Quaternary Research **58**, 149–159 (2002)
doi:10.1006/qres.2002.2352

Holocene Environmental Variability in Southern Greenland Inferred from Lake Sediments

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Received October 18, 2001

Sediments from Qipisarqo Lake provide a continuous Holocene paleoenvironmental record from southern Greenland. Following deglaciation and glacio-isostatic emergence of the basin from the sea ∼9100 cal yr B.P., proxies of lake paleoproductivity, including biogenic silica and organic matter, increased markedly until 6000 cal yr B.P. and thereafter remained stable over the ensuing warm three millennia. Subsequent decreases in these proxies, most dramatically between 3000 and 2000 cal yr B.P., show the lake's responses to initial Neoglacial cooling. Intervals of ameliorated limnological conditions occurred between 1300 and 900 and between 500 and 280 cal yr B.P., briefly interrupting the decreasing trend in productivity that culminated in the Little Ice Age. Increased lake productivity during the latter half of the 20th century, which reflects the limnological response to circum-arctic warming, still has not reached peak Holocene values. The Neoglacial climate of the last 2000 yr includes the most rapid high-amplitude environmental changes of the past nine millennia. The Norse thus migrated around the North Atlantic Ocean region in the most environmentally unstable period since deglaciation. Lacustrine sediment records provide a context with which to consider future environmental changes in the Labrador Sea region. In particular, any future warming will be superposed on a regional climate system that is currently exhibiting highly unstable behavior at submillennial timescales.

INTRODUCTION

Information on past climatic and ecological variability provides a context in which to consider present and future environmental changes, thus improving our understanding of spatial patterns and underlying mechanisms of natural climate variability. The Arctic is particularly important in this regard, given its susceptibility to anthropogenic warming due to the sensitivity of albedo to relatively small changes in snow and ice cover (Overpeck *et al.*, 1997; Rouse *et al.*, 1997; MacDonald *et al.*, 2000). Although a number of paleoenvironmental studies have been conducted in northwestern (Blake *et al.*, 1992), western (Kelly, 1985; Willemse and Törnqvist, 1999; Bennike, 2000), and eastern (Cremer *et al.*, 2001) coastal Greenland, relatively little is known from the southern part of the island, despite its strategic location in relation to adjacent North Atlantic oceanographic currents and dynamics.

Lake sediments offer continuous and datable paleoenvironmental archives that may frequently be interpreted in climatic terms. This study reports paleolimnological inferences regarding Holocene climatic variability from a small lake in southern Greenland. Our reconstruction of the area's environmental history is based on the lake sediments' physical–chemical properties, including magnetic susceptibility, density, water content, and biogenic silica and organic matter concentration, which have been widely demonstrated to be a function of changes in watershed and lake characteristics (e.g., Noon *et al.*, 2001). In turn, watershed and lake physical–chemical properties have been strongly linked to climatic and glacial influences in high-latitude regions, especially where ice cover exists for much of the year (e.g., Rouse *et al.*, 1997; MacDonald *et al.*, 2000; Joynt and Wolfe, 2001; Noon *et al.*, 2001).

Our findings illustrate significant paleoenvironmental variability throughout the Holocene, but especially during the last 3000 yr, which is in contrast to the warmer and more stable conditions of the middle Holocene. The most rapid shifts occur stratigraphically back-to-back in the most recent sediment

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0033-5894/02 $35.00

record, including the Little Ice Age (LIA), which marks the single largest Holocene glacial advance, and subsequent warming that probably includes, in the second half of the 20th century, some degree of amplification from anthropogenic greenhouse gas emissions.

STUDY SITE

Qipisarqo Lake (61°00′41″N, 47°45′13″W) is situated on a coastal lowland (7 m above sea level [asl]) less than 1 km from the shores of the inner reach of Kobberminebugt, a large fiord system that connects directly to the open ocean 30 km to the west. The terminus of a large outlet of the Greenland Ice Sheet, Nordre Qipisarqo Bræ, is less than 2 km from the lake (Fig. 1). However, the lake is not presently influenced by glacial meltwater due to a morainic hill that diverts proglacial drainage to the west and south of the lake's catchment. Thus, any advance greater than 1 km beyond the present terminus would impinge into the catchment of Qipisarqo Lake, supplying glacial sediment to an otherwise nonglacial closed basin system. Limnologically, the lake's waters are highly dilute and ultra-oligotrophic, although proximity to the ocean is clearly reflected by dominance of the sea-salt ions Cl^-, SO_4^{2-}, Na^+ and Ca^{2+} (Table 1).

The climate of southern Greenland is influenced by both cold air masses associated with proximity to the ice sheet, and warmer maritime air related to the regional oceanographic circulation pattern (Funder, 1989). The southwestern coast receives the highest precipitation of all of Greenland due to the presence of the warm West Greenland Current (Fig. 1) which integrates waters of the cold East Greenland Current and warm Atlantic waters derived from the Irminger Current (Putnins, 1970). At Ivigtut, 30 km northwest of Qipisarqo Lake, mean annual temperature is 1.8°C with an annual range of 15.3°C, whereas annual precipitation is 130 cm. Vegetation is moderately lush shrub tundra dominated by shrub birch (*Betula*) and heaths (*Cassiope, Empetrum, Vaccinium*). Although shrub alder (*Alnus*) is regionally common at the heads of fiords, it is not present on the Qipisarqo foreland, presumably due to the glacier's proximity. Local bedrock includes several Proterozoic crystalline lithologies, mainly gneisses and migmatites with various plutonic bodies. Although the exact position of the Greenland Ice Sheet during the Late Wisconsinan is unknown, it most likely extended well onto the continental shelf (Funder, 1989). Retreat to the outer coast was probably complete by about 11,500 ^{14}C yr B.P. (Bennike, 2000).

