-point with the other cores 1) the presence of in-situ aquatic mosses and a thin 2 cm layer of oxygenated sediment indicating the modern water-sediment interface, and 2) the prominent excursion in the LOI profile were used. Assuming overall similar sedimentation rates, periods of increased or decreased LOI seem to be well in phase, suggesting a possibly regional control on lake productivity (Willemse & Törnqvist, 1999). Distinct optimum phases are recorded between ca. 4500-3500 cal yrs B.P. in lakes SFL17, NAUJG1, and SFL4, which conforms to the 5000-3700 cal yrs B.P. 'climatic optimum' inferred from West Greenland pollen records (Fredskild, 1985; Eisner *et al.*, 1995).

Long-term LOI trends are more difficult to interpret. Temporal changes for slowly increasing LOI values from the basal sediments upward are highly lake specific, suggesting interference with more locally forced patterns of variation. Deviating long-term trends (e.g. SFL6.1 and SFL17.1) may be related to natural infilling and the associated shallowing of the lake, leading towards progressively increasing organic deposition rates at the coring site. It should be noted that in some cases (notably SFL4.1 and NAUJG1.1.) longer term fluctuations do exhibit similar trends.

Discussion

Comparative radiocarbon dating of terrestrial and semi-aquatic plant material suggests that reservoir effects in limnic organic material is absent in the shallow lakes, which is consistent with results earlier obtained for a small salt lake at the head of Kangerlussuaq (Bennike, 2000). Datable macrofossil material is, however, rare in lake cores from central areas, and organic-rich bulk sediment samples had to be used in numerous cases. The similarity between single-lake core stratigraphies dated with bulk sediment samples and terrestrial macrofossil material furthermore suggest that ageing of bulk organic material due to reservoir effects or old carbon can be neglected. The absence of ageing effects can be attributed to the rapid water renewal in these shallow lakes and the dominant within-lake source for sedimentary organic matter. Root contamination of bulk sediments in sediment cores from the marginal parts of the lakes is evident from comparative dating with macrofossil material, where bulk sediment samples yield ^{14}C ages that are generally too young.

Overall sediment accumulation rates as derived from extrapolated ^{14}C ages are rather constant for individual cores, but differ from 1.4 to 3.9 cm/100 yrs between lakes. For individual samples, however, the temporal resolution is dependent on small changes in sedimentation rate and the amount of water-sediment interface mixing, like bioturbation. Oxygenated sediments at the water-sediment interface set the limit for benthic activity and sediment mixing due to bioturbation, which is largely dependent on burial rates and the oxygen content of the overlying bottom and pore waters. Depth of benthic activity determines the amount of temporal smoothing in the records, which should be accounted for when comparing homogenised lake records to other time series (Willemse & Törnqvist, 1999). Moreover, it is of consequence for radiocarbon ages, both derived from bulk samples and macrofossil material, where the 'sample time-width' is a function of both sample thickness and depth of sediment mixing (Törnqvist & Bierkens, 1994).

Mean accumulation rates of sedimentary organic matter is rather variable between lakes (4.6-32.6 g m^{-2} yr^{-1}), whereas for individual lakes in the direct vicinity of the Inland Ice margin, influx rates of aeolian derived minerogenic matter varies less (between ca. 50-60 g m^{-2} yr^{-1}, except for core SFL17.2), which is consistent with a regionally homogeneous atmospheric input.

According to statistical analysis of concentration data proportional changes in general reflect variable contributions of biogenic matter, whereas simple dilution of the sediments due to variable increases of either organic matter or atmospheric silts play a more subordinate role. Nevertheless, observed changes over time of the amount of organic matter left in the sediments should be interpreted cautiously. Accumulation of organic matter is the cumulative result of a variety of processes, such as net primary production, grazing by zooplankton, terrestrial erosion, bacterial mineralization and sediment diagenesis. Only a small percentage of organic matter produced in the water column is deposited (most is recycled), and then only a fraction of deposited organic C is actually preserved (Meyers & Lallier-Vergès, 1999). Moreover, changing accumulation rates influence the preservation potential of organic compounds and hence, in addition to bioturbation, affect the proportion of organic matter in the sediments. Sources and diagenetically alterations of organic matter vary both in space and time and accumulation records are therefore usually quite unique.

