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Holocene glaciation and climate evolution of Baffin Island, Arctic Canada

Gifford H. Miller[a,*], Alexander P. Wolfe[b], Jason P. Briner[a,e], Peter E. Sauer[c], Atle Nesje[d]

[a]INSTAAR and Department of Geological Sciences, University of Colorado, 1560 30th St. Campus Box 450, Boulder, CO 80303, USA
[b]Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, AB, T6G 2E3 Canada
[c]Department of Geological Sciences, Indiana University, Bloomington, IN 47405-1405, USA
[d]Department of Earth Science, University of Bergen, Allégaten 41, 5007 Bergen, Norway
[e]Department of Geology, State University of New York at Buffalo, Buffalo, NY 14260 USA

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Abstract

Lake sediment cores and cosmogenic exposure (CE) dates constrain the pattern of deglaciation and evolution of climate across Baffin Island since the last glacial maximum (LGM). CE dating of erratics demonstrates that the northeastern coastal lowlands became ice-free ca.14 ka as the Laurentide Ice Sheet (LIS) receded from its LGM margin on the continental shelf. Coastal lakes in southeastern Baffin Island started to accumulate sediment at this time, whereas initial lacustrine sedimentation was delayed by two millennia in the north. Reduced organic matter in lake sediment deposited during the Younger Dryas chron, and the lack of a glacial readvance at that time suggest cold summers and reduced snowfall. Ice retreated rapidly after 11 ka but was interrupted by a widespread readvance of both the LIS and local mountain glaciers ~9.6 ka (Cockburn Substage). A second readvance occurred just before 8 ka during a period of unusually cold summers, corresponding to the 8.2 ka cold event in the Greenland Ice Sheet. Most local glaciers were behind their present margins before 7 ka, and in some instances much earlier, although the Foxe Dome of the LIS continued to slowly retract toward the present day Barnes Ice Cap throughout the Holocene.

Pollen in lake sediments is rare and dominated by exotic sources prior to 12 ka. Subsequently, grass tundra became established, followed by modern tundra vegetation ca. 8 ka, with subtle changes in pollen assemblages in the late Holocene. Lake primary productivity peaked in the early Holocene, before terrestrial vegetation or marine surface waters reached their apparent thermal maxima. Lacustrine, marine, and glacial proxies all reflect significant late Holocene cooling. The onset of Neoglaciation is well dated in lacustrine records at ca. 6 ka, with intensification after 2.5 ka. The expansion of local glaciers during the Little Ice Age represents the most extensive advance since 7 ka. We suggest that the replacement of Atlantic surface waters by cold, low-salinity Arctic Ocean water, coupled with the steady reduction of summer insolation, resulted in a significant positive sea-ice feedback that produced a larger late Holocene summer temperature depression over the Baffin region than in the Pacific sector of the Arctic.

1. Introduction

Baffin Island is the largest island in the Canadian Arctic Archipelago, stretching 1600 km in length and varying from 300 to 600 km in width. The eastern rim of Baffin Island was uplifted when Greenland was rifted from North America in the early Tertiary, and the uplifted rim was subsequently dissected by fluvial and glacial erosion into rugged mountainous terrain and fiords that connect the interior plateau to the adjacent ocean. Baffin Island is everywhere above treeline, and all but the southernmost portion is in the zone of continuous permafrost. Vegetation across southern Baffin Island is classified as Low Arctic, most of central

*Corresponding author. Tel.: +1 303 492 6387; fax: +1 303 492 6388.
E-mail addresses: gmiller@colorado.edu (G.H. Miller), awolfe@ualberta.ca (A.P. Wolfe), pesauer@indiana.edu (P.E. Sauer), atle.nesje@geo.uib.no (A. Nesje).

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Baffin Island is in the Middle Arctic zone, whereas the northern third is High Arctic (Polunin, 1951). Mean annual temperatures decrease from −5 °C in the south to −15 °C in the north (ESWG, 1995). Mean July temperatures along the coast, where most of the recording stations are situated, are around 4 °C with little latitudinal trend, but interior sites are frequently warmer. Annual precipitation is between 200 and 300 mm for most of the island, with localized maxima of 500–600 mm close to open water in winter (ESWG, 1995). The Barnes and Penny ice caps (both ca. 6000 km^2) are the largest ice caps; smaller mountain ice caps, cirque, and valley glaciers are common along the eastern mountains.

Holocene (the last 11.5 kyr[1]) climate change in the Eastern Canadian Arctic has been influenced by three dominant factors: (1) the impact on atmospheric circulation by the waning Laurentide Ice Sheet (LIS), (2) the steady decrease in summer solar insolation as regular changes in the Earth's orbit diminished Northern Hemisphere seasonality (11% less summer insolation at 70°N at present than at the start of the Holocene), and (3) changes in the balance between relatively warm, salty Atlantic water (West Greenland Current) and the outflow of cold, low-salinity surface water from the Arctic Ocean (Baffin Current; Fig. 1).

Changes in the terrestrial environment across Baffin Island have been reconstructed from surficial mapping of glacial features in the field and on aerial photographs, and from a wide range of climate proxies preserved in lacustrine archives, including biotic (pollen, diatoms, organic content, biogenic silica) and physical (magnetics, grain size, mineralogy) properties. At the peak of the last glacial maximum (LGM), warm-based outlet glaciers occupied most fiords and sounds, but significant portions of Baffin Island were covered by cold-based ice that left few traces of its passage. In these areas, cosmogenic exposure (CE) dating of erratic blocks forms the basis of the deglacial reconstruction (Briner et al., 2003; Briner et al., 2005; Davis et al., in press). By the onset of the Holocene, the Eastern Canadian Arctic had warmed sufficiently that most glaciers were warm-based, and deglaciation can be deciphered from moraines, ice-contact deposits, and ice-dammed lakes, and from diagnostic physical characteristics in lacustrine archives, supplemented by marine seismic surveys and marine sediment cores. Chronological control is provided primarily by radiocarbon dating of organic remains in lake sediment and marine bivalves in marine cores and raised marine deposits.

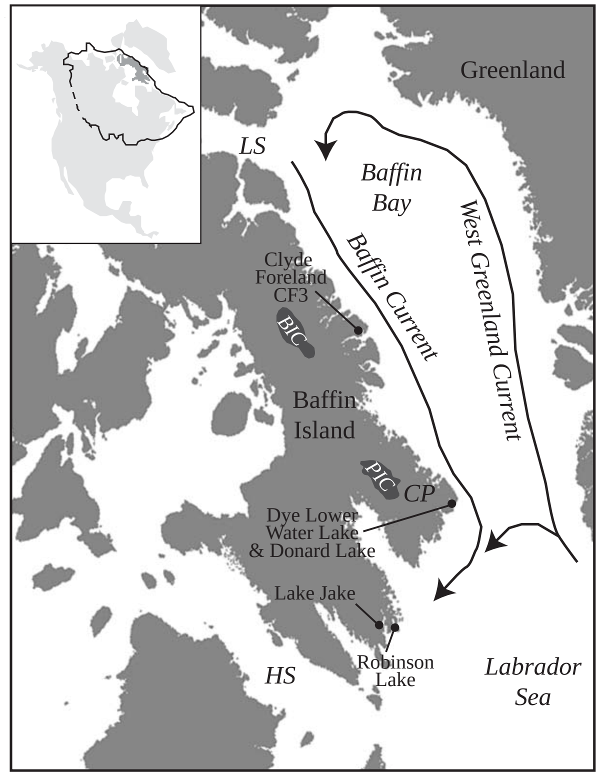

Fig. 1. Baffin Island, eastern Canadian Arctic (Nunavut), showing the locations of lakes and marine currents discussed in the text. LS = Lancaster Sound, BIC = Barnes Ice Cap, PIC = Penny Ice Cap, CP = Cumberland Peninsula, HS = Hudson Strait. Inset shows Baffin Island in relation to the generalized outline of the Laurentide Ice Sheet at the LGM.

The Foxe Dome of the LIS covered almost all of Baffin Island during repeated glaciations of the Quaternary, terminating along the eastern coast or on the adjacent continental shelf. This margin of the LIS is comparable in length to the well-studied southern margin of the LIS that terminated in the northern United States, but has received only a fraction of the research effort. After years of debate, a general consensus on the approximate limit of the LIS margin at the LGM has emerged (Dyke et al., 2002; Miller et al., 2002). The Foxe Dome contributed to large outlet glaciers that occupied deep troughs north (Lancaster Sound) and south (Hudson Strait) of Baffin Island, and fed relatively fast-moving outlet glaciers that occupied the fiords and sounds of eastern Baffin Island. These outlet glaciers terminated on the continental shelf, but probably did not reach the shelf break, except in the largest of the offshore troughs. Many inter-fiord uplands and much of eastern Cumberland Peninsula were not inundated by Laurentide ice at the LGM.

At the onset of the Holocene, the Foxe Dome was still connected to the rest of the LIS, and was probably contiguous with the Innuitian Ice Sheet over the Queen Elizabeth Islands to its north (Dyke et al., 2002). Sea

[1]All ages are in calendar years before present (1950 AD), with conventional radiocarbon dates calibrated using Calib 4.4.1 (http://depts.washington.edu/qil/calib/). kyr = thousand years; ka = thousands of years ago.

level stood ca. 30 m below present, but summer insolation was at a peak. By 8 ka the sea had penetrated the interior of the LIS (Barber et al., 1999). Marine waters finally penetrated Foxe Basin 7.5 ka, leaving a residual ice dome over Baffin Island that slowly retracted, with some minor readvances, toward the Barnes Ice Cap, which has been relatively stable for the last 1 or 2 millennia. The presence of highly depleted $\delta^{18}O$ in basal ice of the Barnes Ice Cap (Hooke and Clausen, 1982) confirms that the ice cap is the final remnant of the LIS. Pleistocene ice is also present in the Penny Ice Cap (Fisher et al., 1998) and may be present in some of the larger mountain ice caps.

Recent summaries of Holocene climate change in the North American Arctic (CAPE, 2001; Gajewski and Atkinson, 2003; Kerwin et al., 2004) rely primarily on changes in diagnostic pollen assemblages to reconstruct summer temperatures. Kaufman et al. (2004) use pollen as well as a wider array of proxy data to show a fundamental asynchrony in the timing of the Holocene thermal maximum across the North American Arctic, and a larger amplitude late Holocene cooling over the Atlantic sector.

The goal of this paper is to summarize the main characteristics of the glacial and climate history of Baffin Island since the LGM. We do this by first presenting biotic and physical climate proxy data derived from six lacustrine sediment cores recovered from four new sites on Baffin Island. The paleoenvironmental implications of these new data are then combined with results published elsewhere to develop a regional picture of conditions during deglaciation, the subsequent rise to the Holocene thermal maximum, the onset of Neoglaciation, and its intensification in the late Holocene.