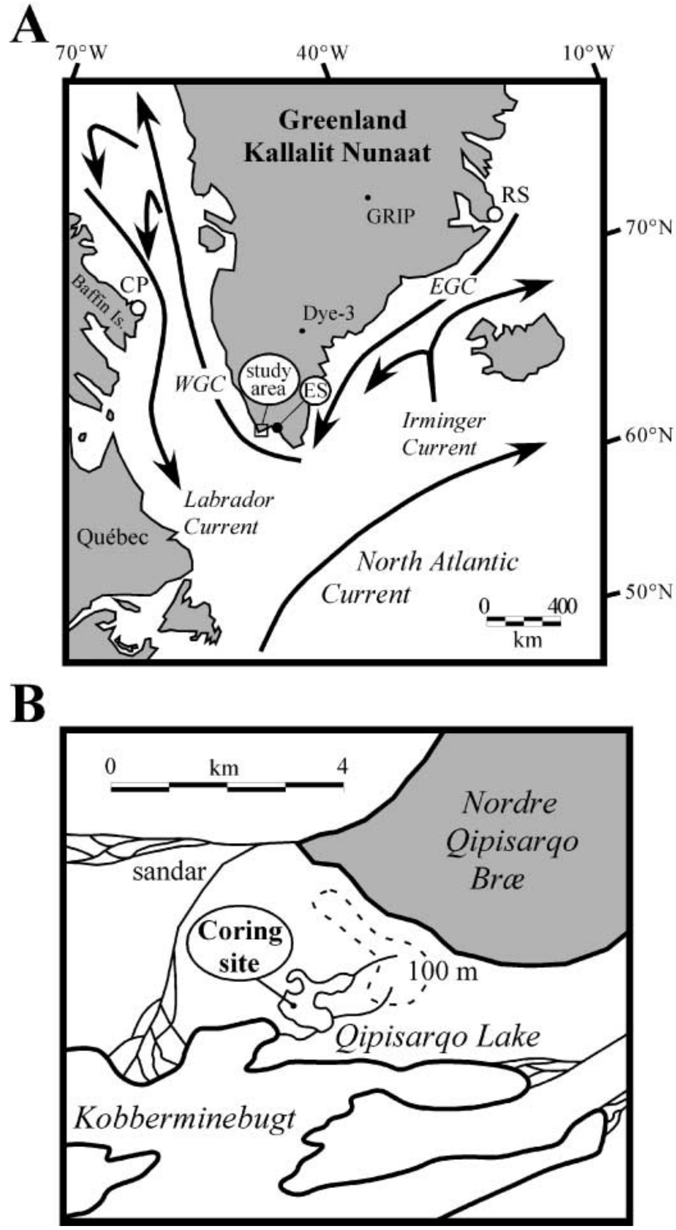

FIG. 1. A: Location of the study area in relationship to major North Atlantic Ocean currents including WGC (West Greenland Current) and EGC (East Greenland Current). The locations of other sites mentioned in the text are abbreviated as follows: CP (Cumberland Peninsula, Baffin Island), RS (Raffles Sø, Jameson Land), and ES (Norse eastern settlement). B: Local situation of Qipisarqo Lake between inner Kobberminebugt and Nordre Qisarqo Bræ. The 100-m contour delimits the hill that presently impedes proglacial drainage from entering the lake.

TABLE 1
Qipisarqo Lake Water Chemistry

Parameter	Value
pH	6.62
Conductivity	$27\ \mu S \cdot cm^{-1}$
Cl^-	$3.23\ mg \cdot L^{-1}$
Na^+	$1.79\ mg \cdot L^{-1}$
SO_4^{2-}	$1.24\ mg \cdot L^{-1}$
Ca^{2+}	$1.10\ mg \cdot L^{-1}$
Mg^{2+}	$0.48\ mg \cdot L^{-1}$
SiO_2	$0.40\ mg \cdot L^{-1}$
K^+	$0.28\ mg \cdot L^{-1}$
PO_4^{3-}	<0.002 (d.l.) $mg \cdot L^{-1}$
NH_4^+	<0.001 (d.l.) $mg \cdot L^{-1}$
NO_3^-	<0.001 (d.l.) $mg \cdot L^{-1}$

d.l.: detection limit.

METHODS

The complete sediment sequence from Qipisarqo Lake was obtained by overlapping a percussion core (Nesje, 1992) for the lower 267 cm with a 17-cm gravity core that preserved the mud–water interface intact (Glew, 1989). Both cores were retrieved through the ice in May 1998 from the lake's western basin, in a water depth of 9 m. The gravity core was extruded in continuous 0.25-cm increments in the field, whereas the percussion core was split lengthwise, described, and sampled in the laboratory. Volumetric subsamples (0.65 cm^3) were taken in the percussion core every 2 cm, except around lithologic boundaries where sampling was every 1 cm. In the gravity core, sampling intervals increase from 0.25 to 0.5 cm between core top and 3.75 cm depth, and from 1 to 2 cm between 3.75 and 10.25 cm depth. All samples were dried at 35°C overnight (i.e., for at least 16 h) and weighed to obtain sediment water content and density. Consistency between neighboring samples and repeated measurements on samples throughout the core, specifically where water content is highest, showed reproducibility indicating sediment water was removed. Magnetic susceptibility (MS) was measured with a Bartington system on both wet and dry samples (there was no significant difference). Dried sediments were heated overnight in ceramic crucibles at 105°C, weighed to obtain % hygroscopic moisture (by mass), and subsequently ignited at 550–600°C to derive mass loss on ignition (LOI), a measure of sediment organic matter content (e.g., Dean, 1974; Heiri *et al.*, 2001).

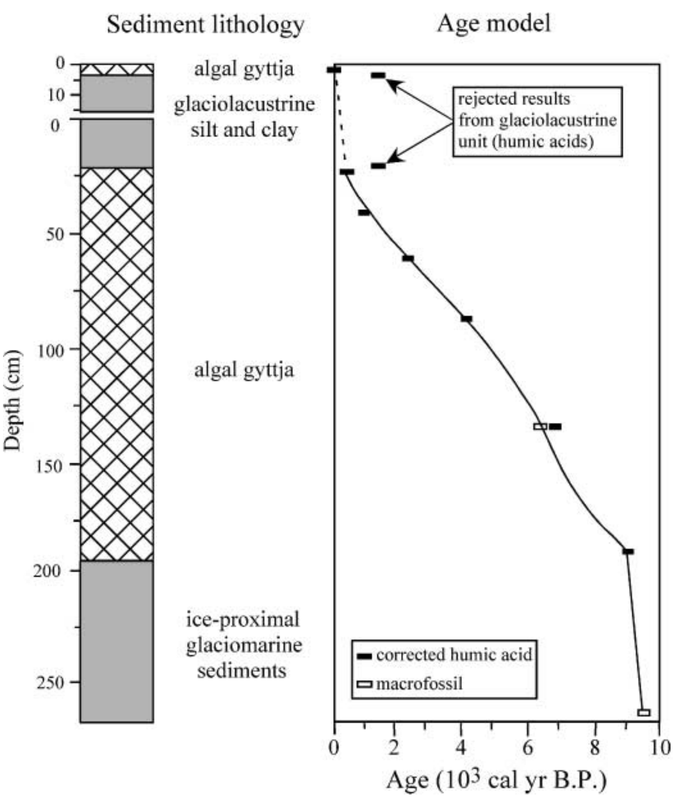

FIG. 2. Sediment lithostratigraphy and ^{14}C ages. The age model for the percussion core is shown on the right (polynomial fit to 194 cm, linear 194–267 cm). We infer the two ^{14}C ages from the glaciolacustrine unit associated with sedimentation during the LIA reflect reworking of old carbon during glacial advance.