Notwithstanding these uncertainties, the observed synchrony among a regional series of detailed lake LOI records provides a strong case for a systematic climate control on lake biological processes (Willemse & Törnqvist, 1999). Aquatic organisms respond to parameters directly controlled by climate such as temperature, light, and watercolumn turbulence, all of which are most rapidly transmitted in small systems. Moreover, covarying changes in the length of the ice cover period and open-water growing season can have effects on regional lake productivity. During colder years the short active biological period can be reduced with 30% or more (Smol, 1988; Doran *et al., 1*996) which would depress net primary production. Warming is accompanied by a shorter period of snow and lake ice cover (Magnuson *et al*., 2000), resulting in enhanced plant growth due to lengthening of the growing season (Myeni *et al*., 1997; Quayle *et al*., 2002), improved exchange of gases and nutrients, wind-driven circulation and light conditions (Schindler *et al., 1*974). The timing for the start of the open water season largely reflects air-temperature variability in the transition seasons but such climatic influences on lakes are extremely complex and include a diversity of direct and indirect effects.

Alternatively, the presence of saline lakes in the area indicate changes in the hydrological balance, and effects of water-level changes could equally influence the nutrient status of the lakes. However, given the amplitude of changes in organic matter content, such water-level changes would be visible as distinct lithofacies changes in the cores (Harrison & Digerfeldt, 1993), but no such indicators are present. Equally, other lake nutrient transfer mechanisms related for instance to spring thaw and runoff (Miller, 1992), and variable hydrological and pedological settings (Engstrom *et al*., 2000), could play another important role. The proposition that lake productivity in these shallow isolated systems is influenced directly by climate parameters therefore requires a more complete analysis of palaeolimnological productivity proxies, and comparison with other records of regional environmental change.

Notwithstanding the good agreement between individual lake records, long-term trends also indicate interference from other factors, like natural lake infill, as the pattern of sediment accumulation and biological activity is sensitive to the constantly changing morphometry of the infilling lake (Håkanson & Jansson, 1983). The recognition of patterns of variation on different timescales points to the importance of comparing separate lake records in order to identify and remove anomalous stratigraphic variability (Snowball & Sandgren, 1996). Constructing a composite record would greatly help to remove anomalous variability, following similar practices used to evaluate other high resolution proxy records such as ice-cores and tree-rings. This would however require stratigraphical alignment well below the resolution level of ^{14}C ages, and this will only be feasible in case of clear and geographically similar trends with durations longer than ^{14}C age resolution. Such problems are intrinsic for lake records lacking annual resolution, but should be endeavoured nonetheless because of the strong and possibly unique signal in such small systems.

Conclusions

Shallow and small closed-basin lakes between the West Greenland Ice Sheet margin and Sukkertoppen Iskappe have generally simple basin morphometries and the accumulation and distribution of organic rich sediments indicate a rather regular lake infill since deglaciation. Accumulated sediments overlying glacial and glacio-lacustrine sediments typically consist of homogeneous highly aqueous, olive green and silt-rich gyttja which lack bedding of any kind. Diatom valves make up as much as 20% of dry bulk weight for core NAUJG1.1, whereas they are much less abundant in the sediments of the lakes close to the Inland Ice margin. Particulate organic matter is largely absent in cores collected from the central parts of the lakes and consists largely of limnic macrofossil material, which is consistent with the characteristics of modern lake-wide sediment distributions. Carbon-nitrogen ratios of surficial sediments typically range between 7 and 11 suggesting that organic matter is almost entirely derived from within-lake sources. Dry bulk sediments of the five SFL-cores contain 75-90% minerogenic matter and consist largely of aeolian sediments (Eisner *et al.*, 1995), whereas such sediments are absent from lakes 90 km west of the present day Greenland Ice Sheet margin.

According to new AMS ^{14}C ages, the westernmost lake studied was deglaciated probably around 10,500 cal yrs B.P., whereas the easternmost lake closest to the Inland Ice margin records a Holocene deglaciation around 7250 cal yrs B.P. The start of organic infill in lakes around the head of Kangerlussuaq follow the general timing for deglaciation as defined by regional moraine chronostratigraphy, but includes evidence for diachronous development of basins. This is probably a result of asynchronous melting of ice-rich sediments in peat-filled depressions and the subsequent subsidence and formation of shallow thaw lakes. Basal ages for the two lakes along Kangerlussuaq suggest minimal ages for deglaciation which differ significantly from those reported in Ten Brink & Weidick (1974). The westernmost lake at the innermost margin of the Sarfartoq/Avadtleq moraine system records deglaciation to have taken place some 1300 yrs earlier than suggested by radiocarbon ages on marine shells from a local marine limit of 125-130 m (Ten Brink, 1975), whereas the lowermost organic sediments in a lake above the Angujaartorfik embayment yielded an age somewhat younger than the reported age of 7900-9100 cal yrs B.P. for the Fjord moraine system. Given the intrinsic uncertainties when defining former ice margin positions by assuming a marine-glacial interrelation of ^{14}C-dated elevated (glacio)marine deposits (Ten Brink, 1975; Van Tatenhove *et al.*, 1996), basal ^{14}C ages for organic infill of lakes in ice-free areas might provide a more direct time control for glacial retreat from individual lake-catchment areas (Steig *et al.*, 1998).