2. Methods

2.1. Sediment cores

Lacustrine sedimentary archives offer the best opportunity to reconstruct terrestrial environmental change in the Baffin region. They provide continuous, datable records incorporating a suite of proxies, most of which are related to summer temperature. We targeted lakes for sediment cores that were deemed to have a high probability of recording either Holocene glacial activity (glacier-dominated lacustrine systems) or Holocene climate change (small lakes lacking large catchments).

Sediment cores were recovered using a modified hammer-driven piston corer (Nesje, 1992) of 7 or 11 cm diameter, using lake ice as a coring platform. This system enables recovery of a wide range of sediment types, including stony diamicton, in up to 100 m water depth. In most Baffin Island lakes the entire sediment fill can be recovered in single core of no more than 3 m length, although some glacier-fed lakes contain more than 3 m of sediment fill. In many lakes we supplemented piston coring with surface sediment cores (Glew, 1991) to capture the sediment–water interface and the top 30–40 cm of sediment. These cores include the uppermost 5–10 cm that are typically lost or disturbed in piston coring, and sufficient additional sediment to splice the two cores together. A box corer ($10 \times 10 \times 25$ cm) was used to recover surface sediment in some lakes, typically the uppermost 10–15 cm.

2.2. Geochronology

Dating arctic lake sediment poses unusual challenges (Wolfe et al., 2005). Radiocarbon dating is complicated by relatively low within-lake primary productivity, especially in the ultra oligotrophic lakes on Baffin Island. Furthermore, low rates of terrestrial plant decomposition results in a relatively large pool of "aged" carbon in catchment soils. Delivery of both particulate organic carbon (POC) and dissolved organic carbon (DOC) by stream and slopewash processes results in pools of POC and DOC in lakewaters that are depleted in ^{14}C activity relative to the atmospheric reservoir (Abbott and Stafford, 1996). This is in contrast to the ^{14}C activity of dissolved inorganic carbon (DIC) in lakes that in most instances is equilibrated with the atmospheric. This equilibration is demonstrated by dating living aquatic moss collected from seven Baffin Island lakes that all had ^{14}C activities indistinguishable from that of the contemporary atmosphere (Miller et al., 1999). Consequently, aquatic organisms utilizing DIC as their primary carbon source have the same ^{14}C activity as the contemporaneous atmosphere, and are a primary target for dating. All lakes in this study reside in the Precambrian crystalline terrain of the Canadian Shield, and only rarely does the surrounding drift contain any carbonate. Consequently, there is rarely any hardwater effect.

Based on our assessment of carbon reservoirs, fragments of aquatic moss are the target medium for ^{14}C dating; chironomid head capsules are equally suitable (Fallu et al., 2004), but more difficult to isolate in sufficient mass. When macrofossils are not present, humic acid extracts provide the most reliable medium for dating the bulk DOC fraction (Abbott and Stafford, 1996). Miller et al. (1999) provide a detailed study of Robinson Lake (Fig. 1) showing that in the first millennium after deglaciation, macrofossil and humic acids have the same ^{14}C activity, but as soils develop the DOC fractions become increasingly depleted in ^{14}C relative to macrofossils from the same levels, eventually reaching an offset of 300 ± 300 years. To correct for this offset, we apply a reservoir correction to humic acid dates by subtracting 300 years from the conventional

Table 1
Radiocarbon dates on organic material from lake sediments reported in the text

Depth (cm)	Material dated	Laboratory ID	$\delta^{13}C^{a}$ (‰)	Conventional radiocarbon age	Conventional age after HA reservoir correction	Calibrated age[b]
Donard Lake (95-DON-03)						
262	Humic acids	CAMS-27265	(−25)	6240±50	5940±300	7150±91
319	Terrestrial plant	AA-17952	−27.3	8940±80		10,095±95
333	Humic acids	CAMS-27266	(−25)	11,790±70		13,855±171
345	Humic acids	CAMS-23555	(−25)	11,970±60		13,737±75
356	Aquatic moss	CAMS-23554	(−25)	12,600±60		15,134±335
Donard Lake (94-DON-01)						
47	Aquatic moss	AA-15684	−28.3	935±60		897±22
221	Aquatic moss	AA-15685	(−25)	4210±80		4806±41
232	Aquatic moss	AA-15686	(−25)	4650±90		5531±46
Dye Lower Water Lake (93-DLW-03)						
34.5	Aquatic moss	CAMS-7787	(−25)	2380±70		2595±95
140.5	Aquatic moss	CAMS-7788	(−25)	6880±90		7715±75
159	Humic acids	CAMS-7783	(−25)	8510±70		9510±40
159	Humins	CAMS-7786	(−25)	8690±70		
0	Living aquatic moss	CAMS-12259	(−25)	Fm = 1.1079±0.0073		
0	Lakewater inorganic carbon	CAMS-12293	(0)	Fm = 1.0607±0.0072		
Lake Jake (91-LJ1)						
6.5	Humic acids	CAMS-17143	(−25)	2410±60	2110±300	2120±370
15.5	Humic acids	CAMS-17139	(−25)	4260±60	3960±300	4410±400
25.5	Humic acids	CAMS-17145	(−25)	5290±50	4990±300	5829±60
35.5	Humic acids	CAMS-17140	(−25)	6440±60	6140±300	7094±63
45.5	Humic acids	CAMS-17144	(−25)	7370±50	7070±300	7872±100
55.5	Humic acids	CAMS-17142	(−25)	8760±170	8460±300	9415±244
65.5	Humic acids	CAMS-17141	(−25)	9010±60	8710±300	9651±98
Lake Jake (91-LJ3)						
25	Plant macrofossils	AA-10639	(−25)	2580±50		2645±120
85	Plant macrofossils	AA-10640	(−25)	5685±60		6455±90
115	Plant macrofossils	AA-10641	(−25)	7455±70		8265±70
165	Plant macrofossils	AA-10642	(−25)	7240±60		8105±50
175	Plant macrofossils	AA-10643	(−25)	7395±65		8200±125
Lake CF3 (02-CF3-01)						
2	Aquatic moss	NSRL-13325	−20.9	1240±30		1175±85
27	Aquatic moss	NSRL-13357	−23.6	3460±25		3730±85
60	Aquatic moss	NSRL-13356	(−25)	6450±35		7370±85
98	Aquatic moss	NSRL-13355	(−25)	8150±40		9130±105
121	Aquatic moss	NSRL-13324	−27.5	8520±45		9515±20
163	Aquatic moss	NSRL-13216	−26.0	9770±40		11,185±20
Large block of lacustrine sediment within Clyde Foreland Neoglacial Moraine						
	Humic acids	NSRL-12104	(−25)	3390±35	3090±300	3255±335

CAMS = Lawrence Livermore National Laboratory, AA = University of Arizona, NSRL = INSTAAR, University of Colorado for target preparation, measured at Woods Hole Oceanographic Institution.
[a]Measured; estimated if in parentheses.
[b]Calibrated age standard deviations are based on $\pm 1\,\sigma$ uncertainties on the uncalibrated date, and then dividing the calibrated age range in two.

humic acid date, and increasing the uncertainty to 300 years, except for humic acid dates within 1000 years of deglaciation which we treat similar to macrofossil dates. We report conventional ^{14}C ages, reservoir-corrected ^{14}C ages for humic acid dates, and calibrated ages (Table 1). For each lacustrine sediment core, an age model is derived from the calibrated ages fitted to an appropriate curve, usually a polynomial regression, that allows every depth to be given a unique calibrated age (Fig. 2). We assume the surface sediment is modern,

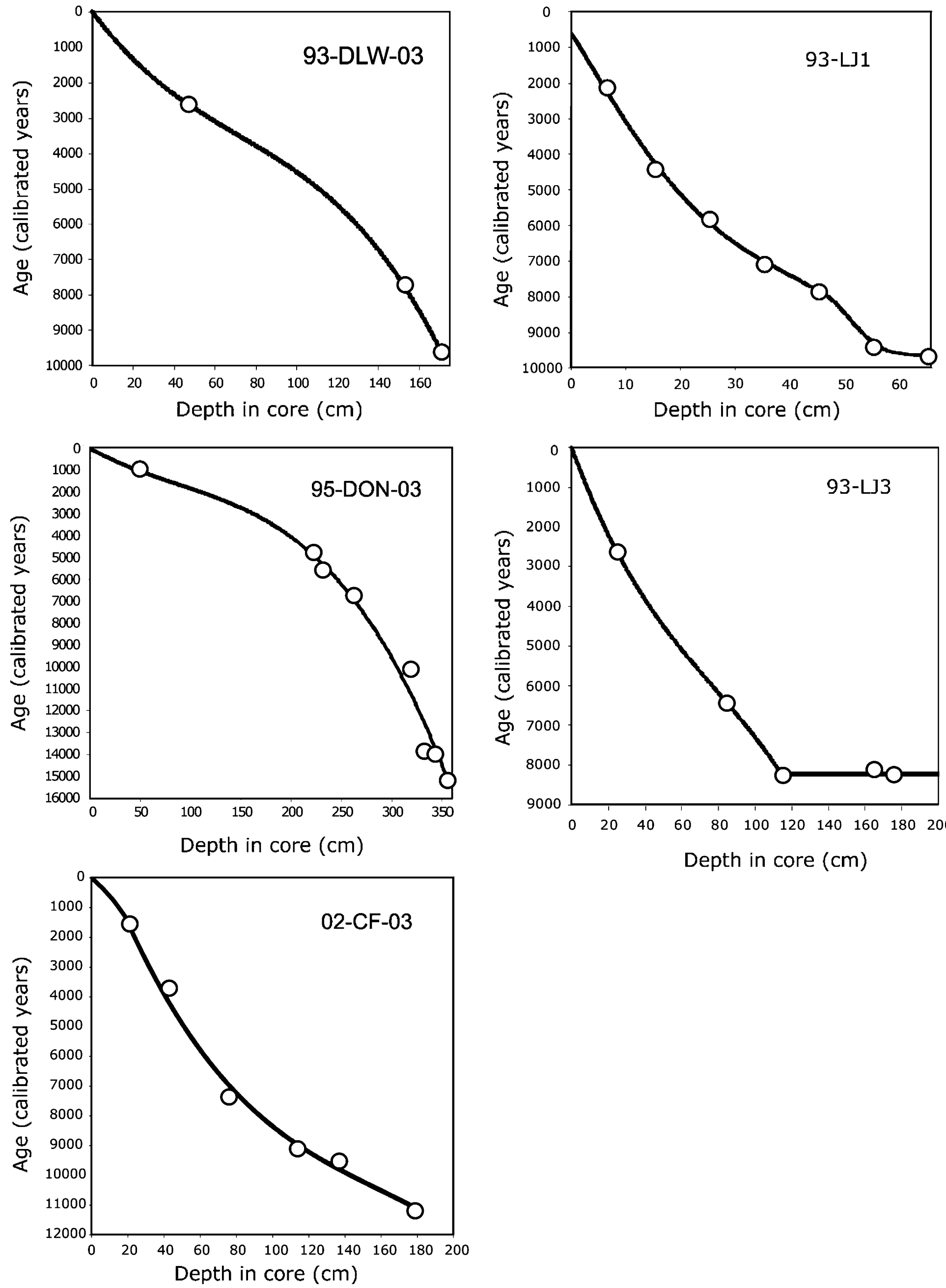

Fig. 2. Age–depth relations established for all cores presented in this paper. Dates are in calendar years (Table 1). Age and depth ranges and uncertainties are contained within the circles. Polynomial regressions (solid lines) are used to convert depth-dependent proxy data into the time domain.

unless compelling additional data suggests a significant temporal loss of surface sediment in the coring process.