In the percussion core, a less densely spaced set of samples, taken every 2 cm in the upper 60 cm of the sequence and 6–12 cm below 60 cm, was prepared for grain size and biogenic silica analyses. In the gravity core, sampling intervals increase from 0.25 to 1 cm between the core top to 3.75 cm depth, and 0.75 to 2 cm between 3.75 and 10.5 cm depth (7.5 cm depth for biogenic silica). Grain size analyses were performed on a Malvern laser diffraction particle size analyzer and are restricted to the lacustrine sediment units. For biogenic silica ($BSiO_2$) a sequential dissolution methodology was used (Mortlock and Froelich, 1989; Carter and Coleman, 1994). On freeze-dried sediment, organic matter and carbonates were first removed with 10% H_2O_2 and 1N HCl for 1 h in a 70–80°C water bath. Then, $BSiO_2$ was dissolved by treating samples with 2N $NaCO_3$ for 5 h, also at 70–80°C, with complete sample agitation every hour. Samples were thereafter centrifuged, and 1.0 ml aliquots of supernatant were removed, to which 0.25 ml 70% HNO_3 and 8.5 ml of deionized H_2O were added. Total Si and Al were measured on these aliquots by inductively coupled plasma-atomic emission spectroscopy (ICP-AES). Reproducibility for duplicate analyses is less than ±4%. To test whether sediment $BSiO_2$ dissolution was complete after 5 h, aliquots of supernatant were taken from a range of samples at 1-h intervals up to 7 h, and analyzed for Si and Al. After 5 h of extraction time, concentrations of Si and Al had stabilized in all samples, and microscopic inspection of the samples revealed no remaining siliceous microfossils. After correcting raw ICP-AES measurements against procedural blanks, total Al was used to correct Si concentrations for contributions from clay mineral dissolution. Total Al was multiplied by 2 (the average molar ratio of Si : Al in most clays; Eggimann *et al.*, 1980) and then subtracted from the total measured Si. We emphasize that correction for the aluminosilicate contribution only slightly alters the value and never changes the sign of a result or conclusions. The first derivative of the raw $BSiO_2$ data set was computed by the central difference approximation technique (Smith, 1985), which estimates the slope between every successive pair of $BSiO_2$ measurements. Although this translates to the between-sample rate of change in %$BSiO_2$ (positive or negative), the calculation itself is independent of the age model assigned to the sediments.

Sediment chronology is based on a series of accelerator mass spectrometry (AMS) ^{14}C measurements, which target the depths of major lithologic boundaries and changes in sediment proxies. Despite an extensive search, macrofossils were only recovered from two depths (136 and 264 cm). All additional dates were performed on sediment humic acid extracts (Abbott and Stafford, 1996). Samples were prepared at the Laboratory for AMS Radiocarbon Preparation and Research (University of Colorado) and measured at either the National Ocean Sciences AMS Facility (Woods Hole Oceanographic Institution) or

Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory.

RESULTS

Sediment Stratigraphy and Chronology

The lower 73 cm of the percussion core comprise massive fine-grained glaciomarine muds that reflect sedimentation into an isostatically depressed lagoonal embayment. The transition to overlying organic lacustrine sediments (gyttja) is very abrupt, implying rapid uplift at the time of isolation from the sea, which is confirmed by the ^{14}C dating results (Fig. 2). The percussion core contains 171 cm of structureless olive-brown gyttja, which is capped by 23 cm of faintly laminated minerogenic, organic-poor sediment. Although the percussion core failed to recover any overlying organic sediments, the gravity core faithfully preserves the return to organic sedimentation in its upper 3.75 cm (Fig. 2). The gravity core gyttja is compositionally very similar to that in the percussion core.

The AMS ^{14}C dating results are reported on Table 2. Paired analyses of humic acid and macrofossils from 136 cm indicate