According to diatom analysis, the small and shallow lakes have remained moderately nutrient rich (mesotrophic) since deglaciation, which seems to be typical for this area (Eisner *et al.*, 1995). Floristic shifts in fossil diatom assemblages in lake SFL4 indicate an early raw-water phase, followed by a pioneering phase with abundant small alkaliphilous taxa and some small planctonic species. Floristic shifts indicate variable salinity throughout the history of the lake pointing to periods of increased/decreased evaporation rates. Conditions unfavourable for the development of lake biota is most clearly demonstrated for the period after ca. 1000 cal yrs B.P. by the sudden disappearance of major hydrophyte pollen from the lake sediments and concurrent decreases in biogenic silica content.

Statistical analysis of the influence of specific sediment components on sediment variability, and corroborative evidence from paleoecological studies, favour the proposition that lacustrine processes drive the observed lithological variation in the lake sediments. According to core stratigraphies and sediment concentration data, input in the lakes of aeolian sediments seems to have been more constant over time than changes in biogenic input, which contrasts earlier postulations on the type of environmental signal present in the lake cores (Eisner *et al.*, 1995). Importantly, similarities between detailed lithological records among widely separated lakes indicate a regional, possibly climatic controlled forcing factor for in-lake processes. Localized and basin-specific controls seem to play a more dominant role for sediment variability over longer (>500 yrs) timescales, and their apparent overall sensitivity to record regional changes over decade-to-century timescales suggests considerable scope for further paleoclimatic studies in this region.

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References

Anderson, N. J., & Bennike, O. 1997. Holocene lake sediments in West Greenland and their palaeoclimatic and palaeoecological implications. – *Geology of Greenland Survey Bulletin* 176: 89-94.

Anderson, N.J., Bennike, O., Christoffersen, K., Jeppesen, E., Markager, S., Miller, G., & Renberg, I. 1999. Limnological and palaeolimnological studies of lakes in southwestern Greenland. – *Geology of Greenland Survey Bulletin* 183: 68-74.

Anderson, N.J., Clarke, A., Juhler, R.K., McGowan, S. & Renberg, I. 2000. Coring of laminated lake sediments for pigment and mineral magnetic analyses, Søndre Strømfjord, southern West Greenland. – *Geology of Greenland Survey Bulletin* 186: 83-87.

Anderson, N.J., Harriman, R., Ryves, D.B. & Patrick, S.T. 2002. Dominant factors controlling variability in the ionic composition of West Greenland Lakes. –*Arctic, Antarctic and Alpine Research* 33 (4): 418-425.

Barlow, L. K., Rogers, J. C., Serreze, M. C. & Barry, R. G. 1997. Aspects of climate variability in the North Atlantic sector: Discussion and relation to the Greenland Ice Sheet Project 2 high-resolution isotopic signal. – *Journal of Geophysical Research* 102, no.C12: 26,333-26,344.

Barry, R. G. & Kiladis, G. N. 1982. Climatic characteristics of Greenland. – In: Radock, U. *et al*.,(eds.). *Climatic and Physical characteristics of the Greenland Ice Sheet*, Cooperative Institute for Research in Environmental Science, University of Colorado, Boulder: 7-34.

Bennike, O. 2000. Palaeoecological studies of Holocene lake sediments from West Greenland. – *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 285-304.

Bennike, O. & Weidick, A. 1999. Observations on the Quaternary geology around Nioghalvfjerdsfjorden, eastern North Greenland. – *Geology of Greenland Survey Bulletin* 183: 57-60.

Berner, R. A. 1980. *Early diagenesis, A theoretical approach*. – Princeton University Press, Oxford, UK: 241 pp.

Billings, W.D. & Peterson, K.M. 1980. Vegetational change and ice-wedge polygons through the thaw-lake cycle in Arctic Alaska. – *Arctic and Alpine Research* 12: 413-432.