Because of limited nutrients and brief ice-free seasons (generally less than 3 months/year), Baffin Island lakes contain a paucity of benthic organisms, resulting in minimal bioturbation. Even lakes consisting dominantly of gyttja that is not visibly stratified exhibits weak stratification in X-radiographs. Sedimentation rates in gyttja-dominated systems (isolated basins without significant catchments) are typically ca. 10 cm kyr^{-1}, with a few lakes as high as 20 cm kyr^{-1}. Despite low sedimentation rates, the lack of a significant benthos allows 0.5–1.0 cm-scale sampling, providing from 25 to 200 year resolution.

Over the past decade, CE dating has been applied widely to date glaciated terrain (see review by Gosse and Phillips, 2001). Most CE dates used to derive the deglacial history of Baffin Island are referred to their primary sources. Key unpublished dates from the Clyde region are given in Table 2.

Table 2
Cosmogenic exposure ages from the Clyde Foreland and Aston Lowland that define initial deglaciation of the coastal margin (arranged from youngest to oldest for each group)

Sample	Sample type	Lat. (N)	Long. (W)	Elevation (m asl)	^{10}Be (10^5 atoms g^{-1})	^{10}Be age (ka)
Coastal central Clyde Foreland						
AL12-01-2	Boulder	70° 42 19.1′	69° 05 13.9′	36	0.55±0.03	10.5±0.6
CF02-184	Cobble	70° 42.431′	69° 07.530′	40	0.61±0.06	11.9±0.5
AL2-01-2	Boulder	70° 35 05.6′	68° 55 43.3′	220	0.87±0.03	13.9±0.5
AL9-01-1	Boulder	70° 41 29.4′	69° 11 3.7′	178	0.88±0.03	14.6±0.5
AL12-01-1	Boulder	70° 42 23.2′	69° 05 17.5′	52	0.81±0.03	15.3±0.5
AL7-01-1	Boulder	70° 41 7.4′	69° 08 35.9′	93	0.86±0.05	15.5±0.8
AL14-01-2	Boulder	70° 37 35.9′	69° 10 25.8′	45	0.89±0.04	16.9±0.7
CF02-64	Boulder	70° 36.823′	68° 48.078′	98	0.94±0.09	17.0±0.7
AL7-01-2	Boulder	70° 40 58.3′	69° 09 42.9′	130	0.99±0.03	17.3±0.5
CF02-65	Boulder	70° 38.812′	68° 54.527′	85	1.22±0.11	22.7±0.7
AL8-01-1	Boulder	70° 41 29.9′	69° 10 23.4′	172	1.39±0.07	23.3±1.2
CF02-58	Cobble	70° 34.843′	68° 55.791′	300	1.58±0.05	23.8±0.7
AL10-01-1	Boulder	70° 41 31.1′	69° 12 3.9′	185	1.45±0.13	23.9±2.1
CF02-109	Cobble	70° 37.482′	69° 10.743′	50	2.59±0.22	50.5±1.3
AL2-01-1	Boulder	70° 35 16.4′	68° 59 39.8′	208	4.54±0.16	73.9±2.7
Coastal central Aston Lowland						
PT01-04	Boulder	69° 53.642′	67° 35.100′	85	0.66±0.03	11.9±0.5
CAD-COS-1	Cobble	69° 54.177′	67° 36.000′	87	0.71±0.09	13.1±1.2
PT01-07	Boulder	69° 54.452′	67° 33.833′	80	0.74±0.02	13.5±0.4
CAD02-8	Boulder	69° 52.079′	67° 27.388′	75	0.76±0.03	13.9±0.6
CAF02-1	Cobble	69° 55.210′	67° 33.066′	41	0.75±0.03	14.2±0.5
CAD02-5	Boulder	69° 53.164	67° 29.801′	79	0.78±0.03	14.4±0.6
CAD02-9	Boulder	69° 53.086′	67° 27.357′	47	0.79±0.04	14.9±0.7
CAD02-10	Boulder	69° 53.751′	67° 26.581′	37	0.88±0.03	16.8±0.6
PT01-06	Boulder	69° 54.627′	67° 33.888′	80	0.95±0.03	17.5±0.5
CAD02-6	Boulder	69° 53.164′	67° 29.801′	73	0.99± 0.03	18.4±0.6
CAD02-7	Cobble	69° 53.164′	67° 29.801′	73	1.06±0.04	19.6±0.7
CAD02-4	Boulder	69° 53.089′	67° 33.629′	78	1.42±0.04	25.8±0.7
PT01-05	Boulder	69° 53.700′	67° 34.972′	85	5.13±0.13	95.4±2.4

Sample processing at the University of Colorado, targets measured at Lawrence Livermoore National Laboratory.
Note: ^{10}Be production rate used is 5.1 atoms g^{-1} yr^{-1} (Stone, 2000). Isotope concentrations were scaled for elevation following Lal (1991) and Stone (2000).

2.3. Climate proxies

Data are presented for a variety of physical and biological parameters isolated from lacustrine sedimentary archives. Magnetic susceptibility (MS), grain size, loss-on-ignition (LOI), and palynology were all determined following standard laboratory procedures described elsewhere. Whole-core MS was measured in the field shortly after recovering the cores, and again in the laboratory at higher resolution. For one core (95-DON-03), both mass and volume MS and a full paleomagnetic analysis was completed. Pollen was extracted and identified in three cores. Interpretation of Baffin Island pollen spectra is complicated by over-representation of far-traveled wind-blown exotic pollen taxa. A regional survey of pollen from surface sediment in lakes across the eastern Canadian Arctic provides a more quantitative basis for climate interpretation of pollen data (Kerwin et al., 2004), although many pollen assemblages older than 8 ka lack modern analogs. LOI has been shown to be a reliable index of the total organic content of Baffin Island lake sediments, because the crystalline terrain of Baffin Island provides few clay minerals or other hydrated minerals that can obscure the relationship. In lakes with small catchments, most of the carbon in the sediment fill is fixed by primary producers within the lakes. Consequently, for these lakes, LOI is a reasonable proxy for within-lake biological productivity (e.g. Battarbee et al., 2001, 2002). Biological productivity in Baffin Island lakes occurs dominantly during the brief, ice-free summer season; unless nutrient sources change, LOI is controlled by the duration of the open-water season, a variable closely associated with mean summer temperature.

3. Lacustrine archives

Paleoenvironmental proxies are presented below for six lake sediment cores from four different lakes that range from southern to northeastern Baffin Island (Fig. 1). The most complete Holocene record is derived from Donard Lake, supplemented by cores from nearby Dye Lower Water Lake. Three cores from Lake Jake document an early Holocene glacier readvance, and a core from the Clyde Foreland illustrates the latitudinal gradient in the timing of the Holocene thermal maximum.

3.1. Donard lake

The Donard Lakes are a series of lakes near Cape Dyer, at the eastern edge of Cumberland Peninsula (Fig. 3). Only the uppermost basin has been cored, and we refer to this basin as Donard Lake (66° 40′N; 61° 47′W; 460 m asl). Immediately to the north, the Caribou Glacier occupies a valley running at right angles to the valley of Donard Lake, and is separated from it by a low col (Fig. 4).

Sediment flux to Donard Lake is determined by the thickness of the Caribou Glacier. Currently, the glacier is thick enough that it has breached the col separating it from the Donard Lakes drainage, and an arm of the Caribou Glacier terminates a few hundred meters from Donard Lake. In this state, most of the surface meltwater that follows the left margin of the Caribou Glacier is diverted into Donard Lake (Fig. 4), increasing the catchment area for Donard Lake by a factor of 10, and delivering glacier rock flour directly to the lake. When the Caribou Glacier is not thick enough to breach the col, meltwater bypasses the Donard Lake drainage, and is instead delivered to Sunneshine Fiord. Because of the unusual threshold relation between Donard Lake and Caribou Glacier, the lake provides an ideal situation to capture the onset of Neoglaciation. As long as Caribou Glacier remains below the col separating it from the Donard Lakes drainage, Donard Lake will receive sediment only from slopewash and a small headwater stream. Whenever Caribou Glacier thickens sufficiently to breach the col, Donard Lake is immediately transferred to a glacier-dominated system, with a high flux of minerogenic sediment. To capitalize on this unusual relation, we mounted coring campaigns to Donard Lake in 1994 and 1995. Additional descriptions of the lake are in Moore (1996) and Moore et al. (2001).

Donard Lake has a simple bathymetry, with a broad central basin 19–20 m deep, and a deepest hole just over 21 m (Fig. 4). Nine sediment cores were recovered in 1994 and 1995, primarily from 20 to 21 m water depth in the central basin, but also along a transect toward the

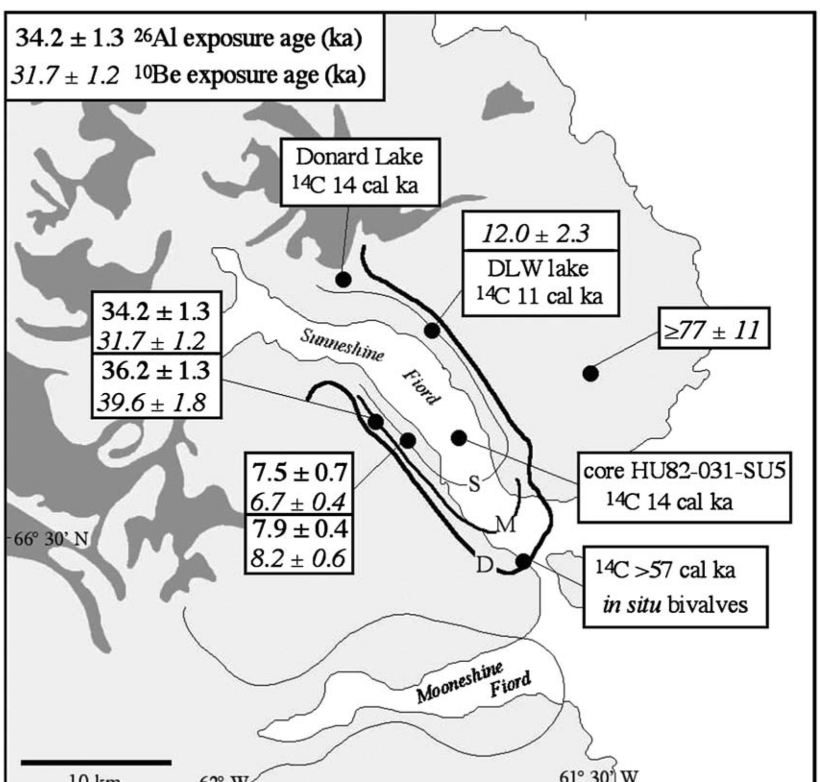

Fig. 3. Map of the Cape Dyer region, easternmost Cumberland Peninsula, showing the locations of Donard Lake and Dye Lower Water Lake, and a marine core from Sunneshine Fiord (Andrews et al., 1996), all with their basal radiocarbon ages. An ice-contact raised marine delta from outer Sunneshine Fiord was deposited >57 ka based on a ^{14}C age on in situ *Hiatella arctica*. CE ages (^{10}Be and ^{26}Al) on moraine boulders, and one age on weathered bedrock (⩾77 ka) are described in Miller et al. (2002). Three moraine systems mapped in Sunneshine Fiord (Locke III, 1987) include the Duval Moraine (D), >57 ka, the Mooneshine Moraine (M) stabilized ⩾35 ka, and the Sunneshine Moraine, of LGM age. Dye Lower Water Lake (DLW) is dammed between crests of the Duval Moraine. Modified from Miller et al. (2002).