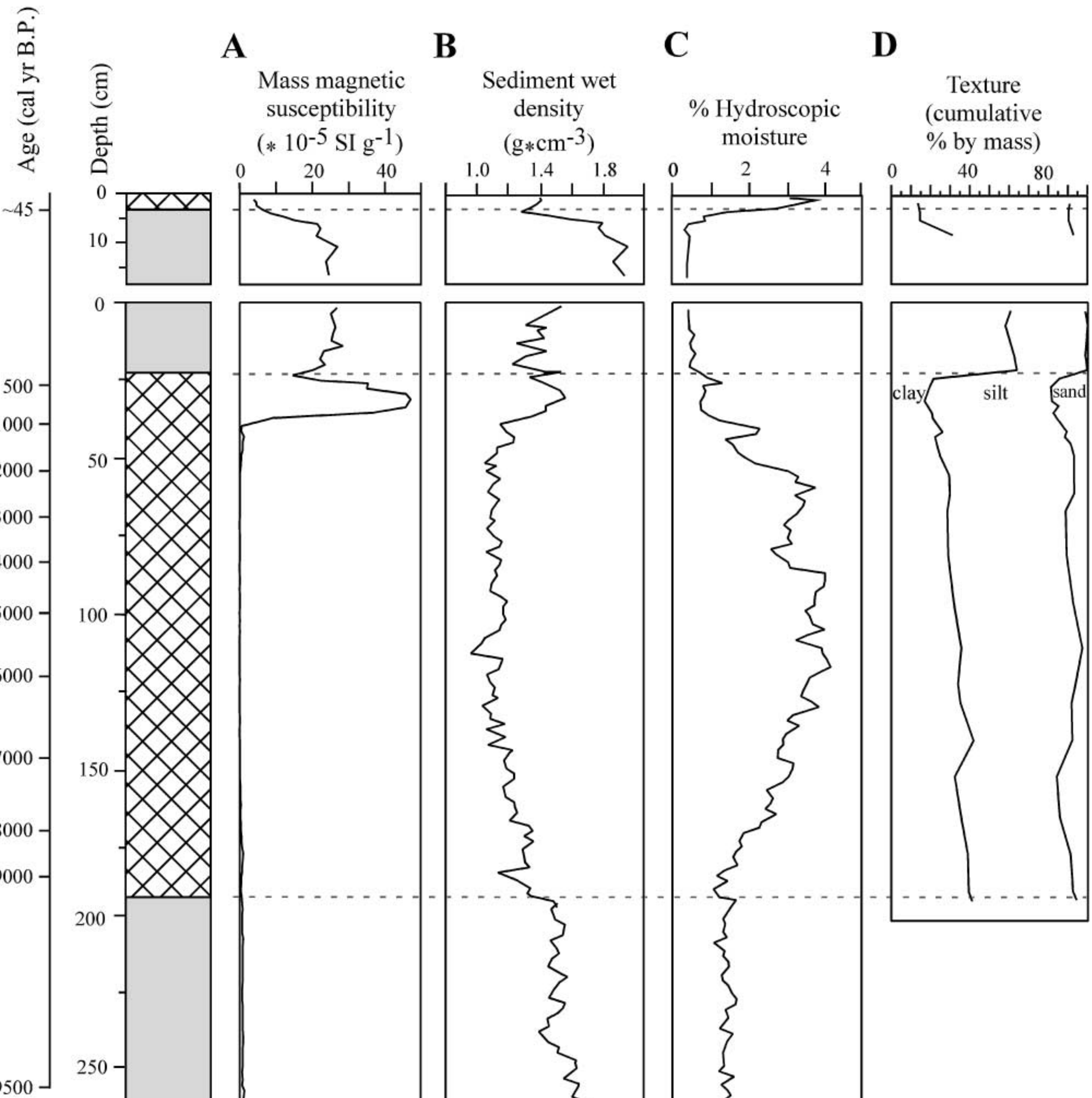

FIG. 3. Physical properties of the composite sediment stratigraphy from Qipisarqo Lake.

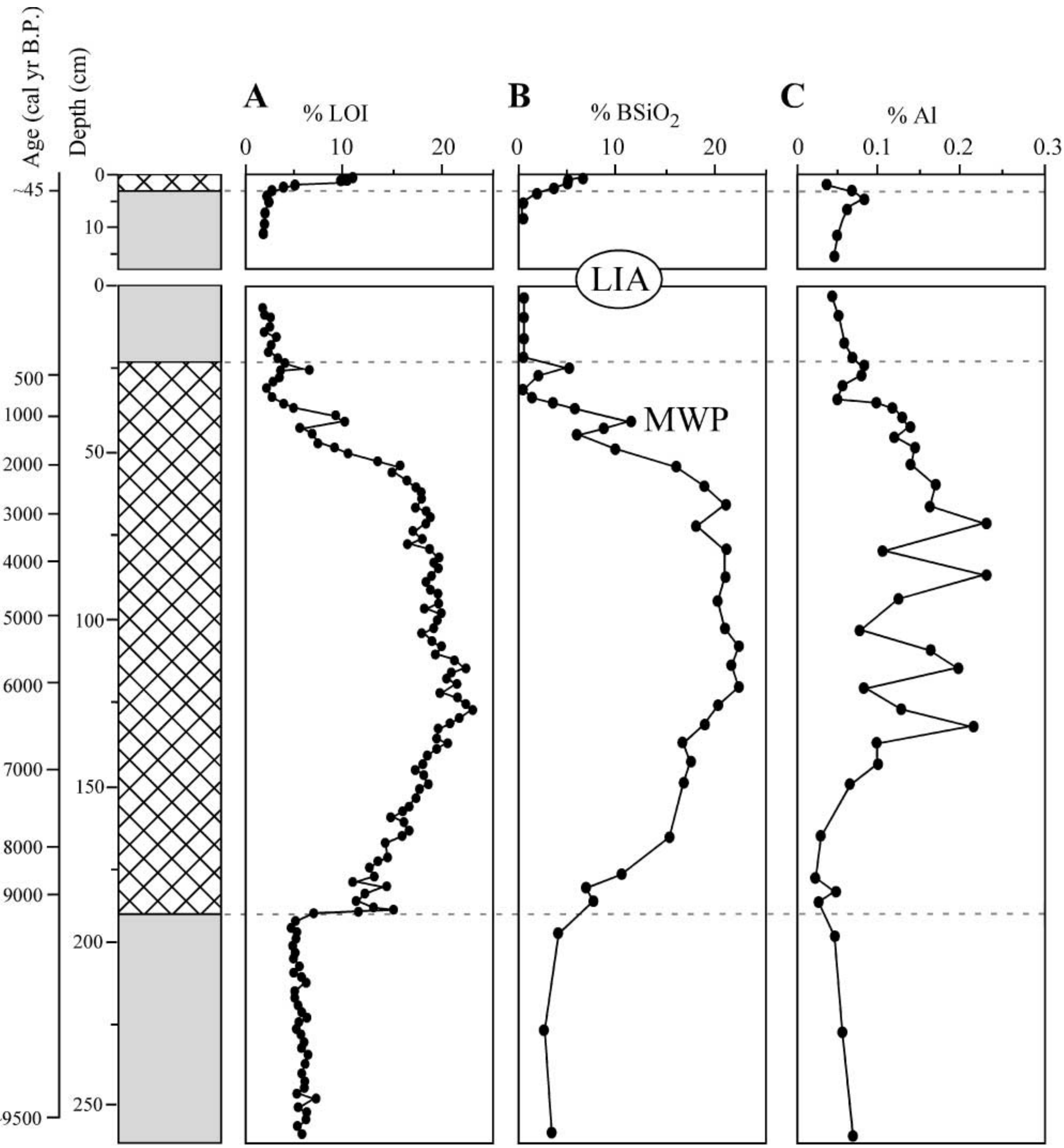

FIG. 4. Profiles of loss on ignition (LOI), $BSiO_2$, and Al from the composite sediment stratigraphy from Qipisarqo Lake. LIA = Little Ice Age; MWP = Medieval Warm Period.