Böcher, T. W. 1949. Climate, soil and lakes in continental West Greenland in relation to plant life. – *Meddelelser om Grønland* 147 (2): 63 pp.

Bradley, R. S., & Miller, G. H. 1972. Recent climatic change and increased glacierization in the eastern Canadian Arctic. – *Nature* 237: 385-387.

Bradley, R. S., Retelle, M. J., Ludlam, S. D., Hardy, D. R., Zolitschka, B., Lamoureux, S. F. & Douglas, M. S. V. 1996. The Taconite Inlet Lakes Project: a systems approach to paleoclimatic reconstruction. – *Journal of Paleolimnology* 16: 97-110.

Britt-Florin, M. & Wright, H. E. 1969. Diatom evidence for the persistence of stagnant glacial ice in Minnesota. – *Geolological Society of America Bulletin* 80: 695-704.

Brodersen, K.P. & Anderson, N.J. 2000. Subfossil insect remains (Chironomidae) and lake-water temperature inference in the Sisimiut-Kangerlussuaq region, southern West Greenland. – *Geology of Greenland Survey Bulletin* 186: 78-82.

Brodersen, K.P. & Anderson, N.J. *In Press.* Distribution of Chironomids (Diptera) in low arctic West Greenland lakes: trophic conditions, temperature and environmental reconstruction. – *Freshwater Biology*.

Cole, G.A. 1994. *Textbook of limnology*. – Waveland Press Inc., Prospect Heights Ill.: 412 pp.

Cumming, B. F. & Smol, J. P. 1993. Development of Diatom-based salinity models for paleoclimatic research from lakes in British Columbia (Canada). – *Hydrobiologia* 269/270: 179-196.

Dean, W. E. Jr. 1974. Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods. – *Journal of Sedimentary Petrology* 44: 242-248.

De Wolf, H. 1982. Method of coding ecological data from diatoms for computer utilisation. – *Mededelingen van de Rijksgeologische Dienst* 36: 95-98.

Denys, L. 1992. A check-list of the diatoms in the Holocene deposits of the western belgian coastal plain with a survey of their apparent ecological requirements. Part I: Introduction, ecological code and complete list. – *Professioneel Paper van de Belgische Geologische Dienst* 1991/2 No. 246: 40 pp.

Dickinson, R. E., Meleshko, V., Randall, D., Sarachik, E., Silva-Dias, P. & Slingo, A. 1996. Climate Processes. – In: Houghton, J.T. *et al.*,(eds.).*Climate Change 1995. The Science of climate change*, Cambridge University Press: 572 pp.

Dixit, S. S., Dixit, A. S. & Smol, J. P. 1991. Multivariable environmental inferences based on diatom assemblages from Sudbury (Canada) lakes. – *Freshwater Biology* 26: 251-266.

Dijkmans, J. W. A. 1988. Seasonal frost mounds in an eolian sand sheet near Søndre Strømfjord, W.Greenland. – In: Senneset, K. (ed.). *Proceedings 1, Permafrost Vth Intern. Conference,* Tapir Publ., Trondheim: 728-733.

– 1989. Frost wedges in an eolian sand sheet near Søndre Strømfjord, W. Greenland and their palaeoenvironmental implications. – *Zeitschrift für Geomorphologie* 33: 339-353.

Dijkmans, J. W. A. & Törnqvist, T. E. 1991. Modern periglacial eolian deposits and landforms in the Søndre Strømfjord area, West Greenland and their palaeoenvironmental implications. – *Meddelelser om Grønland, GeoScience* 25: 39 pp.

Doran, P. T. 1993. Sedimentology of Colour Lake, a nonglacial high arctic lake, Axel Heiberg Island, N.W.T., Canada. – *Arctic and Alpine Research* 25: 353-367.

Doran, P. T., McKay, C. P., Adams, W. P., English, M. C., Wharton Jr., R. A. & Meyer, P. 1996. Climate-forcing and thermal feedback of high arctic residual lake ice covers. – *Limnology and Oceanography* 41: 839-848.

Douglas, M. S. V., Smol, J. P. & Blake Jr., W. 1994. Marked post-18th century environmental change in high-arctic ecosystems. – *Science* 266: 416-419.

Eisner, W. R., Törnqvist, T. E., Koster, E. A., Bennike, O. & Van Leeuwen, J. F. N. 1995. Paleoecological studies of a Holocene lacustrine record from the Kangerlussuaq (Søndre Strømfjord) region of West Greenland. – *Quaternary Research* 43: 55-66.