modern glacier margin. Cores from the central basin reveal 2–2.5 m of continuously finely laminated sediment, overlying 1 or more meters of massive or faintly stratified sediment. Moore et al. (2001) showed that the upper 50 cm were annually laminated (varved), from which we conclude that the deeper levels with comparable laminations are also varved, and reflect sedimentation under a glacier-dominated sediment delivery system. These data suggest that once Caribou Glacier thickened sufficiently to breach the col and divert meltwater into Donard Lake, it remained in the catchment to the present day.

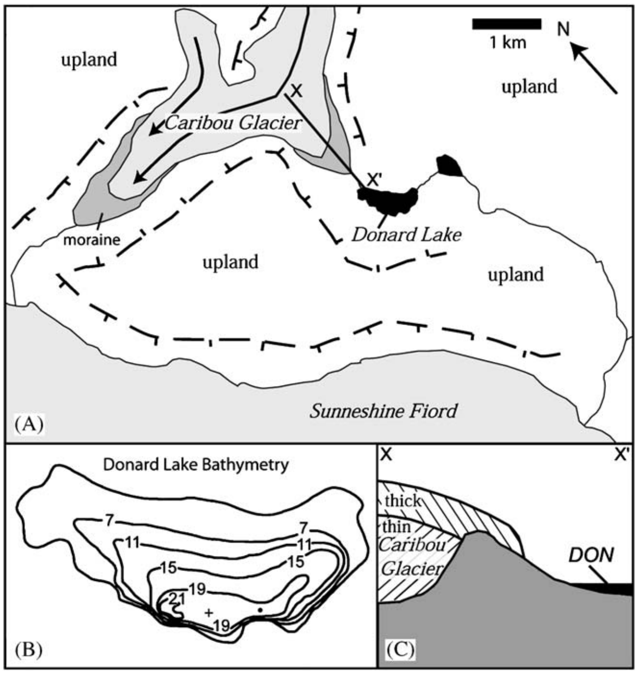

Fig. 4. (A) Map view showing the relation of Donard Lake to the Caribou Glacier with the current ice limit (light stipple) and its Little Ice Age moraines (darker stipple). (B) Bathymetric map of Donard Lake; contours are in meters. Cores 95-DON-03 (+) and 94-DON-01 (●) are located in the central deep. (C) Cross section from X to X′ in panel A, illustrating how as the Caribou Glacier thickens it breaches the col separating Donard Lake from the Caribou Glacier, and in so doing diverts surface meltwater into Donard Lake.

The most complete core, 95-DON-03 from the western half of the central basin (Fig. 4), was recovered in two separate drives of an 11-cm-diameter core barrel in 19.75 m water depth. 95-DON-03A recovered a continuous section from the sediment–water interface to 250 cm depth. 95-DON-03B recovered sediment from 175 to 390 cm depth, providing more than 50 cm overlap between cores. The two cores were spliced together by diagnostic volume MS signatures made in the field (Fig. 5A–C). After settling and dewatering, the total sediment length was 356 cm. Five AMS ^{14}C dates were obtained from 95-DON-03; two on plant macrofossils, and three on humic acid extracts (Table 1). A second core, 94-DON-01, was used to help constrain the geochronology of the basin, because it contained aquatic moss macrofossils in several layers. The dates obtained from 94-DON-01 (Table 1) were mapped into 95-DON-03 using diagnostic volume MS signatures (Fig. 5D,E).

A polynomial fit to the dates is used to define an age model (Fig. 2). Magnetic parameters (volume and mass MS and standard paleomagnetic factors) were measured in discrete 1-cm^3 samples in the laboratory on

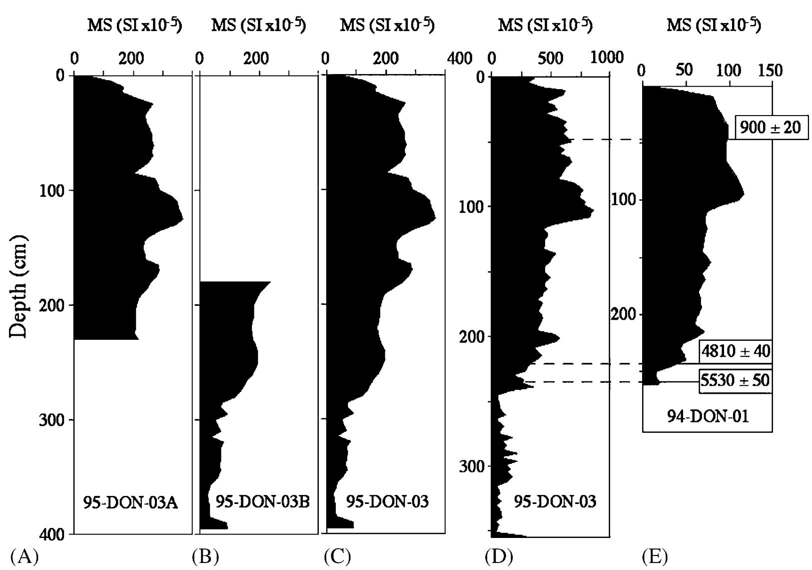

Fig. 5. Magnetic susceptibility (MS) used as a correlation tool to splice cores together and for inter-core correlation. (A) Field measurements of volume MS in 95-DON-03A. (B) Field measurements of volume MS in 95-DON-03B. (C) MS of the two core segments combined to form a composite curve for the full sediment fill. (D) Paleomagnetic analyses in the laboratory provide a more detailed volume MS record on discrete 1-cm^3 samples, that correlates well with the field measurements. Dewatering during transit resulted in a shortening of the total sediment thickness from 395 to 355 cm. (E) Dates obtained on core 94-DON-01 can be mapped into corresponding levels in core 95-DON-03 (dashed lines) based on diagnostic changes in the MS of each core. Correlations based on thesis research of Moore (1996).

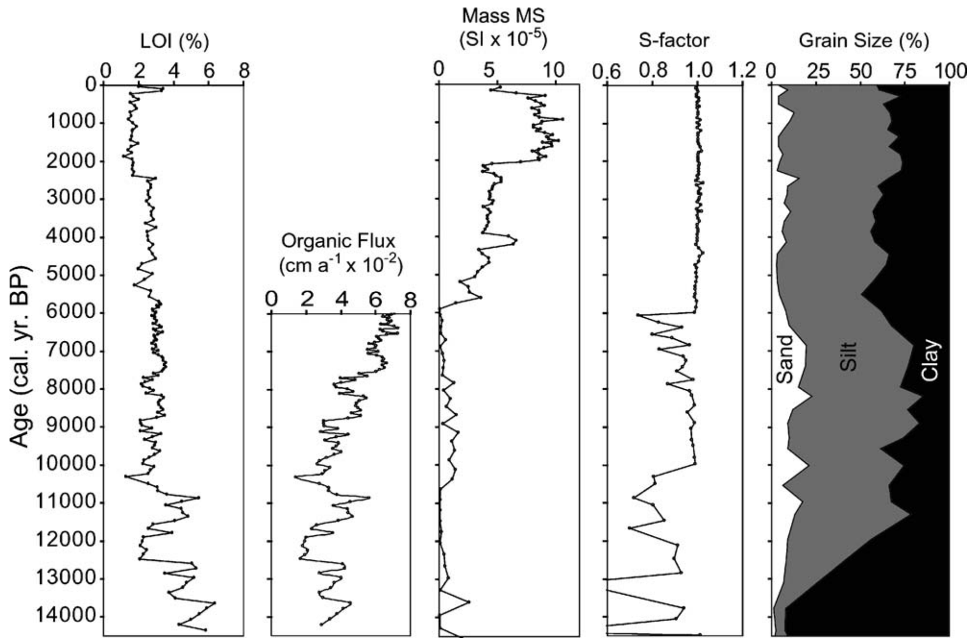

Fig. 6. Down-core changes in parameters measured in core 95-DON-03. Loss-on-ignition (LOI) is a proxy for in-lake biological productivity. Organic flux (LOI × sedimentation rate) provides a more quantitative measure of lake productivity. Mass magnetic susceptibility was measured on 1 cm^3 subsamples, as was the paleomagnetic *S*-factor, the ratio of IRM to SIRM, and a proxy for grain freshness. Complete grainsize analysis of 45 samples is available at http://www.ncdc.noaa.gov/paleo/data.html; only the % sand, silt, clay is presented here.

95-DON-03; mass MS and *S*-factor are plotted in Fig. 6, volume MS on Fig. 5D. *S*-factor, the ratio of isothermal remanent magnetization to saturation isothermal remanent magnetization (IRM/SIRM), is a proxy for the proportion of magnetite to hematite grains, which in this environment can be considered an index for the degree of weathering of the sediment. Higher values indicate fresher sediment. Grain size was determined on 48 1-cm-thick samples, spaced 10 cm apart using a Sedigraph; sand, silt, and clay are shown in Fig. 6. Pollen was counted in 30 samples spaced 20 cm apart in the upper half of the core and 10 cm apart in the lower half, to compensate for lower sedimentation rate in the latter; select taxa are shown in Fig. 7; full pollen counts are available from www.ncdc.noaa.gov/paleo/data.html. LOI was measured in 174 samples, as a continuous series at 1 cm intervals from 235 to 349 cm depth, and every 4 cm from 0 to 232 cm depth, spaced to provide one sample about every 100 years. Each 1-cm-thick sample represents 20–100 years, depending on sedimentation rate. LOI is primarily a function of the rate of primary productivity within the lake and its dilution by clastic sediment supplied by the catchment. To reduce the influence of changes in clastic sedimentation rate on the LOI curve, we also calculate the flux of organic material (LOI × sedimentation rate) for the interval from 6 to 14.5 ka (Fig. 6). At younger levels, the change in sediment source obscures this relationship. Plotted as a flux, LOI gives a more quantitative representation of the within-lake biotic activity.