that the former is 230 yr older than the latter, an offset that reflects the lagged sequestration of the terrestrial component of humic acid to the lake's sediments (Abbot and Stafford, 1996). On Baffin Island, this offset is typically in the order of 300 yr (Miller *et al.*, 1999). In contrast, macrofossils from limnologically similar environments to Qipisarqo Lake, including aquatic bryophytes, have been documented as equilibrated with respect to atmospheric CO_2 (Abbott and Stafford, 1996). Thus, the remaining unpaired humic acid ages were corrected by subtracting the 230-yr offset measured at 136 cm, and then calibrated to "calendar" years following Stuiver *et al.* (1998). All ages reported subsequently are in calibrated years (i.e., cal yr B.P.). The resulting model of sediment accumulation is almost linear for the organic portion of the percussion core (Fig. 2), resulting in an average sediment accumulation rate of 21 cm 10^3 yr^{-1}. Dating of the minerogenic organic-poor unit proved problematic, however,

TABLE 2
Radiocarbon Dating Results from the Qipisarqo Lake Cores

Depth (cm)	Radiocarbon age (^{14}C yr B.P.)	δ^{13}C ‰	Lab number[a]	Radiocarbon age corrected (^{14}C yr B.P.)[b]	Calibrated Age (solution range) (cal yr B.P. $\pm 1\sigma$)[c]	Material Dated
Glew core						
2	>Modern[d]	−25.1	CURL-5578	>Modern	>Modern	Humic acid extract
3.75	1300 ± 35	−25.0[e]	CURL-4477	1070 ± 35	966 (934–1047)	Humic acid extract
Piston core						
20.0	1240 ± 35	−25.1	CURL-4475	1010 ± 35	930 (923–952)	Humic acid extract
23.0	530 ± 40	−22.9	CURL-3036	300 ± 40	313 (230–329)	Humic acid extract
40.0	1330 ± 40	−23.2	CURL-4486	1100 ± 40	1040 (957–1056)	Humic acid extract
60.0	2570 ± 35	−22.2	CURL-4476	2340 ± 35	2347 (2339–2353)	Humic acid extract
86.5	3970 ± 40	−20.9	CURL-3037	3740 ± 40	4090 (3991–4149)	Humic acid extract
136.0	5840 ± 60	−20.5	CURL-4487	5610 ± 60	6364 (6308–6446)	Humic acid extract
136.0	5610 ± 60	−25.0[e]	CURL-4488	5610 ± 60	6365 (6308–6446)	Bryophyte and vascular leaf
187.5	8400 ± 40	−18.1	CURL-3038	8170 ± 40	9088 (9027–9249)	Humic acid extract
264.0	8560 ± 50	−25.0[e]	CURL-4914	8560 ± 50	9533 (9502–9547)	Bryophyte

[a] CURL: University of Colorado Radiocarbon Laboratory.
[b] 230 years subtracted from humic acid ages to correct for offset measured between paired samples at 136 cm (see text).
[c] Calibrated according to Stuiver *et al.* (1998).
[d] Fraction modern: 1.208 ± 0.0048.
[e] For three samples δ^{13}C was estimated due to insufficient remaining material for measurement.

with two humic acid results (CURL-4475 and 4477) almost 1000 yr older than expected, one each from the percussion core and gravity core (Table 2). The latter age is on the ~3.75 cm transition to overlying gyttja. The simplest explanation of these inverted ages is that the minerogenic organic-poor unit contains a pool of old dissolved organic carbon. We infer that the upper unit in the percussion core (corresponding to the lower unit in the gravity core) is glaciolacustrine in origin based on the physical properties described below, specifically high silt clay percent and lack of fossil material, deposited due to the nearby proximity of the outlet glacier. The glacial advance would have reworked surrounding soils, causing old dissolved organic carbon to enter the sediment-laden meltwaters and thereafter transfer to the lake. The effects of such contamination would have been maximized by the low rates of autochthonous organic matter production at this time (see below).

Based on the ^{14}C evidence for the presence of reworked dissolved organic carbon (i.e., possible change in erosional regime in the catchment), including at the 3.75-cm lithologic transition in the gravity core (Fig. 2), ^{210}Pb dating of the gravity-core gyttja and underlying glaciolacustrine unit was not pursued. In the absence of reliable ^{14}C ages for the glaciolacustrine unit, an alternate approach was taken to assess the age of the transition to overlying organic sediments. First, from the 45–195 cm interval of the percussion core, which is lithologically almost identical to the recently deposited gyttja, mean dry density (0.340 g cm^{-3}) and sediment accumulation rate (0.0214 cm yr^{-1}) were used to derive a term for dry mass flux (6.86 mg cm^{-2} yr^{-1}). The cumulative dry mass of the 3.75 cm of gyttja having accumulated above the glaciolacustrine unit was calculated as the summed product of depth and dry density for measurements over this interval ($n = 8$), resulting in a value of 0.308 g cm^{-2}, which was then divided by the average Holocene dry mass flux (6.86 mg cm^{-2} yr^{-1}), resulting in an age of 45 yr for this lithological transition (Fig. 2). However approximate the age of 45 yr may be for this hydrological change, given the similar properties between the gyttja in the gravity and percussion cores (Figs. 3 and 4) and the "modern" ^{14}C age at 2 cm (Table 2), this exercise strongly suggests that the lithological transition occurred recently in the 20th century.

Sediment Physical Properties

For the most part, the sediment physical properties (Fig. 3) closely reflect sediment lithology. The glaciomarine unit carries a low MS signal and is of intermediate density (1.4–1.8 g cm^{-3}) and hygroscopic moisture content (1–2%). The onset of deposition of the lacustrine gyttja is marked by a decrease in sediment density and a progressive increase of hygroscopic moisture. Toward the top of the gyttja, the sediment physical properties begin to change, well before the abrupt transition to minerogenic glaciolacustrine sediment. For example, moisture content begins to decrease around 60 cm, followed by rapid increases of MS and sediment density, both of which likely reflect an enhanced proportion of sand (Fig. 3D). After deposition of the minerogenic unit began, large proportions of clay were delivered, resulting in increased sediment density and partial dilution of the MS signal (Fig. 3). We infer that this clay-rich sediment was deposited due to advance of Nordre Qipisarqo Bræ into the lake's catchment based on physical differences from modern (i.e., gravity-core top) sediments (Figs. 2, 3). The physical characteristics of sediments deposited after retreat of the glacier from the catchment closely resemble those of percussion-core gyttja (Fig. 3).