Engstrom, D. & Wright, H. E. 1984. Chemical stratigraphy of lake sediments as a record of environmental change. – In: Haworth, E. Y. & J. W. G. Linch (eds.). *Lake sediments and environmental history: studies in paleolimnology*. University of Minnesota Press, Minneapolis: 11-68.

Engstrom, D.R., Fritz, S.C., Almendinger, J.E. & Juggins, S. 2000. Chemical and biological trends during lake evolution in recently deglaciated terrain. – *Nature* 408: 161-166.

Escher, A., Sorensen, K. & Zeck, H. P. 1976. Nagssugtoqidian mobile belt in West Greenland. – In: Escher, A and Watt, W. S. (eds.). *Geology of Greenland*. Grønlands Geologiske Undersøgelse, København: 76-95.

Fee, E. J., Hecky, R. E. & Welch, H. E. 1987. Phytoplankton photosynthesis parameters in central Canadian lakes. – *Journal of Plankton Research* 9: 305-316.

Foged, N. 1958. The Diatoms in the basalt area and adjoining areas of Archean rock in West Greenland. – *Meddelelser om Grønland* 156 (4).

Foged, N. 1972. The diatoms in four postglacial deposits in Greenland. – *Meddelelser om Grønland* 194(4): 66 pp.

Fredskild, B. 1977. The development of the Greenland lakes since the last glaciation. – *Folia Limnologia Scandinavia* 17: 101-106.

Fredskild, B. 1984. Holocene palaeo-winds and climatic changes in West Greenland as indicated by long-distance transported and local pollen in lake sediments. – In: Mörner, N. A., and W. Karlen (eds.). *Climatic changes on a yearly to millennial basis*. Reidel, Dordrecht: 163-171.

Fredskild, B. 1985. Holocene pollen records from West Greenland. – In: Andrews, J. T. (ed.). *Quaternary environments: Eastern Canadian Arctic, Baffin Bay, and Western Greenland*. Allen & Unwin, Boston: 643-681.

Fritz, S. C., Cumming, B. F., Gasse, F. & Laird, K. 1999. Diatoms as indicators of hydrologic and climatic change in saline lakes. – In: Stoermer, F. & Smol, J. P. (eds.). *The Diatoms: application for the environmental* *and earth sciences*, Cambridge University Press, Cambridge, UK: 41-72.

Funder, S. 1989. Quaternary geology of the ice-free areas and adjacent shelves of Greenland. – In: Fulton, R. J. (ed.). *Quaternary geology of Canada and Greenland*. Geological Society of Canada, Geology of Canada 1: 741-792.

Håkanson, L. & Jansson, M. 1983. *Principles of lake sedimentology*. – Springer-Verlag, Berlin: 274 pp.

Hansen, K. 1967. The general limnology of arctic lakes as illustrated by examples from Greenland. *Meddelelser om Grønland* 178: 1-77.

Hansen, K. 1970. Geological and geographical investigations in Kong Frederik IX's Land. Morphology, sediments, periglacial processes and salt lakes. *Meddelelser om Grønland* 188: 77 pp.

Hardy, D. R., Bradley, R. S. & Zolitschka, B. 1996. The climatic signal in varved sediments from lake C2, northern Ellesmere Island, Canada. – *Journal of Paleolimnology* 16: 227-238.

Harrison, S. & Digerfeldt, G. 1993. European lakes as palaeohydrological and palaeoclimatic indicators. – *Quaternary Science Review* 12: 233-248.

Hasholt, B. & Søgaard, H. 1978. Et forsøg på en klimatisk-hydrologisk regioninddeling af Holsteinborg kommune (Sisimiut). – *Geografisk Tidsskrift* 77: 72-92.

Hedegaard, K. 1982. *Wind vector and extreme wind statistics in Greenland*. – Danish Meteorological Institute, Copenhagen: 106 pp.

Hendey, N. J. 1964. *Fishery Investigations Series IV. An introductory account of the smaller algae of British Coastal Waters. Part V: Bacillariophycae (Diatoms)*. – H.M.S.O., London.

Hobbie, J. 1973. Arctic Limnology: A review. – In: Britton, M. (ed.). *Alaskan Arctic Tundra, Calgary*. Arctic Institute of North America Technical Paper 25: 127-168.

Holt, S. 1980. Vegetation patterns and effects of grazing on Caribou ranges in the Søndre Strømfjord area, West Greenland. – In: Reimers, E., Gaare, E., & Skjenneberg, S. (eds.). *Proceedings of the 2nd International Reindeer/Caribou Symposium, Røros, Norway* 1979.