The sedimentology of the deepest levels reflects unusual environmental conditions. Clay, almost exclusively $<1\,\mu m$, constitutes $>80\%$ of the sediment between 14.5 and 13 ka, gradually decreasing to a background level of ca. 20% by 12 ka, with only minor variability thereafter. No other Baffin Island lake we have studied ($n = \sim 30$) has such a dominance of clay at any level. The source of this clay is unlikely to be from glacial meltwater, as glacial erosion products in the crystalline terrain of Baffin Island are dominantly in the silt grades. The clay is also not wind transported from a distant source. X-ray diffractometer studies show that the mineralogy of the clay fraction at the base of the core is indistinguishable from clay-sized sediment in the glacier-dominated portion of the core (the upper 2 m), suggesting that clay-sized sediment in both levels is derived from abrasion of local bedrock. Consequently, we interpret the unusually high percentages of fine clay to indicate that during this interval the lake remained mostly ice-covered throughout the year with only a narrow moat opening in summer. Resuspension of sediment by wave action would be minimal, and runoff during cold summers would be less than at present, resulting in the mobilization and transportation of only the finest particles from the lake margin to the central deep.

3.2. Dye lower water lake

A small lake used as the primary water supply for the Dye Main Lower Camp of the DEW Line (now North Warning System), and informally named by us as Dye Lower Water Lake, (DLW; 66° 38′N; 61° 25′W; 306 m asl) is dammed between crests of the Sunneshine Moraines, a lateral moraine complex above the break

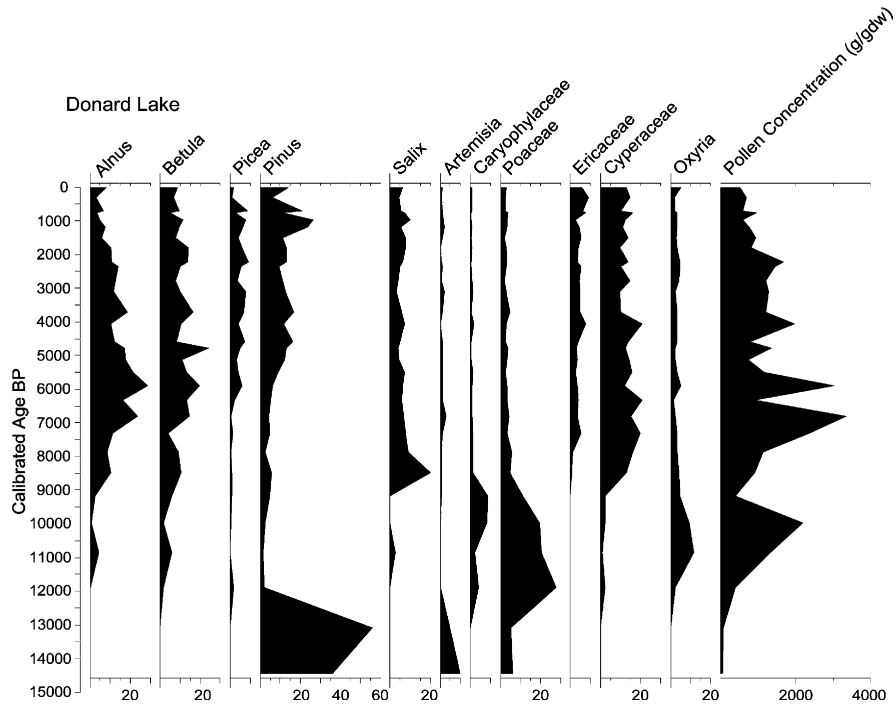

Fig. 7. Pollen percentage diagram of key plant taxa, and total pollen concentration (pollen grains per gram dry weight) from Donard Lake core 95-DON-03 plotted against calibrated age. Pollen data are given as percentages of the total pollen sum.

in slope of Sunneshine Fiord, easternmost Cumberland Peninsula (Fig. 3). CE dating of moraine boulders confirms that the moraines are from the last glaciation (Miller et al., 2002). DLW is roughly circular, about 200 m in diameter, and has a simple bathymetry with a central deep reaching almost 7 m water depth. The lake has no significant inflow stream. Five cores recovered in 1993 from the center of the basin were between 150 and 170 cm in length; the longest core, 93-DLW-03, was selected for radiocarbon, pollen and MS analyses. MS is only marginally above background (Fig. 8), reflecting the lack of a terrigenous sediment supply. The lack of glacial-lacustrine sediment at the base of any of the cores suggests that the lake formed as the moraines stabilized after glacier recession.

The Sunneshine Moraines contain small amounts of carbonate in the form of marine shells and rare limestone clasts dredged from the floor of Sunneshine Fiord by the advancing outlet glacier. To test for hardwater from these carbonates, we dated living aquatic moss from the lake, and DIC precipitated as $BaCO_3$ on site. Both showed small, but measurable hardwater influences of 100–250 years (Table 1). This introduces a small uncertainty in the age model. Four AMS ^{14}C dates were obtained from 93-DLW-03, two on aquatic moss macrofossils, and both a humic acid and humin extract from near the base of the core. As observed elsewhere on Baffin Island lakes, the humin fraction is older than the humic acid fraction; the latter was used for the age model (Fig. 2). Because there is little minerogenic sediment input to the lake, the uppermost lacustrine sediments are poorly consolidated and could not be recovered well with the piston corer. To capture the surface sediment, a 12-cm-long box core was recovered close to the piston core locality. Because MS exhibits little variability in DLW sediments, the box core was spliced into the piston core based on the diagnostic latest Holocene *Ambrosia* rise derived from pollen analyses. This suggested that about 6–8 cm of sediment was lost from the piston core. Pollen was counted from six 1-cm core slices in the box core (every second centimeter), and from 40 levels in the piston core at 2.5 cm intervals over the first 20 cm, then at 5 cm intervals for the remainder of the core. Full pollen counts are available at www.ncdc.noaa.gov/paleo/data.html; counts for key taxa are shown in Fig. 8.

3.3. Lake Jake

Lake Jake (unofficial name; 63° 40′N; 65° 10′W; 300 m asl) is a glacier-dominated lacustrine system situated 2.3 km downstream from a major outlet glacier of the 93 km^2 Cornelius Grinnell Ice Cap, Hall Peninsula (Fig. 9). The lake is 2 km long, covering 1.1 km^2, with a catchment of 23 km^2, 30% of which is glacierized. Lake Jake consists of two main basins; the larger eastern basin reaching 40 m water depth is separated from a shallower (25 m water depth) western basin by a sill only 8 m below the surface (Fig. 9). Two sediment cores collected in 1991 from the eastern basin

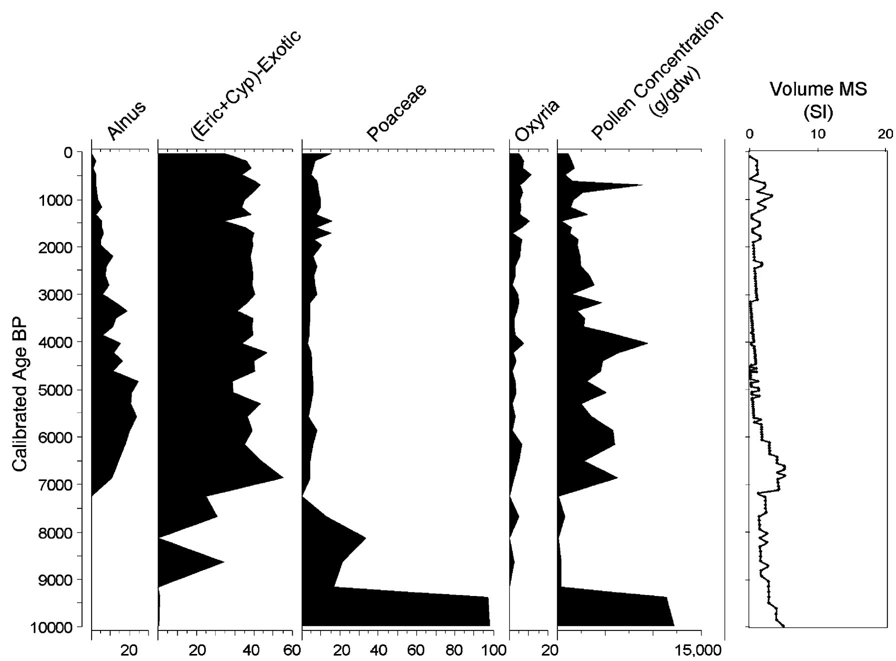

Fig. 8. Pollen percentage diagram of key plant taxa, and total pollen concentration (pollen grains per gram dry weight) from Dye Lower Water Lake core 93-DLW-03 plotted against calibrated age. Pollen data are given as percentages of the total pollen sum. Volume MS is consistently low, supporting the lack of an inflow stream and little clastic input to the lake.

(91-LJ1, 91-LJ2) and one from the western basin (91-LJ3) are presented here. All three cores terminated on bedrock or on stones in diamict. X-radiography of the cores reveals the distribution of ice-rafted detritus (IRD), which was further quantified by sieving core 91-LJ3.

3.3.1. Proximal basin

Core 91-LJ3 was taken in 23.1 m water depth, 100 m from the inlet to the lake in the western basin. One hundred and ninty-five centimeters of sediment were recovered, of which the uppermost 115 cm are lacustrine muds, underlain by 80 cm of high MS stony diamict. The diamict is massive, and contains abundant pebbles visible in X-radiographs and quantitatively as high concentrations of sediment grains coarser than 2 mm. Terrestrial plant fragments are also present in the diamict. We interpret the diamict to represent deposition beneath or at a readvancing glacier margin. Inwashed terrestrial macrofossils, although rare, are present through much of the lacustrine portion of the core. Plant macrofossils from three levels in the lacustrine muds were selected for radiocarbon dating (Table 1), and used to construct an age model (Fig. 2). IRD was quantified by sieving the whole core in 10 cm increments. Sediment between 2 and 4 mm diameter, and clasts coarser than 4 mm were separated and quantified as the number of clasts per 300 cm^3 of bulk sediment (Fig. 10).

3.3.2. Distal basin

Core 91-LJ1, recovered from 37.4 m water depth, contains 64 cm of lacustrine sediment. Core 91-LJ2, located 150 m west of 91-LJ1 in 38.3 m water depth contains 110 cm of sediment, of which the upper 70 cm is lacustrine mud, and the lower 40 cm is a stony diamict. MS records show nearly identical trends in the lacustrine portions of both cores, with low MS through must of the lacustrine section except for an interval of relatively high MS between 50 and 60 cm depth in both cores (Fig. 9). IRD is qualitatively represented in X-radiography of the two cores. In both cores, the upper 25 cm contains scattered ice-rafted clasts, but IRD is generally lacking in deeper levels of the lacustrine sediment in both cores except around 55 cm depth. Both cores are devoid of plant macrofossils.