Biologically Mediated Sediment Properties

LOI and $BSiO_2$ are very tightly coupled ($r^2 = 0.93$) throughout the Qipisarqo Lake sequence. After a marked, gradual increase from ∼180 to 125 cm, both proxies maintain high values (∼20%) between 125 and 60 cm in the percussion core, which subsequently drop dramatically in the uppermost 30 cm of gyttja that immediately underlie the glaciolacustrine sediment (Fig. 4). This trend is punctuated by two minor recoveries prior to the lithological change at ∼24 cm. Minimum values of LOI and $BSiO_2$ are attained in the glaciolacustrine unit, although it is noteworthy that $BSiO_2$ begins to recover before the return to organic sedimentation. Finally, there is striking, overall similarity between trends of Al and $BSiO_2$ (and LOI). Aluminum in lake sediments has both allogenic and authigenic sources (Engstrom and Wright, 1984), both of which are likely represented by the total Al concentrations from Qipisarqo Lake. Although there is much variability in parts of the Al profile, based on the pronounced decrease in the 60–25 cm interval which changes in concert with decreasing $BSiO_2$, despite enhanced (sand and silt) minerogenic input (Fig. 3), we infer that the Al stratigraphy is primarily influenced by weathering on the landscape and, perhaps more important, autochthonous source pools such as chelation by aquatic humic substances.

DISCUSSION

Lake Production and Paleoclimate

Given that $BSiO_2$ is a direct measure of the paleoproduction of siliceous algae, it is by definition a proxy of purely aquatic origin (Conley, 1988). Biogenic silica in Qipisarqo Lake sediments is primarily composed of diatom valves, with occasional chrysophyte cysts, but no chrysophyte scales. Close correspondence between $BSiO_2$ and LOI thus imply that much of the Qipisarqo Lake sediment organic matter is also of aquatic origin, although the ^{14}C dating results suggest that fluxes of (aged) terrestrial dissolved organic carbon must also be considered. Given that LOI values may also be a function of allochtonous organic material and the rate of in-lake decomposition of organic matter, we strongly rely on $BSiO_2$ measurements when inferring changes in paleoproduction and assume LOI faithfully tracks biogenic silica. In all likelihood, diatom and chrysophyte paleoproduction covaries with that of the soft-bodied algae in this oligotrophic lake, resulting in increased autochthonous organic matter sedimentation in response to elevated total algal production. Thus, the diatoms themselves are likely also important vectors for sediment organic matter accumulation and dominantly influence LOI values in Figure 4. This is especially true given that the dominant diatoms in Qipisarqo Lake are relatively large and voluminous taxa (e.g., *Pinnularia biceps, Aulacoseira* spp., *Tabellaria flocculosa;* unpublished data), implying a high ratio of protoplasm mass to silica on a per cell basis.

In several previous studies, both $BSiO_2$ (Colman *et al.*, 1995; Edlund and Stoermer, 2000) and LOI (Levesque *et al.*, 1993; Willemse and Törnqvist, 1999) have been interpreted as direct proxies of paleoclimatic variability. Such inferences assume a strong linkage between lake production and air temperature, which is supported by long-term monitoring studies (Schindler *et al.*, 1990). The results from Qipisarqo Lake are entirely compatible with these notions, given the striking similarity between the lake's $BSiO_2$ record and the borehole paleothermometry reconstruction from Dye3 (Dahl-Jensen *et al.*, 1998), 500 km to the north (Figs. 1 and 5B, 5D). Unless a common paleoclimatic history is invoked, there is no *a priori* basis for any linkage between Qipisarqo Lake sediment proxies and the ice sheet's thermal history. Hence, the amplitude-locked correspondence between these records lends reassurance that climate history is indeed recorded by the lake sediment record.

Holocene Climatic Evolution of Southern Greenland

Following glacio-isostatic isolation of the basin from the sea ca. 9100 cal yr B.P., Qipisarqo Lake records progressive warming over the next three millennia. This warming is lagged relative to summer insolation high by about 4000 yr (Fig. 5). In all likelihood, proximity to the Greenland Ice Sheet, and perhaps even the Laurentide, restrained early Holocene warming locally, through both direct atmospheric cooling as well as indirect effects involving meltwater and iceberg discharge into the neighboring ocean (Manabe and Broccoli, 1985). Despite the direct proximity of Qipisarqo Lake to the ice margin, there is no pronounced lithostratigraphic evidence for an early Holocene glacial advance corresponding to the regional North Atlantic cooling episode centered on 8200 cal yr B.P. (Alley *et al.*, 1997; Barber *et al.*, 1999). Although the $BSiO_2$ sampling density in this section of the core may be insufficient to document short-lived events, LOI and other physical proxies suggest that southern Greenland remained relatively cool until ca. 8000 cal yr B.P. (Figs. 3 and 4). The interval from 6000 to 3000 cal yr B.P. was marked by warmth and stability, and as such encompasses the Holocene thermal maximum expressed in southern Greenland. This interval integrates the influences of moderately high summer insolation and a warmer surrounding ocean, for which significantly reduced sea ice minimized the importance of high-albedo feedbacks (Kerwin *et al.*, 1999).

Marked cooling is recorded shortly after 3000 cal yr B.P. in the sediments of Qipisarqo Lake, corresponding closely with Neoglacial glacier advances regionally (Kelly, 1985; Fig. 5E) and likely in part a response to decreased solar insolation (Fig. 5A). The 3000–2000 cal yr B.P. interval is potentially a crucial interval of Neoglacial cooling, because it is so clearly manifested at a broad geographical scale around the Labrador Sea and Greenland (Fig. 5). On eastern Baffin Island, for example, molluscs with Arctic ecological affinities migrated southward