Hughen, K. A., Overpeck, J. T., Anderson, R. F. & Williams, K. M. 1996. The potential for paleoclimate records from varved Arctic lake sediments: Baffin Island, Eastern canadian Arctic. – In: Kemp, A.(ed.). *Palaeoclimatology and palaeoceanography from laminated sediments*. Geological Society of London Special Publications 116: 57-71.

Jacobs, J. D., Andrews, J. T. & Funder, S. 1985. Environmental background. – In: Andrews, J.T. (ed.). *Quaternary environments. Eastern Canadian Arctic, Baffin Bay and Western Greenland*. Allen and Unwin, Boston. Ch.2: 27-68.

Johnsen, S. J., Dahl-Jensen, D., Dansgaard, W. & Gundestrup, N. S. 1995. Greenland temperatures derived from GRIP borehole temperature and ice core isotope profiles. – *Tellus Series* B 47: 624-629.

Kapsner, W. R., Alley, R. B., Shuman, C. A., Anandakrishan, S., and Grootes, P. M. 1995. Dominant influence of atmospheric circulation on snow accumulation in Greenland over the past 18,000 years. – *Nature* 373: 52-54.

Kattenberg, A., Giorgi, F., Grassl, H., Meehl, G. A., Mitchell, J. F. B., Souffer, R. J., Tokioka, T., Weaver, A. J. & Wigley, T. M. L. 1996. Climate models – Projections of future climate. – In: Houghton, J. T. et al (eds.). *Climate change 1995. The Science of climate change*. Cambridge University Press: 572 pp.

Keen, R. A. 1980. *Temperature and circulation anomalies in the eastern Canadian Arctic summer 1946-76*. – Institute for Arctic and Alpine Research Occasional Paper 34, University of Colorado, Boulder: 159 pp.

Kelts, K.& Hsü, K. J. 1978. Freshwater carbonate sedimentation. – In: Lerman, A. (ed.). *Lakes: Chemistry, Geology, Physics*. Springer-Verlag, Berlin: 295-323.

Krammer, K. & Lange-Bertalot, H. 1986. Bacillariophycae, teill II, Epithemiacae, Surirellacae. – In: Ettl, H., Gerloff, J., Heynig, H. & Mollenhauer, D. (eds.). *Süsswasserflora von Mitteleuropa*. Band 2/2, Gustav Fischer Verlag, Stuttgart/New York: 876 pp.

Krammer, K. & Lange-Bertalot, H. 1986. Bacillariophycae, Teil I, Naviculacae. – In: Ettl, H., Gerloff, J., Heynig, H. & Mollenhauer, D. (eds.). *Süsswasserflora von Mitteleuropa*. Band 2/1, Gustav Fischer Verlag, Stuttgart/ New York: 876 pp.

Lamoureux, S. F. & Bradley, R. S. 1996. A 3300 year varved sediment record of environmental change from northern Ellesmere Island, Canada. – *Journal of Paleolimnology* 16: 239-255.

Larsen, C. P. S. & MacDonald, G. M. 1993. Lake morphometry, sediment mixing and the selection of sites for fine resolution palaeoecological studies. – *Quaternary Science Review* 12: 781-792.

Magnuson, J. J., Robertson, D. M., Benson, B. J., Wynne, R. H., Livingstone, D. M., Arai, T., Assel, R. A., Barry, R. G., Card, V., Kuusisto, E., Granin, N. G., Prowse, T. D., Stewart, K. M. & Vuglinski, V. S. 2000. Historical Trends in Lake and River Ice Cover in the Northern Hemisphere. – *Science* 289: 1743-1746.

Meyers, P. A. & Lallier-Vergès, E. 1999. Lacustrine sedimentary organic matter records of Late Quaternary paleoclimates. – *Journal of Paleolimnology* 21: 345-372.

Miller, R. 1992. What is preserved in arctic lakes? – In: *Proceedings of the 2nd Arctic Workshop, Boulder*: 104-105.

Myeni, R. B., Keeling, C. D., Tucker, C. J., Asrar, G. & Nemani, R. R. 1997. Increased plant growth in the northern high latitudes from 1981 to 1991. – *Nature* 386, 698-702.

Nicholls, N., Gruza, G.V., Jouzel, J., Karl, T. R., Ogallo, L. A. & Parker, D. E. 1996. Observed climate variability and change. – In: Houghton, J.T. et al (eds.). *Climate change 1995. The Science of climate change*. Cambridge University Press: 133-192.