Core 91-LJ1 was selected for pollen analysis and AMS ^{14}C dating. Humic acids extracted from 1 cm core slices at seven levels were dated (Table 1), and used to derive an age model (Fig. 2). Pollen was counted at 38 levels between 1 and 65 cm depth, using 1 cm slices of core (ca. 50–100 years of record for each sample). At least 300 grains were counted at each level, except for three levels with very little pollen between 45 and 50 cm depth; diagnostic taxa are presented in Fig. 11; full counts are available at www.ncdc.noaa.gov/paleo/data.html.

Cores 91-LJ1 and 91-LJ2 document deglaciation of the basin ca. 9.5 ka, at the end of the Cockburn Substage

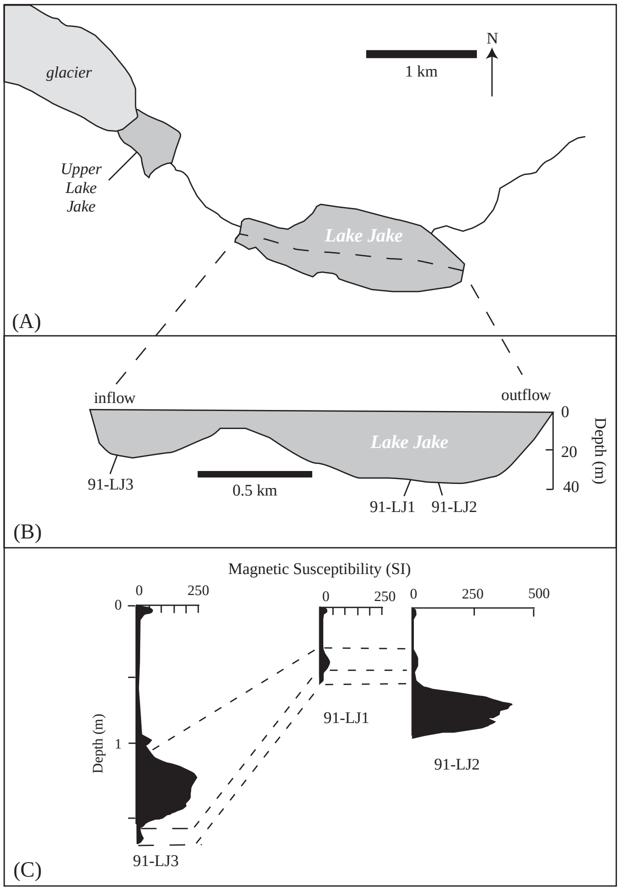

Fig. 9. (A) Plan view of Lake Jake, showing an outlet glacier from the Cornelius Grinnell Ice Cap terminating at an upper lake that catches most of the coarse sediment transported by meltwater streams, and Lake Jake farther downstream. (B) A cross-section through the central axis of Lake Jake, showing the position of cores 91-LJ1, 91-LJ2 in the distal basin, and 91-LJ3 in the proximal basin. (C) Whole-core volume magnetic susceptibility of the three cores scaled to depth. Radiocarbon dating reveals that the high MS zone (diamict) in 91-LJ3 was deposited by a glacier readvance into the proximal basin, and that this correlates with the small rise in MS near the base of core 91-LJ1, not the high MS zone (diamict) at a deeper level seen in 91-LJ2, which was deposited during regional deglaciation more than a millennium earlier.

(Andrews and Ives, 1978). Following a millennium of normal lacustrine sedimentation, a glacier readvanced into the proximal basin, depositing the diamict at the base of core 91-LJ3. Terrestrial plant remains incorporated within the diamict confirm that the diamict was deposited during a readvance, rather than during regional deglaciation. The age of the diamict is constrained by AMS ^{14}C dates on plant remains from two levels within the diamict, and additional plant remains in lacustrine sediment directly overlying the diamict. All three ages are statistically indistinguishable ($\pm 2\sigma$) at about 8.2 ka, indicating a glacier readvanced into the proximal basin at this time. There is no diamict of this age in the distal basin, but sediment of this age (45–55 cm) in core 91-LJ1 has IRD visible in X-

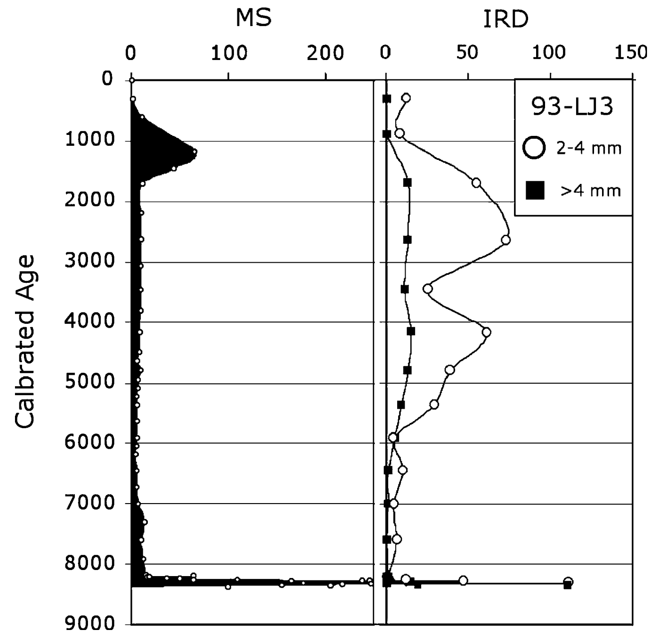

Fig. 10. Whole core volume magnetic susceptibility (MS) and ice-rafted detritus (IRD) from Lake Jake core 93-LJ3 plotted against calibrated age. IRD is the number of grains between 2 and 4 mm (open circles) and >4 mm (solid squares) sieved from 300 cm^3 of core.

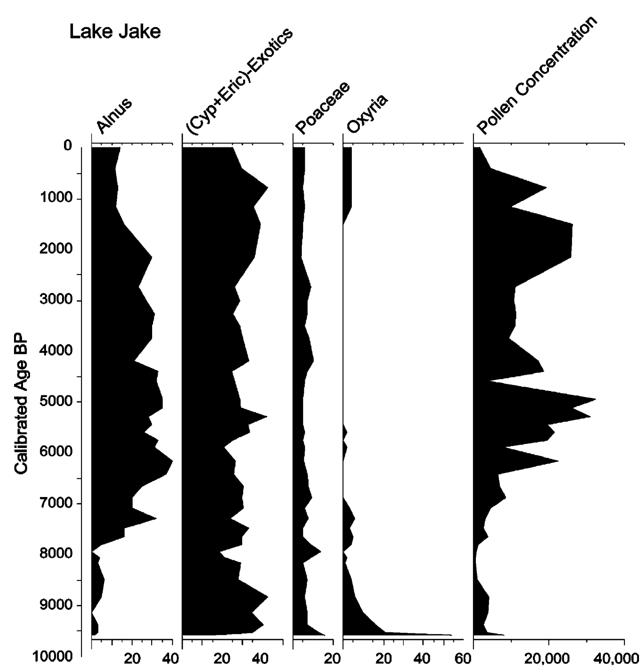

Fig. 11. Pollen percentage diagram of select plant taxa, and total pollen concentration (pollen grains per gram dry weight) from Lake Jake core 93-LJ1 plotted against calibrated age. Second column is the sum of Cyperaceae (Cyp) and Ericaceae (Eric), two of the dominant modern tundra taxa, after removing all exotic taxa from the total pollen sum.

radiographs, anomalously low pollen concentrations (implying an elevated sedimentation rate), and high MS (greater proportion of clastic sediment). Similar MS and

IRD occur at comparable depths in core 91-LJ2. We interpret sediment at these levels to be delivered to the distal basin by sediment-laden meltwater from the glacier that occupied the proximal basin.

3.4. Clyde Foreland

The Clyde Foreland (Fig. 1) was deglaciated ca.14 ka. The most detailed record of Holocene environmental change is from a 1.7-m-long sediment core from lake CF3 (70° 32′N, 68° 22′W; 25 m asl), a small ($\sim$1 km^2), isolated basin above the postglacial marine limit, lacking significant fluvial inflow; lake depth at the core site is 6.1 m. The basal few centimeters of the core are fine sand, overlain by 160 cm of lacustrine sediment, ranging from a macrofossil-rich gyttja at the base, into a light-brown gyttja, that grades into a silty gyttja toward the top of the core. Aquatic moss macrofossils occur throughout the core, and were selected for radiocarbon dating from six levels (Table 1). MS, measured both in the field (volume) and in the lab (mass), grain-size, and LOI (every centimeter) provide the primary paleoenvironmental proxies.

The basal radiocarbon age, 11.2 ka (Table 1), is similar to ($\pm$0.5 kyr) basal ages in five other lakes from the foreland (Briner, unpublished data), suggesting that lake basins only began to accumulate sediment at the end of the Younger Dryas chron. We presume that this discrepancy indicates the lakes remained permanently frozen following deglaciation until ca. 11.5 ka.

4. Climate implications of the lacustrine records

Donard Lake provides the most complete record of the Holocene on Baffin Island. The lake began to accumulate sediment early in the deglacial cycle, and captures a longer record of terrestrial environmental change than any other site. We use the Donard Lake core, supplemented by cores from the other lakes to derive an interpretation of environmental change since the last deglaciation.

4.1. The early postglacial: 14.3–9 ka

Sedimentation in Donard Lake began just prior to 14 ka, and is dominated by clay. Most of the rare pollen grains preserved in this sediment are exotic, derived from plants living south or west of the LIS (mostly *Pinus*, but also rare *Juglans*; Fig. 7). These two proxies suggest that summers were colder than present and local vegetation was either absent or reproducing vegetatively. Despite apparently cold, brief summers, primary productivity was occurring in the lake. Aquatic moss macrofossils, rare in higher levels of the core, are present in these deepest levels, confirming aquatic biotic activity even though terrestrial plants were virtually absent. The concentration of organic matter was moderately high from 14.3 to ca. 12.5 ka when it dropped dramatically until ca. 11.5 ka, after which it recovered to values higher than before 12.5 ka. This interval of low LOI is broadly coincident with the Younger Dryas chron. The chronological control lacks sufficient precision to make a definitive correlation, but dates below (13.7 ka) and above (10.1 ka) the LOI minimum provide reasonable certainty for the correlation. Reduction in primary productivity, whether measured as LOI or as the flux or organic matter (Fig. 6), suggests colder summers. The lack of evidence for glacier readvances at that time, and the conclusion that Caribou Glacier remained thinner than it is at present throughout the Younger Dryas chron, suggests that conditions may have also been drier. A drier, colder climate is supported by the dramatic decrease in snow accumulation rates and depressed mean annual temperatures over Greenland during the Younger Dryas chron (Cuffey and Clow, 1997).