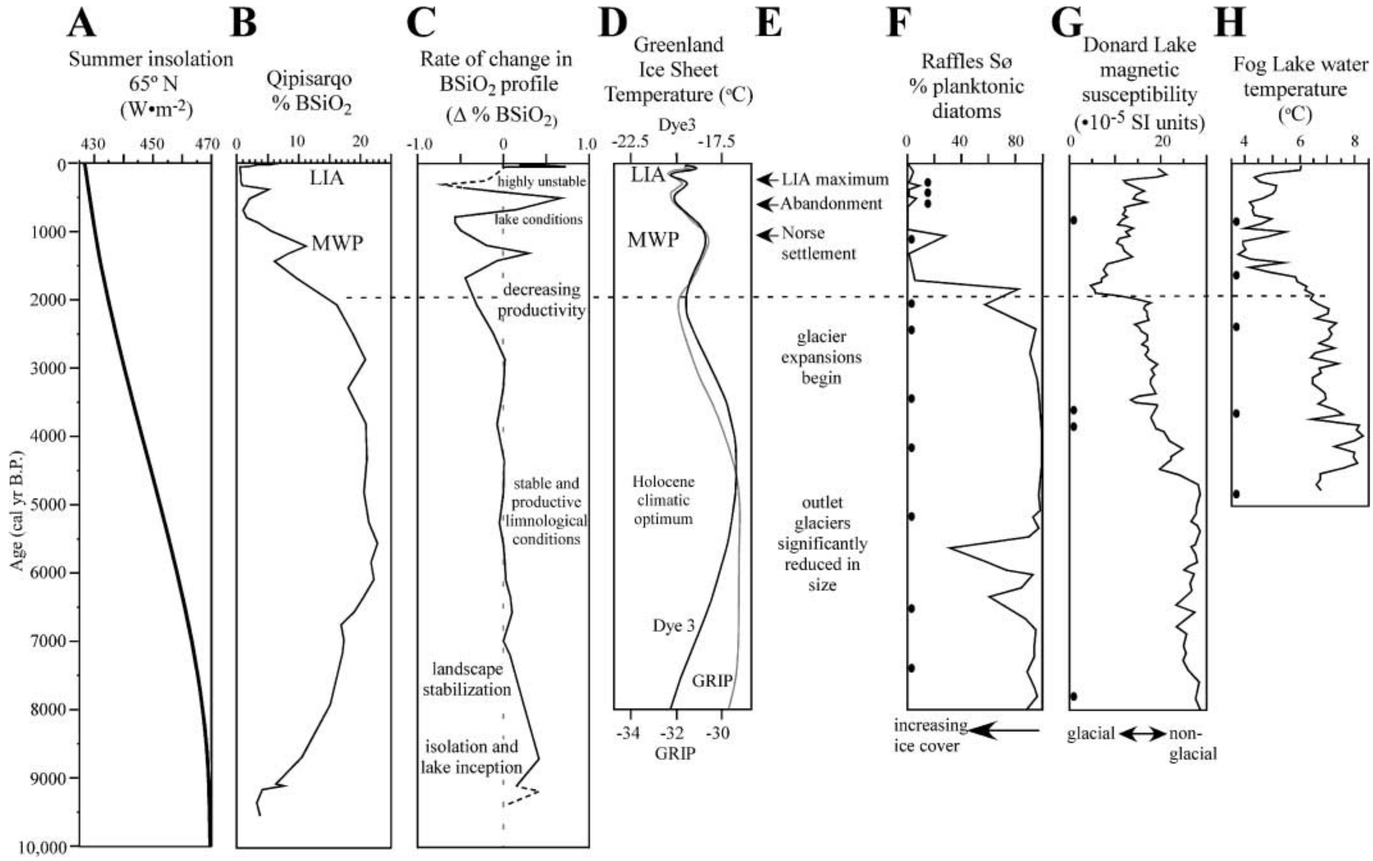

FIG. 5. Northern Hemisphere high-latitude summer insolation (A; Berger, 1978) shown in relation to Qipisarqo Lake $BSiO_2$ raw data (B) and rates of change (C); borehole paleothermometry from the Greenland Ice Sheet (D; Dahl-Jensen *et al.*, 1998); the glacial geological record of west Greenland (E; Kelly, 1985) and documented Norse chronology for eastern settlement (E; see Ogilvie *et al.*, 2000); the diatom record from Raffles Sø, east Greenland (F; Cremer *et al.*, 2001), and two records from Cumberland Peninsula, Baffin Island: magnetic susceptibility from Donard Lake (G; Moore, 1996) and diatom-inferred water temperatures from Fog Lake (H; Joynt and Wolfe, 2001). Black circles in F, G, and H correspond to radiocarbon ages. Horizontal dashed line across the figure highlights the dramatic 3000–2000 cal yr B.P. cooling observed in the Greenland/Baffin Bay region. On the rate of change in $BSiO_2$ curve (C), dashed sections (including glaciolacustrine and glaciomarine sediments) may not be directly comparable to the solid line (gyttja) because they reflect parts of the cores where sediment accumulation rates were dramatically higher (Fig. 2).

along the coast (Dyke *et al.*, 1996), and both glacial (Moore, 1996) and nonglacial (Joynt and Wolfe, 2001; Wolfe, 2002) lakes on Cumberland Peninsula record pronounced cooling. Synchronous cooling is also registered by diatom assemblage changes and decreased sediment $BSiO_2$ on Jameson Land (Cremer *et al.*, 2001), 1800 km to the northeast of Qipisarqo Lake. This geographic pattern of change (Figs. 1 and 5) implies that the East Greenland, West Greenland, and Baffin currents cooled synchronously in the late Holocene, perhaps in response to an enhanced delivery of Arctic Ocean waters through Denmark Strait and northern Baffin Bay. The corollary to this is that North Atlantic waters delivered via the Irminger Current had diminished influence on the climate of the southern and eastern Greenland coast during this interval.

Neoglacial cooling continued throughout the late Holocene until its culmination during the LIA, although two reversals in the $BSiO_2$ and LOI records, dated 1300–900 cal yr B.P. and 500–280 cal yr B.P., punctuate this overall trend and indicate a climate not simply forced by decreased insolation. The earlier of these events represents the limnological response to the Medieval Warm Period (MWP), associated with a temperature rise over southern Greenland of approximately 1.5°C from 2000 to 1100 cal yr B.P. (Dahl-Jensen *et al.*, 1998). The second warming reversal is also coherently recorded in the borehole record (Figs. 5C and 5D), and coincides with brief ameliorated conditions indicated by many European records (e.g., Grove, 1988); at Dye3, Dahl-Jensen *et al.* (1998) reconstruct a warming of up to 1°C between A.D. 1500 and A.D. 1750, prior to pronounced cooling during the second and more severe portion of the LIA. The Qipisarqo Lake lithostratigraphy demonstrates that the last cold interval produced the most extensive Holocene glacial position, as this was the only time the ice sheet's outlet glacier advanced into the lake's catchment and produced a diagnostic sedimentological signature. The evidence indicates that the outlet glacier's advance into the lake's catchment was relatively abrupt (i.e., a short time before the lithologic transition).