Nichols, H. 1967. The disturbance of arctic lake sediments by "bottom ice": A hazard for palynology. – *Arctic* 20: 213-225.

Nichols, N., Kelly, P. M. & Andrews, J. T. 1978. Holocene palaeo-wind evidence from palynology in Baffin Island. – *Nature* 273: 140-142

Odgaard, B.V. 1993. Wind-determined sediment distribution and Holocene sediment yield in a small, Danish, kettle lake. – *Journal of Paleolimnology* 8, 3-13.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A. & Zielinski, G. 1997. Arctic environmental change of the last four centuries. – *Science* 278: 1251-1256.

Patrick, R. & Reimer, C. W. 1966. *The diatoms of the United States. Part I: Fragilariacae, Eunotiacae, Achnantacae, Naviculacae*. – Academy for the Natural Sciences, Philidelphia: 213 pp.

Pienitz, R., Smol, J. P. & Birks, H. J. 1995. Assessment of freshwater diatoms as quantitative indicators of past climatic change in the Yukon and Northwest Territories, Canada. – *Journal of Paleolimnology* 13: 21-49.

Putnins, P., 1970. The climate of Greenland. – In: Orvig, S. (ed.).*Climates of the Polar regions*, World Survey of Climatology 14, Elsevier, Amsterdam. 3-128.

Quayle, W. C., Peck, L. S., Peat, H., Ellis-Evans, J. C. and Harrigan, P. R. 2002. Extreme responses to climate change in Antarctic lakes. – *Science* 295: 645.

Reeh, N. 1989. Dynamic and climatic history of the Greenland Ice Sheet. – In: Fulton, R.J. (ed.). *Quaternary geology of Canada and Greenland*. Geological Survey of Canada, Geology of Canada 1: 793-819.

Røen, U. 1962. Studies on freshwater entomostraca in Greenland II. – *Meddelelser om Grønland* 170(2).

Rott, H. & Obleitner, F. 1992. The energy balance of dry tundra in West Greenland. – *Arctic and Alpine Research* 24: 352-362.

Schindler, D. W., Welch, H. E., Kalff, J., Brunskill, G. J. & Kritsch, N. 1974. Physical and chemical limnology of Char Lake, Cornwallis Island (75N Lat.). – *Journal of the Fisheries Research Board Canada* 31: 585-607.

Schindler, D. W., Bayley, S. E., Parker, B. R., Beaty, K. G., Cruikshank, D. R. 1996. The effects of climate warming on the properties of boreal lakes and streams at the Experimental Lakes Area, northwestern Ontario. – *Limnology and Oceanography* 41: 1004-1017.

Smol, J. P. 1988. Paleoclimatic proxy data from freshwater Arctic diatoms. – *Verhandelungen der Internationalen Verein für Limnology* 23: 37-44.

Smol, J. P., Cumming, B. F., Douglas, M. S .V. & Pienitz, R. 1996. Inferring past climatic changes in Canada using paleolimnological techniques. – *GeoSciences Canada* 21: 113-118.

Smol, J. P., Walker, I. R. & Leavitt, P. R. 1991. Paleolimnology and hindcasting climatic trends. – *Verhandelungen der Internationalen Verein für Limnologie* 24: 1240-1246.

Snowball, I. & Sandgren, P. 1996. Lake sediment studies of Holocene glacial activity in the Kårsa valley, northern Sweden: contrast in interpretation. – *The Holocene* 6: 367-372.

Steig, E.J., Wolfe, A.P. & Miller, G.H. 1998. Wisconsinan refugia and the glacial history of eastern Baffin Island, Arctic Canada: coupled evidence from cosmogenic isotopes and lake sediments. – *Geology* 26: 835-838.

Ten Brink, N. W. 1975. Holocene history of the Greenland Ice Sheet based on radiocarbon dated moraines in West Greenland. – *Meddelelser om Grønland* 201(4): 44 pp.

Ten Brink, N. W. & Weidick, A. 1974. Greenland Ice Sheet history since the Last Glaciation. – *Quaternary Research* 4: 429-440.

Törnqvist, T. E. & Bierkens, M. F. P. 1994. How smooth should curves be for calibration of radiocarbon ages? – *Radiocarbon* 36: 11-26.

Tramoni, F., Barry, R. G. & Key, J. 1985. Lake ice cover as a temperature index for monitoring climate perturbations. *Zeitschrift für Gletscherkunde und Glazialgeologie* 21: 43-49.