Pollen sampling was not at sufficient resolution to evaluate vegetation changes during the Younger Dryas interval, but the pollen spectrum is dominated by grass from 12.5 until at least 10.5 ka. The relatively high levels of *Oxyria* (sorrel), an initial colonizer of disturbed, or freshly deglaciated landscapes, suggests that local hillslopes remained unstable until 10 ka. The Donard Lake pollen results are supported by pollen from basal sediment in nearby Dye Lower Water Lake (9.5 ka), which is also dominated by grass and *Oxyria*, with a relatively high influx of pollen to the lake (Fig. 11). The pollen assemblages prior to 8 ka have no modern analogs, so cannot be quantitatively interpreted in terms of summer temperatures, although qualitatively, modern grass-dominated arctic sites are cold and dry (Gajewski and Atkinson, 2003; Kerwin et al., 2004).

Readvances of both local mountain glaciers and the LIS are recorded throughout most of Baffin Island during the Cockburn Substage, between 10 and 9 ka, but proxy evidence for this event is not obvious in ice cores from Greenland or Arctic Canada. A possible correlative change is present in the Donard Lake core. LOI recovers to its pre-Younger Dryas levels in 95-DON-03 by ca. 11.5 ka, then decreases again to intermediate levels after 10.5 ka, and remains at a similarly low level until just after 9 ka, when it rises again. Pollen influx in the Donard Lake core is relatively low through the same interval.

4.2. Establishment of modern tundra vegetation and the 8.2 ka event

Between 10 and 8 ka, pollen assemblages in core 95-DON-03 indicate regional vegetation changed from a grass/Caryophyllaceae-dominated tundra to the modern sedge/heath tundra, which became fully established by 8 ka, with only modest variability subsequently. A similar shift from grass-dominated tundra to the modern sedge/heath tundra occurs in the Dye Lower Water Lake core, although about 1000 years later, according to the chronology. However, the same shift occurs just before 9 ka at the more southerly Lake Jake site (Fig. 8). Other sites from Baffin Island show that the modern sedge-heath tundra vegetation was established by 8 ka (Kerwin et al., 2004).

A glacier readvance centered on 8.2 ka is well constrained at Lake Jake by dated terrestrial plant remains within diamict and additional plant debris of the same age in lacustrine sediment immediately overlying the diamict. This is the first documentation of a glacier readvance on Baffin Island at this time. Support for cooler summers at about the same time comes from Donard Lake. LOI in 95-DON-03 exhibits only modest variability between 9 and 2 ka, except for two brief decreases. The best defined of these is a 35% drop in LOI lasting several centuries about 8 ka. The age model is only constrained by ages at ca. 7 and 10 ka, hence a precise date on this brief, but well-defined decrease in organic content is illusive, but the age is certainly close to 8 ka and it is a reasonable candidate for the 8.2 ka event. Decreased LOI during an interval without substantive changes in either MS or grainsize (Fig. 6) suggests that reduced organic matter reflects reduced primary productivity, most plausibly caused by lower summer temperatures. This interpretation is supported by an increase in grass pollen at this time in cores from both Lake Jake (Fig. 11) and Dye Lower Water Lake (Fig. 8).

4.3. Neoglaciation

The threshold setting of Donard Lake provides a clear signal for the expansion of the adjacent Caribou Glacier. In core 95-DON-03, the early Holocene sedimentation rate is relatively low, as is MS, and the sediment contains variable amounts of weathered magnetic mineral grains (*S*-factor; Fig. 6). Between 6 and 5.5 ka, MS underwent a dramatic, step-function increase and magnetic minerals became exclusively fresh. At the same time, the sediment character shifted from weakly stratified to finely laminated. These shifts signal the onset of a glacial-lacustrine-dominated environment at Donard Lake, indicating that Caribou Glacier thickened sufficiently to breach the col separating it from the valley of Donard Lake, greatly increasing the catchment size and delivery of suspended sediment to the lake. We interpret these changes to reflect the onset of Neoglaciation. Once the col was breached, it does not appear that Caribou Glacier ever thinned sufficiently to drop below the col, as the MS and *S*-factor never drop to levels similar to those before 6 ka. A second step-function change occurred ca. 2 ka, with a doubling of the MS and a nearly 50% decrease in LOI. We interpret this change to reflect intensification of glacial erosion and an even greater dominance of glacially eroded sediment to the lake. The uppermost few centimeters of the core have lower MS and higher LOI, presumably reflecting 20th century warming, glacier recession and less effective diversion of meltwater to the lake than before.

The IRD record in sediment cores from Lake Jake supports the onset of Neoglaciation derived from Donard Lake. IRD is absent in core 91-LJ1 X-radiographs from ca. 8 ka until 5.6 ka (23 cm, and at the same depth in LJ2), after which IRD is present continuously until near the tops of both cores. In core 91-LJ3, IRD is common in the basal diamict, but virtually absent in overlying sediment until it reappears at 80 cm depth (6 ka) after which IRD is present continuously until near the top of the core. IRD (2–4 mm diameter grains) rises from a background level of 5–10 grains/300 cm^3 to >30 grains/300 cm^3 at 5.6 ka. There is also a modest, but significant increase in the MS of LJ3 by 5 ka, suggesting a higher minerogenic flux to the lake. There is no obvious change in the pollen record to suggest that vegetation shifted in response to the onset of Neoglaciation.

5. Baffin Island since the Last Glacial Maximum

5.1. Deglaciation and deglacial readvances

The limits of Laurentide ice at the LGM and the timing of its recession have been intensely debated in the literature since the first radiocarbon dates appeared in the early 1960s. The debate was between those who argued that the LIS terminated at the continental shelf break, and those who postulated an LIS margin restricted to the fiord heads (summarized by Miller et al. (2002)). Over the past decade, compelling data derived from detailed field mapping and the application of CE dating to constrain the ages of moraines and glaciated surfaces for the most part have resolved this dispute (Steig et al., 1998; Kaplan, 1999; Marsella et al., 2000; Dyke and Hooper, 2001; Miller et al., 2002; Briner et al., 2003; Briner et al., 2005). The Foxe Dome of the LIS dominated Baffin Island at the LGM. Low-gradient outlet glaciers and ice streams occupied fiords and sounds along eastern Baffin Island, terminating on the continental shelf landward of the shelf break. Many inter-fiord uplands remained above the limit of Laurentide ice throughout the LGM; some of these areas may have been covered by locally derived, non-erosive, cold-based ice, while others may have remained unglaciated (Miller et al., 2002). Deglaciation began by 15 ka, with the coastal fringe ice-free about 14 ka (Fig. 12; Briner

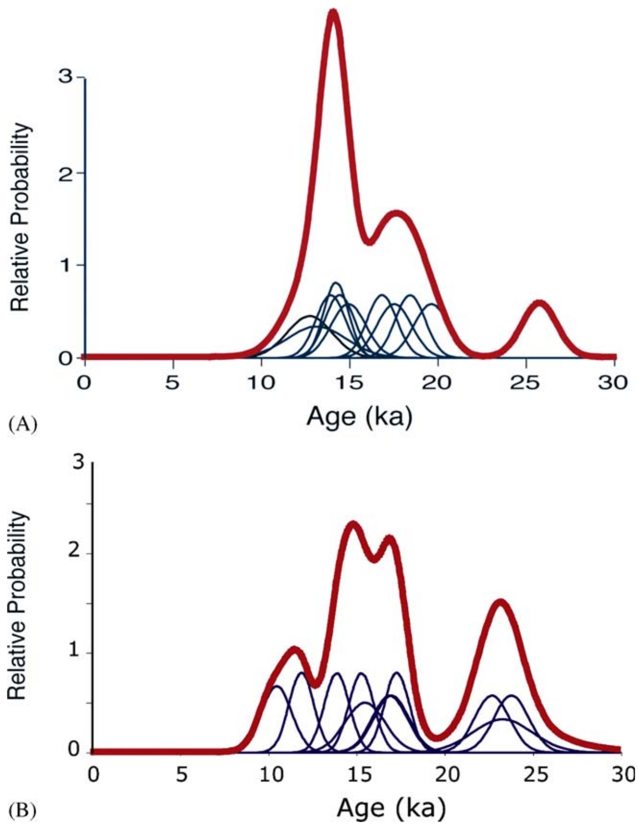

Fig. 12. Cosmogenic exposure (CE) dates on large (>2 m high) glacially derived boulders and glacial-fluvially derived quartz cobbles that define the timing of deglaciation from the Cape Aston Lowland ($n = 13$, panel A) and the coastal portion of the central Clyde Foreland ($n = 15$; panel B). Ages are expressed as stacked probability distribution functions (heavy line) by summing the probability distribution functions of the individual samples (Table 2). Some inheritance is apparent in both datasets, with a clear mode at ca. 14 ka for the Aston Lowland, and a broader dominant mode on the Clyde Foreland between 14 and 18 ka, with a secondary mode at ca. 23 ka that either represents inheritance or reflects erratics deposited during an earlier deglacial event preserved beneath cold-based LGM ice. Two much older CE dates from Clyde Foreland and one from Aston Lowland that reflect substantial inheritance are listed in Table 2, but not shown here. Modified from Briner et al. (2005) and Davis et al. (in press).

et al., 2005). Most fiords were occupied by Laurentide outlet glaciers until ca. 11 ka. Shortly after 11 ka, outlet glaciers receded rapidly to the fiord heads, in some instances in less than 1000 years, with modest still stands at physiographically controlled pinning points within the fiords that are probably unrelated to regional climate. In Clyde Inlet, for example, CE ages on erratics and ice-sculpted bedrock indicate deglaciation of the outer fiord about 10.5 ka, the middle sector between 9.6 and 9.4 ka, with the terminus at the head of Clyde Inlet by 9 ka (Briner et al., 2003).

This pattern of rapid ice sheet retreat was interrupted throughout Baffin Island by a readvance of both Laurentide and local glaciers forming the Cockburn Moraines (Falconer et al., 1965; Miller and Dyke, 1974; Andrews and Ives, 1978). The readvance is defined by till overlying marine sediments dated between 9.5 and 9.9 ka, and overlain by marine sediment dated between 8.9 and 9.4 ka. Because both Laurentide and local glaciers readvanced, the readvance is most likely a response to regional climate forcing, although a compelling mechanism for climate change at this time has not been identified. The lack of a significant negative excursion in $\delta^{18}O$ of Greenland ice cores (Grootes et al., 1993) indicative of substantial cooling during the Cockburn Substage suggests that the readvance was not triggered by temperature. A more likely scenario is that the Cockburn readvance reflects a response to increased precipitation. Snow accumulation rates derived from Summit ice cores almost tripled from Pleistocene levels between 11 and 10 ka (Cuffey and Clow, 1997). At this time, the residual LIS may have had sufficient size to maintain low summer temperatures along its downwind eastern margin, despite strong insolation forcing, resulting in a positive net mass balance. The Cockburn Moraines represent a widespread readvance during the warmest phase of the present interglacial. A decrease in LOI and reduced influx of pollen in Donard Lake sediment between 10.5 and 9 ka may be a correlative response in lacustrine systems.