Changes in sediment characteristics do record the initial destabilization of the catchment, leading to increased erosion, well before the glacier advanced into the lake's drainage (Fig. 3). Although volcanic and solar forcing mechanisms appear sufficient to account for LIA cooling (Crowley, 2000), a possible linkage to reduced North Atlantic Deep Water formation has also been suggested, but not confirmed (Keigwin and Boyle, 2000).

The subsequent retreat of Nordre Qipisarqo Bræ from the lake basin is entirely consistent with naturally initiated post-LIA warming since A.D. 1850, which is recorded throughout the Arctic (Overpeck *et al.*, 1997). Continued warming led to a return to nonglacial sedimentation in the mid-20th century, by which time anthropogenic greenhouse gases had assumed an important role in climate forcing (e.g., Crowley, 2000). The 45-yr age estimate for initiation of 20th-century gyttja deposition is not necessarily associated with any specific climate event, because glacial retreat leading up to the rerouting of proglacial meltwaters away from the Qipisarqo Lake basin must have been initiated as a response to the 19th-century warming (Overpeck *et al.*, 1997). It is also particularly noteworthy that none of the paleoproduction proxies from Qipisarqo Lake have rebounded to their 6000 cal yr B.P. maximum, implying that current warming, however rapid, has not yet reached peak Holocene warmth.

Rates of Environmental Change and the Context of Norse Settlement

The overall character of Holocene environmental variability on southern Greenland is well illustrated by the first derivative of the raw Qipisarqo Lake $BSiO_2$ time series (Methods; Fig. 5C), which expresses the rate of change in a proxy that is demonstrably faithful in tracking climate. The rate of change in $BSiO_2$ between the lacustrine gyttja, glaciolacustrine, and glaciomarine sediments may not be directly comparable (Fig. 5) given that sediment accumulation rates are likely to have been dramatically higher in the non-gyttja units (Fig. 2). Nonetheless, this analysis (even if confined to the lacustrine sediments) indicates that the late Holocene (3000–0 cal yr B.P.) is inherently less stable than the previous six millennia and that this instability appears to be ever-increasing, with the highest rates of change confined to the last 1000 yr. These observations are contrary to the intuitive notion of greater climatic instability during the early Holocene, which is frequently observed, due to the potential impacts of disintegrating vestigial Wisconsinan ice masses (e.g., Barber *et al.*, 1999). Major climate reversals during the final stages of deglaciation are indeed firmly recorded at the summit of the Greenland Ice Sheet (Alley *et al.*, 1997), and the evidence presented in Figures 3 and 4 suggests southern Greenland remained relatively cold until close to 8000 cal yr B.P. However, it is restated that for at least the last 9000 cal yr the sensitively positioned Qipisarqo Lake basin records only climate change associated with the LIA as pronounced enough for outlet ice to reach the lake.

When viewed in relationship to the Qipisarqo Lake and other records with sufficient temporal resolution (Fig. 5), it seems probable that climate played an important role in the history of the Norse settlements in Greenland and perhaps Newfoundland (Buckland *et al.*, 1996; Barlow *et al.*, 1997; Ogilvie *et al.*, 2000). Our $BSiO_2$ rate of change curve strongly suggests that Norse movement around the region occurred at perhaps the worst time in the last 10,000 yr, in terms of the overall stability of the environment for sustained plant and animal husbandry. The Norse established eastern settlement less than 150 km from Qipisarqo Lake around A.D. 985 (Fig. 1). The western settlement was also established around this time, 1000 km north of Qipisarqo Lake, but was then abandoned ca. A.D. 1350, whereas the larger eastern settlement remained until ca. A.D. 1450–1550 (Ogilvie *et al.*, 2000). Colonization around the northwestern North Atlantic occurred during peak MWP conditions that ended in southern Greenland by A.D. 1100. Abandonment of the settlements coincided with subsequent initial LIA cooling which persisted for the next four centuries. The climatic amelioration around A.D. 1500 may have been too late or of insufficient magnitude to reverse the settlements' demise. Our conclusions are consistent with those of Barlow *et al.* (1997), who carried out a comprehensive study of (the) western settlement that included modeling the detrimental effects of deteriorating climate on farming.

CONCLUSIONS

Qipisarqo Lake sediment characteristics have been tightly coupled to lake production and Holocene climatic evolution, thereby validating the utility of proxies such as $BSiO_2$ in paleoclimatic applications from Arctic lake sediments. The reconstruction highlights that following a prolonged interval of warm and stable conditions during the middle Holocene, a progression toward increasingly unstable environmental conditions in the late Holocene, likely in part tracking decreased solar insolation. The pronounced interval of Neoglacial cooling between 3000 and 2000 cal yr B.P. appears to be coherently registered in east Greenland as well as both sides of Baffin Bay, suggesting a climate history not just linearly influenced by Milankovitch forcing and a teleconnection involving decreased ocean heat transport in the entire East Greenland–West Greenland–Baffin current system. Following the MWP the LIA contains the most severe climatic conditions regionally of the entire Holocene. Although the congruous timing of Norse history and climate change could be considered circumstantial, at the very least, environmental conditions during this time were highly variable which must have been unfavorable for long-term plant and animal husbandry.

Present and future environmental changes in southern Greenland will likely be superposed on high degrees of already inherent variability. In addition, although lake productivity has not yet reached peak values achieved 6000 cal yr B.P., given the probability of significant future anthropogenic warming (Wigley and Raper, 2001), it seems plausible that the two most extreme Holocene climate states in southern Greenland may occur successively within less than 500 yr (cf., Bradley, 2000). The rates and amplitudes of such changes will likely induce ecological changes for which there are no adequate past analogs. Evidence is mounting from lakes in adjacent parts of the Arctic (Douglas *et al.*, 1994; Overpeck *et al.*, 1997; Wolfe and Perren, 2001) that such changes may already be arising.

ACKNOWLEDGMENTS

Primary support by NSF-ATM 9709918 and, secondary, a Weeks Postdoctoral Fellowship from the Department of Geology and Geophysics, University of Wisconsin-Madison (Kaplan). We thank Jocelyn Turnbull and Scott Lehman for ^{14}C dating, Wendy Freeman for sedimentological analyses, Dierck Kersten and Peter Foister for core sampling assistance, Fred Luiszer for ICP-AES measurements, David Bahr for mathematical assistance, and John Smol and an anonymous reviewer for helping to improve the manuscript. Fieldwork was made possible by Grønlandsfly helicopters and the assistance of Ole Bennike, Jonathan Overpeck, and Alexis Robertson. This is PALE/PARCS contribution number 184.

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