Van den Broeke, M. R., Duynkerke, P. G. & Oerlemans, J. 1994. The observed katabatic flow at the edge of the Greenland Ice Sheet during GIMEX-91. – *Global and Planetary Change* 9: 3-15.

Van der Plicht, J. 1993. Groningen radiocarbon calibration program. – Radiocarbon 35: 231-237.

Van Tatenhove, F. G. M and Olesen, O. B. 1994. Groundtemperature and related permafrost characteristics in West Greenland. – *Permafrost and Periglacial Processes* 5: 199-215.

Van Tatenhove, F. G. M., Van der Meer, J. J. M. & Koster E. A. 1996. Implications for deglaciation chronology from new AMS age determinations in Central West Greenland. – *Quaternary Research* 45: 245-253.

Van Tatenhove, F.G.M. 1996. Changes in morphology at the margin of the Greenland Ice Sheet (Leverett glacier), in the period 1943-1992: A quantitative analysis. – *Earth Surface Landforms and Processes* 21: 797-816.

Vincent, W.F. & Pientitz, R. 1996. Sensitivity of high-latitude freshwater ecosystems to global change: temperature and solar ultraviolet radiation. – *Geoscience Canada* 23: 231-236.

Warrick, R. & Oerlemans, J. 1990. Sea level rise. – In: Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (eds.). *Climate change, the IPCC scientific assessment*, Press Syndicate Cambrige University: 257-282.

Weidick, A. 1993. Neoglacial change of ice cover and related response of the Earth's crust in West Greenland. – *Grønlands Geologiske Undersøgelse Rapport* 159: 121-126.

Weidick, A., Oerter, H., Reeh, N., Thomsen, H. H. & Thoring, L. 1990. The recession of the Inland Ice margin during the Holocene climatic optimum in the Jacobshavn Isfjord area of West Greenland. – *Palaeogeography, Palaeoclimatology, Palaeoecology* 82: 389-399.

Weidick, A. & Ten Brink, N. 1970. Quaternary deposits between the Sukkertoppen ice cap and Nordre Strømfjord. – *Grønlands Geologiske Undersøgelse Rapport* 28: 23-25.

Welch, H. E., Legault, J. A. & Bergmann, M. A. 1987. Effects of snow and ice on the annual cycles of heat and light in Saqvaqjuac lakes. – *Canadian Journal of Fisheries and Aquatic Sciences* 44: 1451-4161.

White, J. W. C. Barlow, L. K., Fisher, D., Grootes, P., Jouzel, J., Johnsen, S.J., Stuiver, M. & Clausen, H. 1997. The climate signal in the stable isotopes of snow from Summit, Greenland: Results of comparison with modern climate observations. – *Journal of Geophysical Research* 102 C12: 26,425-26,439.

Willemse, N. W. & Törnqvist, T. E. 1999. Holocene century-scale temperature variability from West Greenland lake records. – *Geology* 27: 580-584.

Williams, W. D. 1991. Comments on the so-called salt lakes of Greenland. – *Hydrobiologia* 210: 67-74.

Williams, L. D. & Bradley, R. S., 1985, Paleoclimatology of the Baffin Bay Region. – In: Andrews, J.T. (ed.). *Quaternary environments. Eastern Canadian Arctic, Baffin Bay and Western Greenland*. Allen and Unwin, Boston. Ch.26: 741-772.

Wohlfart, B., Skog, G., Possnert, G. & Holmquist, B. 1998. Pitfalls in the AMS radiocarbon-dating of terrestrial macrofossils. – *Journal of Quaternary Science* 13: 137-145.

Zolitschka, B. 1996. Recent sedimentation in a high arctic lake, northern Ellsmere Island, Canada. – *Journal of Paleolimnology* 16: 169-186.

MONOGRAPHS ON GREENLAND

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Feedback mechanisms of global warming in the Arctic could have a significant impact on climate dynamics outside the polar regions. Our current understanding of the arctic influences on lower latitude climate dynamics is, however, limited.

Small freshwater lakes in the Arctic form ideal natural archives recording the climate variability, because of their continuous sediment records, wide availability, general climatic sensitivity, and the lack of human influence. This monograph presents detailed information on the environmental background, paleoenvironmental record, and climatic significance of the shallow freshwater lakes in the interior continental area of West Greenland which will help to link paleoclimatic studies from Baffin Island with those from the Greenland ice cap.

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