The response of Baffin Island glaciers to the 8.2 ka event (Alley et al., 1997; Barber et al., 1999) has not been previously documented. We present new evidence for a glacier readvance ca. 8.2 ka at Lake Jake, coupled with a modest vegetation response suggesting landscape instability consistent with colder summers. A significant reduction in lake primary productivity at Donard Lake occurred at about the same time. Collectively, these records are consistent with a brief climate shift to colder summers, sufficient to impact vegetation, reduce aquatic primary productivity in lakes, destabilize hillslopes, and produce positive glacier mass balances of sufficient duration to lead to glacier readvances. Other than its clear representation in ice cores, the only other site in the Canadian Arctic to capture the 8.2 ka event is a high-resolution pollen record from lake sediment in the central Arctic reported by Seppä et al. (2003), where a dramatic decline in *Betula* pollen occurs about 8.2 ka, and is followed by a brief peak in grass pollen, consistent with our reconstructions of Baffin Island conditions.

5.2. Timing of peak Holocene warmth

Lake sediment records suggest peak primary productivity, which we equate with peak summer temperatures, occurred between ca. 10.2 and 8.5 ka, in line with the melt record from Agassiz Ice Cap (Fisher and Koerner, 2003) and both records broadly follow summer insolation at 70°N (Fig. 13). LOI values in Clyde Foreland

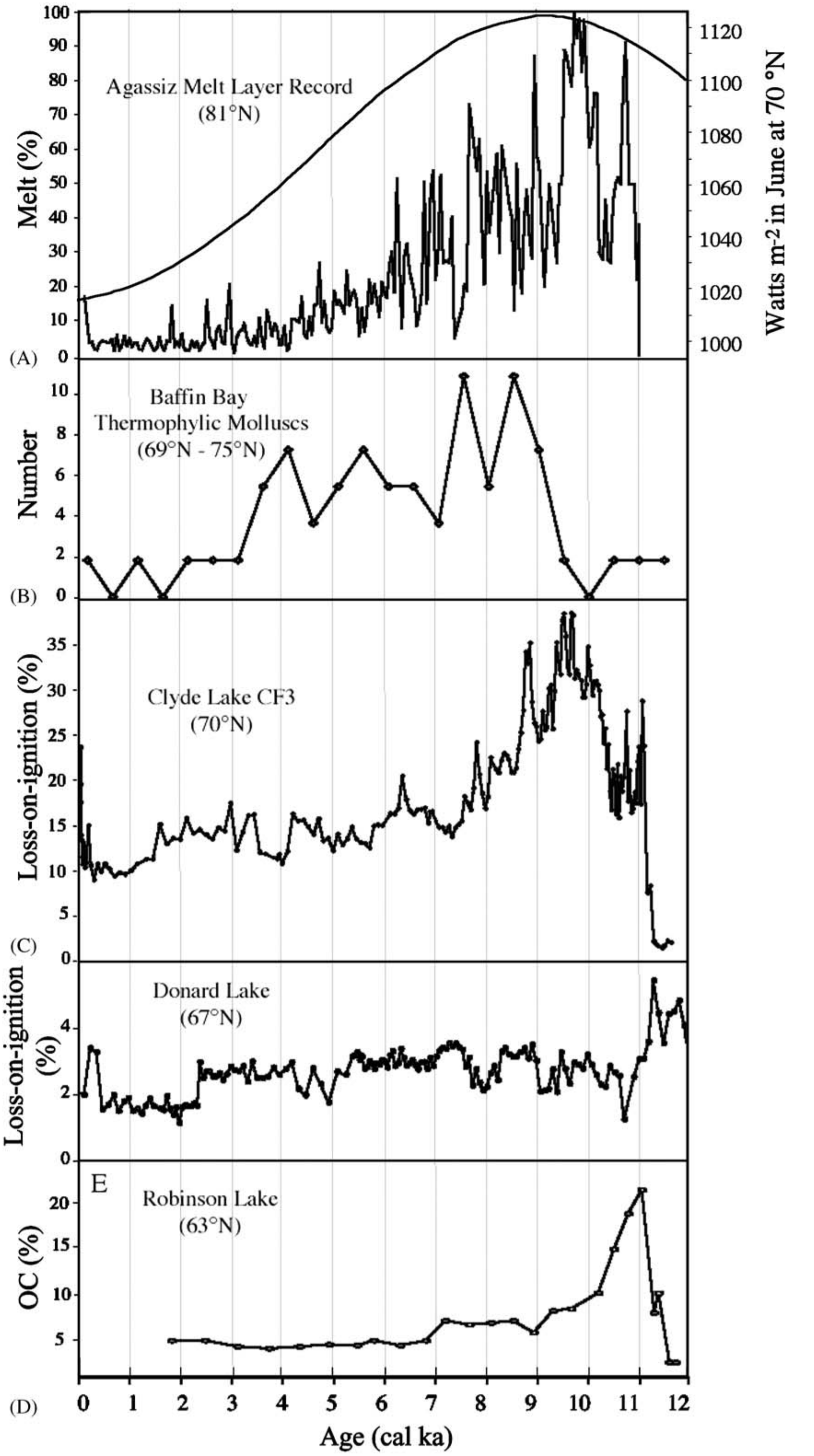

Fig. 13. North–south transect of continuous records of summer temperature proxies from the eastern Canadian Arctic. (A) Summer melt record from Agassiz Ice Cap (Fisher and Koerner, 2003) and June insolation at 70°N (Berger, 1988). (B) Changes in the number of thermophilous mollusk taxa present in Baffin Bay (Dyke et al., 1996). (C) LOI for 02-CF-03 and 95-DON-03, and (D) LOI for Robinson Lake (Miller et al., 1999), southern Baffin Is. Lakes on southern Baffin Island reach maximum summer temperature indicators about 1000 years earlier than at higher latitudes, and peak sea-surface temperatures lag another 1000 years behind the northern terrestrial sites.

lake CF3 rise from near zero to 20% between 12 and 11 ka. At ca. 10.2 ka, LOI rises to ~30%, peaking at 38% at 9.5 ka. After ca. 8.5 ka, LOI declines, with considerable sub-millennial structure, into the late Holocene, reaching minimum Holocene values during the last millennium, before rising sharply during the 20th century. If expressed instead as the flux of organic matter (not shown) these trends are exaggerated. Similar down-core trends in LOI are apparent in five other lakes from the Clyde Foreland (not shown). Although the chronologies for these lakes are not complete, they have similar basal ages to lake CF3 (Briner, unpublished data).

Glacier recession and establishment of the modern tundra vegetation both appear to lag peak terrestrial summer warmth. Many glaciers were not behind their modern margins until 7 ka. Baffin Island pollen data only have reliable modern analogs after 8 ka, precluding direct comparisons with changes in lacustrine productivity. However, the pollen evidence suggests summers warmed after 8 ka, reaching peak warmth after 7.5 ka, which persisted until at least 5 ka (Kaufman et al., 2004; Kerwin et al., 2004). The discrepancy between peak biological productivity in lakes and peak summer warmth reconstructed from pollen assemblages suggests that the establishment of modern vegetation may have been delayed until pedogenesis provided a suitable substrate for vegetation to equilibrate with climate.

Marine sea surface temperatures (SST) also peak later than terrestrial temperatures. Whalebones from raised beaches in the eastern Canadian Arctic north of Baffin Island first appear between 12 and 11 ka, indicating summer sea ice cleared from Baffin Bay by that time (Dyke et al., 1996). But peak SSTs are best defined by diagnostic thermophilous mollusk taxa that first immigrate into northeastern Baffin Bay ca. 9.2 ka. Their distribution peaks between 8.5 and 7.5 ka, and only diminishes to contemporary limits after 3.5 ka (Dyke et al., 1996). The delay in peak oceanic warmth suggests that for Baffin Island, summer insolation dominates over oceanic heat transport in determining terrestrial warmth during the Holocene.

5.3. Late Holocene cooling and the onset of Neoglaciation

Neoglaciation is apparent throughout Baffin Island in the form of fresh moraines beyond modern glacier termini. Most of these moraines are from the Little Ice Age, but moraines as old as 3–4 ka have been identified where they were protected from subsequent overriding (Miller, 1973; Davis, 1985). The onset of Neoglaciation is better defined from lacustrine archives. The most reliable indicator is the onset of finely laminated, exceptionally fresh minerogenic sediment at Donard Lake beginning ca. 6 ka. Similarly, Lake Jake sediment shows a significant and sustained increase in IRD shortly after 6 ka. Supporting evidence comes from lake sediment incorporated in a Neoglacial moraine bordering the Clyde Foreland radiocarbon dated to ca. 3.3 ka (Table 1), a minimum age for Neoglaciation. Intensification of Neoglaciation after about 2.5 ka is apparent in many records from the Arctic, including Donard Lake, where lake productivity decreased and the input of glaciogenic sediment increased.

These reconstructions are supported by $\delta^{18}O$ in cellulose extracted from aquatic moss in lake sediment cores that suggest abrupt cooling in the Cumberland Sound area at about 6 ka and intensified cooling ca. 2.5 ka (Sauer et al., 2001). Baffin Island pollen-based late Holocene summer temperature reconstructions suggest late Holocene cooling set in between 5 and 3 ka (Kerwin et al., 2004), somewhat later than the response of glacial systems. Pollen data suggest intensification of cooling after 2 ka and an overall late Holocene cooling of 1–2 °C (Kerwin et al., 2004). The diatom record from Fog Lake suggests similar late Holocene lake-water cooling beginning ca. 2.5 ka (Wolfe, 2002).

5.4. The Little Ice Age

Glaciers throughout the Canadian Arctic show clear evidence of Little Ice Age expansion, persisting until the late 1800s, followed by variable recession over the past century. Some glaciers remain at their Little Ice Age maximum, although most have receded significantly. In a few instances, the relative magnitude of the Little Ice Age advance can be compared to earlier advances. In all such cases, the Little Ice Age is the most extensive Late Holocene advance. Lichenometeric evidence suggests a two-part Little Ice Age extending over 800 years (Davis, 1985).

Late Holocene cooling is stronger in the Atlantic sector of the Arctic than in the Pacific sector (Kaufman et al., 2004), despite hemispherically symmetric postulated forcings (solar irradiance, volcanism, greenhouse gases). This is particularly apparent in Baffin Island where Little Ice Age snowline depression over the north central plateau exceeded 500 m. A plausible explanation for the larger amplitude of cooling is a strong sea ice feedback. As relatively warm, saline Atlantic water in Baffin Bay was replaced by cold, low-salinity Arctic Ocean surface water (e.g. Dyke et al., 1996), it became more susceptible to sea ice formation. The modern distribution of winter sea ice across Baffin Bay is strongly asymmetric, with no sea ice well north along West Greenland, where surface currents are dominated by Atlantic water, whereas winter sea ice extends well into the Labrador Sea on the Canadian side, where Arctic Ocean surface water dominates. The positive feedback on cooling associated with sea ice, enhanced by the steady reduction in summer insolation at high northern latitudes (Berger, 1988) may explain the larger late Holocene temperature depression over Baffin Island than in the Pacific sector.

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